Contrasting soil development on the sedimentary Kootenai Formation and granitic Boulder batholith in southwestern Montana

by Roger Joseph Veseth

A thesis submitted in partial fulfillment of the requirements for the degree of MASTER OF SCIENCE in SOILS
Montana State University

Abstract:
Characteristics of soils on the Kootenai Formation and Boulder batholith were compared to demonstrate relationships between soils and geologic units in Montana. The Early Cretaceous Kootenai Formation includes clayey shale, fine-grained sandstone, limestone and a basal conglomeratic sandstone. The batholith is composed of a uniform coarse-grained granitic rock, 25 pedons on each parent material were selected by point intercept transects for description, sampling and laboratory analysis.

Thirty of 44 soil properties contrasted sharply between the two parent materials. Boulder batholith soils have sandy loam or loamy sand textures; 40 percent higher frequency of a C horizon in the upper 100 cm; mean pH 0.6 unit higher; larger structure size and weaker grade; and 2-22 percent coarse fragment content. Kootenai Formation soils are more variable, having 7 textural classes ranging from sandy loam to clay and a clay content range of 3 to 74 percent. They have 20 cm greater mean solum thickness; more variable consistence; redder color hues; and a 0-65 percent coarse fragment content. There are 6 family textural classes identified in Kootenai Formation soils compared to 2 on the Boulder batholith.

Less than 10 samples (generally < 5) are required to estimate mean pH; coarse fragment content; sand fraction contents; and sand, silt and clay contents in all horizons of Boulder batholith soils within + 5 units (0.5 for pH) about the mean. Thickness of the B2t and C horizons and depth to the C are the most variable properties, requiring up to 50 samples. In Kootenai Formation soils, about 25 percent of these horizon properties require 10-35 samples and 25 percent require > 30 samples. Coarse fragment content, depth and thickness of the B2t and C horizons, and fine sand, total sand and clay contents of the C horizon require 34-72 samples; clay content being the most variable.

From laboratory analyses of five soils on each parent material, Kootenai Formation soils have 1/3 larger available water holding capacities in the B2t and C horizons; higher extractable Ca, CEC and percent BS; and more variable clay mineralogy ranging from dominantly montmorillonite to kaolinite, illite and interstratified 2:1 clays. Boulder batholith soils are dominated by illite clay with < 25 percent montmorillonite.

This documentation of soil differences related to geologic units will allow more accuracy and efficiency in soil mapping and management .

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by

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A thesis submitted in partial fulfillment of the requirements for the degree

of

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ABSTRACT

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Less than 10 samples (generally < 5) are required to estimate mean pH; coarse fragment content; sand fraction contents; and sand, silt and clay contents in all horizons of Boulder batholith soils within ± 5 units (0.5 for pH) about the mean. Thickness of the B2t and C horizons and depth to the C are the most variable properties, requiring up to 50 samples. In Kootenai Formation soils, about 25 percent of these horizon properties require 10-35 samples and 25 percent require > 30 samples. Coarse fragment content, depth and thickness of the B2t and C horizons, and fine sand, total sand and clay contents of the C horizon require 34-72 samples; clay content being the most variable.

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Chapter 1

INTRODUCTION

The importance of geologic parent material in soil development has long been recognized, being included by Jenny (1941) as one of the five independent soil forming factors. However, lack of knowledge of geologic weathering products and soil-geology relationships in Montana has limited the potential value of this factor in predicting soil properties. The influence of geologic parent material on soil physical and chemical properties is most pronounced in cool, dry regions and in the early stages of soil development (Birkland, 1974; Stauffer, 1935). Montana's young, often glaciated landscapes and relatively cool, dry climate should provide for a strong expression of geologic parent material properties in soils.

The writer did not find any studies that statistically compare the properties in soils developed on contrasting geologic parent materials. This study documents two contrasting examples of geologic parent material influences on soil development in Montana. It statistically compares the predictability and range in variation of selected properties in soils derived from the two parent materials. Efforts are made to explain the occurrence and variablilty of soil properties in relation to properties of the parent materials.

The granitic Boulder batholith was chosen as one of the parent materials because its relatively uniform composition should be reflected in uniformity of some soil properties. This study also provides baseline information for the recently initiated Butte-Whitehall soil survey which includes portions of the batholith exposure. The data are applicable to other exposures of geologic materials of similar origin and composition in the State.

The Early Cretaceous Kootenai Formation, in contrast, contains a variety of lithologies including clayey shales, sandstones, limestones and conglomeratic sandstone, and should contribute to greater soil variability. Although the Formation contains a variety of rock types, it remains fairly consistent in most exposures in the State. Soils on the Kootenai Formation elsewhere in Montana should have properties similar to those of this report.

This study will hopefully stimulate further interest in soil-geology relationships for their potential value as predictors of soil properties.

Chapter 2

LITERATURE REVIEW

Description of the Geologic Parent Materials

The Boulder batholith is an intrusive igneous body formed by the intrusion of molten magma into older geologic materials and subsequent slow cooling beneath the surface. It covers over 2850 km^2 of western Montana, extending 113 km along the Continental Divide between Butte and Helena (Knopf, 1957). Emplacement of the magma occurred during late Cretaceous time, roughly 78-69 million years ago (Knopf, 1957; Robinson et al., 1960; Tilling et al., 1968). Subsequent erosion, mainly during the Eocene Epoch of the Tertiary Period, removed up to 1.6 km of the overlying rock, exposing the igneous rock beneath (Perry, 1962). Figure 1 shows the exposures of intrusive rocks in Montana.

A review of the geologic studies in the area show that the most extensive rock type in the Boulder batholith is quartz monzonite with subordinate granodiorite (Becraft et al., 1963; Klepper et al., 1957; Knopf, 1957; Roberts and Gude, 1953; Robertson et al., 1960; Ruppel, 1963; Smedes, 1966). Rock mineral composition is typically plagioclase and potassium feldspar, with about 20 percent quartz and 15 percent biotite mica and hornblende. The most common grain size is very coarse sand (Wentworth scale), typically 1-3 mm (Becraft et al., 1963; Roberts and Gude, 1953; Ruppel et al., 1963).

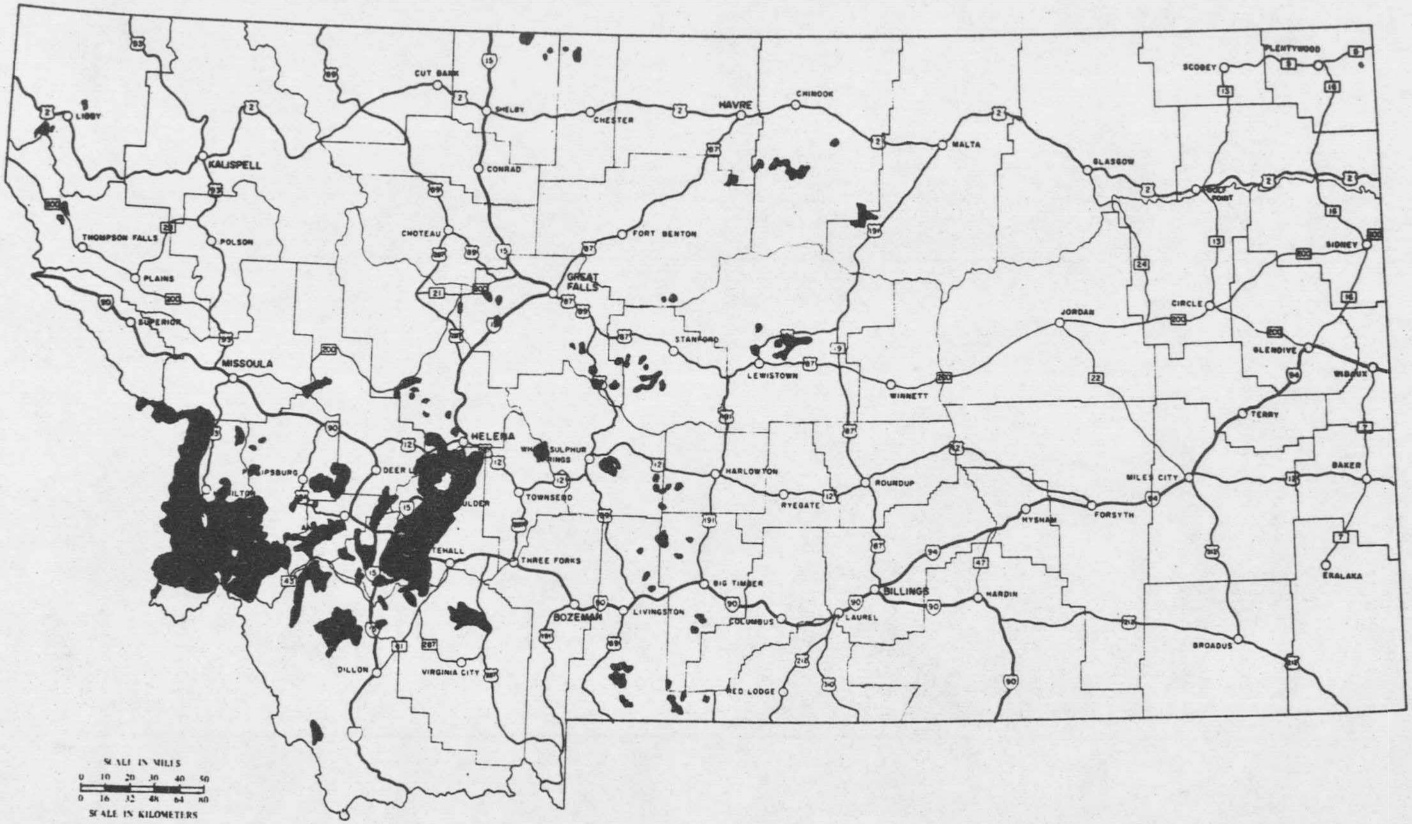

Figure 1. Exposures of coarse- to medium-grained intrusive igneous rocks in Montana (Veseth and Montagne, 1980).

Near the margins of the batholith, the rock composition tends to change somewhat. It becomes darker with an increasing content of iron, calcium and magnesium due to the assimulation of these elements from the "country rock" the igneous rock was intruded into (Perry, 1962; Sahinen, 1950; Tansley et al., 1933). The quartz content decreases and rocks become more basic, grading into granodiotite, diorite and other rock types.

Weathering and erosion along joint planes leave cores of sub-rounded boulders called tor piles, which project above the general land surface (Becraft et al., 1963; Sahinen, 1950). On a larger scale, the batholith landscapes are usually gently rounded, grass and forest covered mountains and foothills. A typical landscape is illustrated in the schematic block diagram in Figure 2.

The rocks have been variably affected by hydrothermal alteration resulting in a variable resistance to disaggregation and erosion (Becraft et al., 1963; Ruppel, 1963). Potassium feldspar and mica are often partially altered to clays by the hydrothermal activity and natural weathering processes (Becraft et al., 1963; Clayton, 1974). The batholith rock commonly disaggregates to a coarse, loose material called grus from mechanical weathering (freeze-thaw, wetting-drying, etc.) and expansion due to clay formation (Arnold et al., 1975; Becraft et al., 1963; Clayton, 1974). On the Idaho batholith to the west, illite, kaolinite and halloysite clay materials predominate in

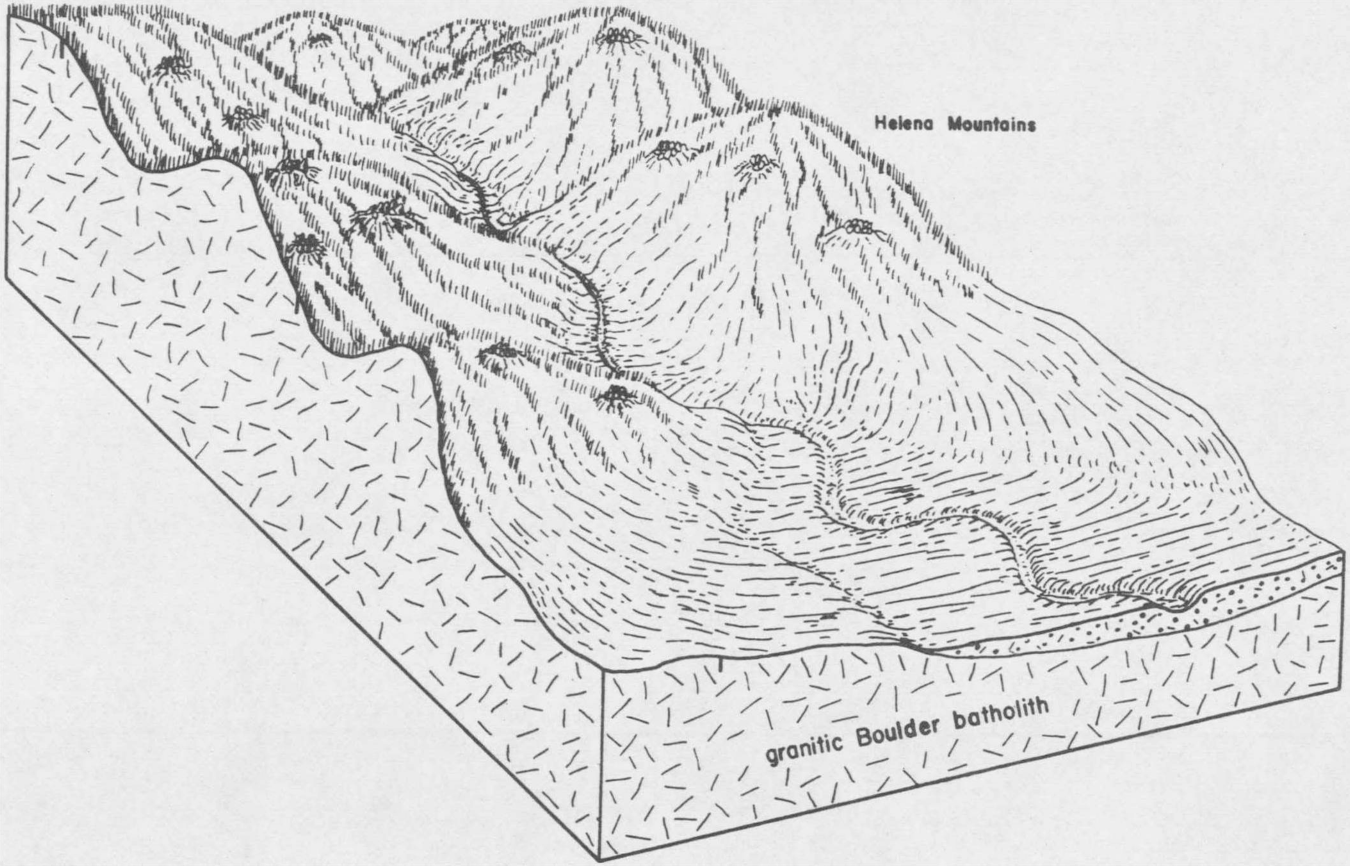

Figure 2. Schematic portrayal of Boulder batholith landscapes (adapted from Veseth and Montagne, 1980).

the soils, with some montmorillonite found at dryer, lower elevations and in poorly drained areas (Clayton, 1974).

The Early Cretaceous Kootenai Formation is commonly exposed in small outcrops in mountain and foothill areas of Montana where mountain building deformation has taken place. Figure 3 shows the exposure of the Kootenai Formation and other soft shale and hard sandstone formations of Pennsylvanian through Early Cretaceous age in Montana. The Kootenai Formation is found in most of the map delineations. It was deposited on the Jurassic Morrison Formation or older rocks and is overlain by Cretaceous marine shales (Imlay, 1952).

A composite description of the nonmarine floodplain, alluvial fan and lacustrine derived rocks of the Kootenai Formation in Montana is summarized by Veseth and Montagne (1980) from numerous geologic studies across the State. This description includes an upper unit of soft varicolored claystone or mudstone, fine-grained sandstone, freshwater limestone and a basal conglomeratic sandstone. The Formation ranges from 100 to 180 m in thickness in most Montana exposures. Of these lithologies, the basal conglomeratic sandstone is the most persistent with the others varying somewhat in proportions from one area to another. Limestones are generally absent in southern Montana east of Bozeman (Foose et al., 1958; Knappen et al., 1931; Moberly, 1960; Richards, 1957; Suttner, 1969). Exposures in the Great Falls-Lewistown area (Ballard, 1966; Fox and Groff, 1966; Harris,

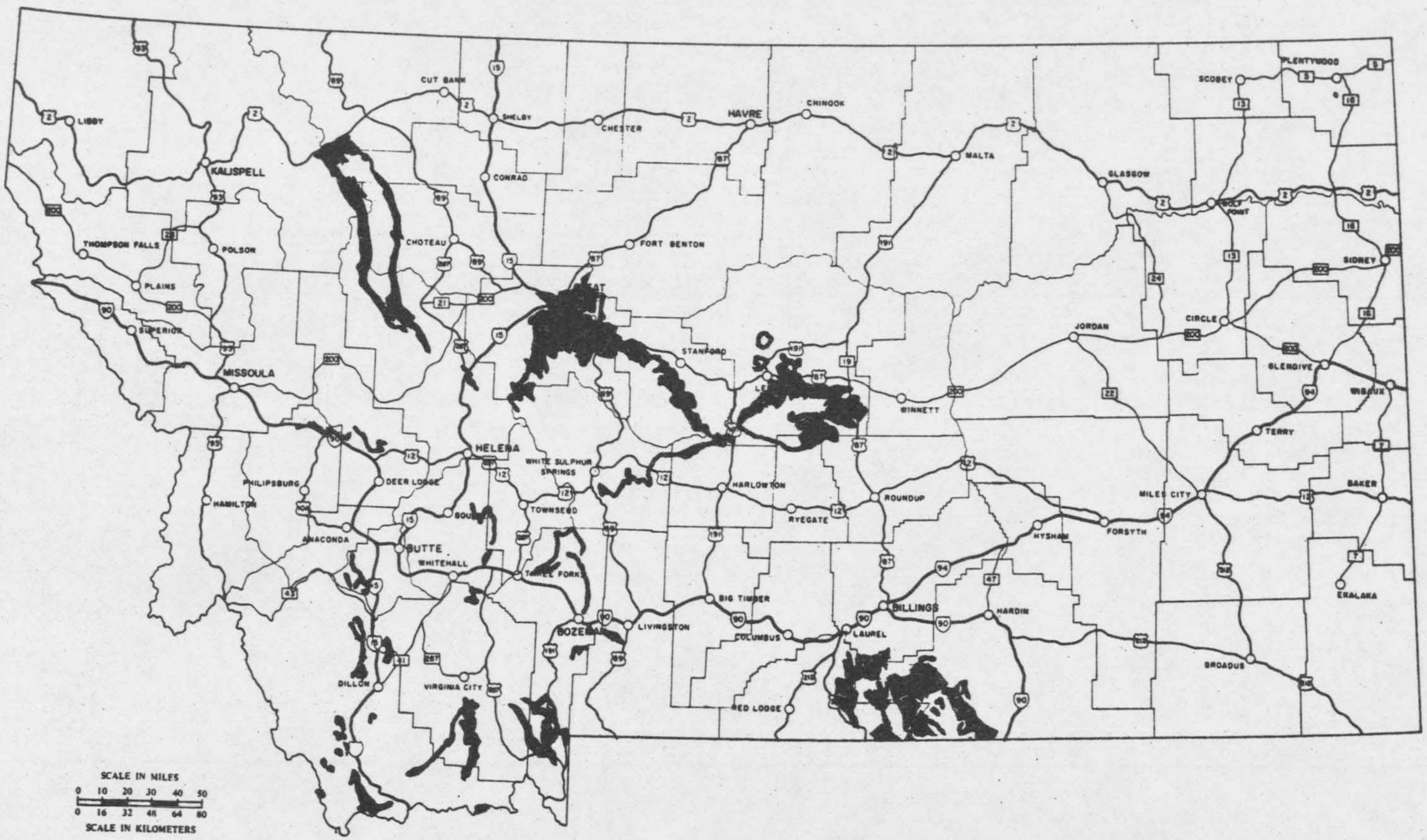

Figure 3. Exposures of soft red-varicolored shales and hard sandstones in Montana (Veseth and Montagne, 1980).

1966; Vine, 1956) and the northwestern exposures (Mudge, 1972) are predominantly clayey shale and sandstone, with only a few thin limestones. In southwestern and western Montana, however, a fossiliferous "gastropod limestone" generally 15 m or less in thickness is commonly found in the upper section of the Kootenai Formation, overlying the middle shale-sandstone unit and basal conglomeratic sandstone (Hadley, 1960; Klepper et al., 1957; Robertson, 1963; Suttner, 1969; Wanek and Barclay, 1966; Witkind, 1969). Kootenai Formation exposures also thicken markedly to the west to 210-425 m (Mudge, 1972; Suttner, 1969; Wanek and Barclay, 1966). The schematic diagram in Figure 4 portrays a mountain foothill landscape of the Kootenai Formation in central Montana.

Clay mineralogy of the Kootenai Formation appears to be somewhat variable. South of Billings in the Pryor-Bighorn Mountains, the Kootenai (Cloverly) Formation seems to be dominantly montmorillonite (Moberly, 1960). In central Montana, Harris (1966) found predominantly mixed-layer illite and montmorillonite clays with some kaolinite, whereas Berg et al. (1968) show kaolinite and illite to be dominant, with some montmorillonite.

A geologic study of the Kootenai Formation area sampled in this thesis was made by Hall (1961). The Formation in this area was divided into three members. The lower member, 24-46 m thick, consists of coarse, massive, salt-and-pepper sandstone with a basal chert-

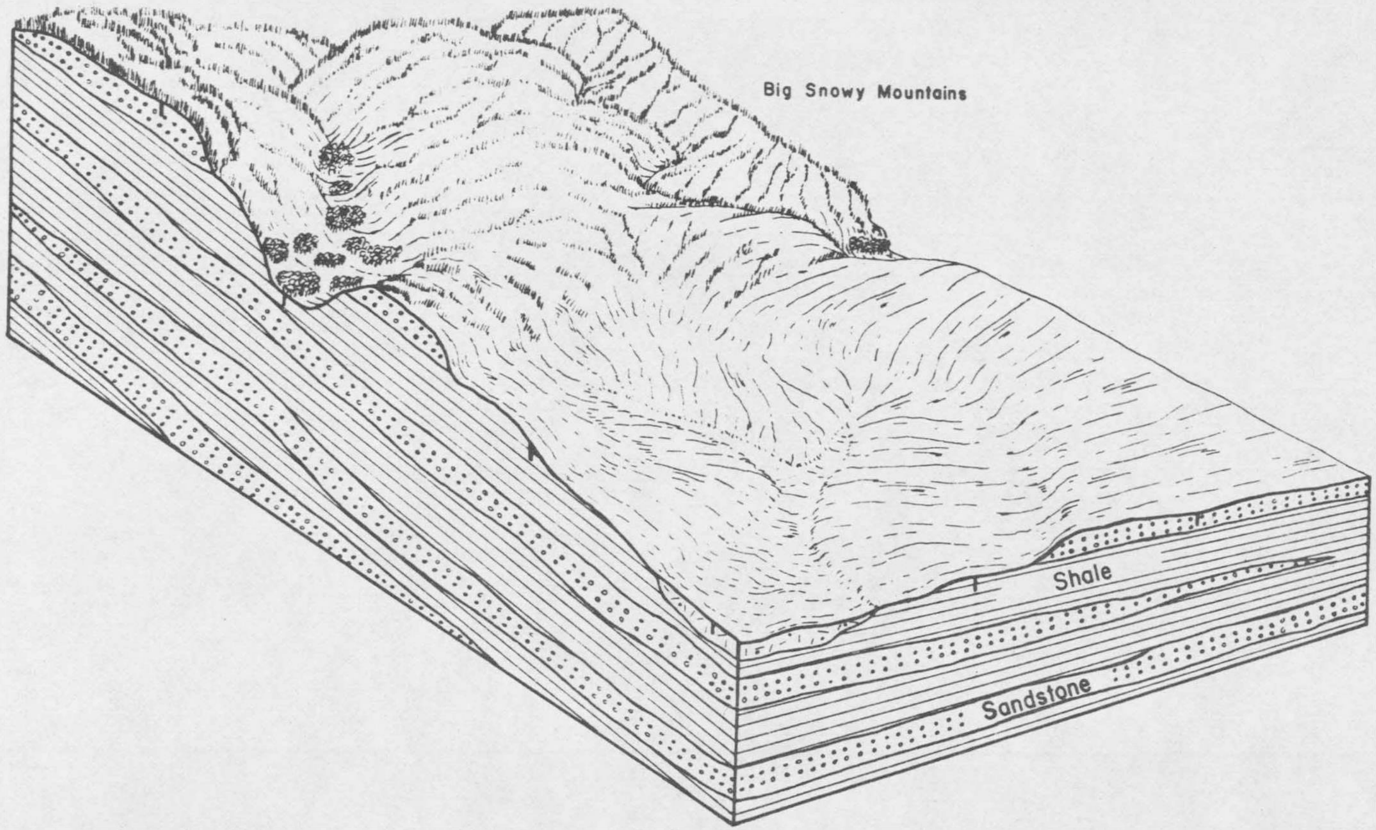

Figure 4. Schematic portrayal of Kootenai Formation landscapes in the Big Snowy Mountain foothills in Central Montana (adapted from Veseth and Montagne, 1980).

pebble conglomerate. It is quite resistant, forming prominent cliffs above the underlying soft Morrison Formation shale. A middle member about 76 m thick consists of red-maroon to yellowish tan claystone, limestones and limy siltstone at the base; 6-18 m of freshwater limestone in the middle; and more red clayey shale, limestones and limy siltstones above. Shales in this member are generally noncalcareous. The limestone in the center section of this member is correlated with the "gastropod limestone" of many other exposures in western Montana. The upper member, 15 m thick, is composed of well sorted, fine-grained, clean quartz arenite that could easily be confused with the Quadrant Sandstone of Pennsylvanian age. It weathers to a distinctive dark reddish brown color due to concentrations of hematite cement on the exposed surfaces. It is a resistant ridge forming sandstone.

The Kootenai Formation area sampled in this study is a northeasterly dipping flank of the northwest trending Buck Creek anticline. The axis area of the anticline is largely removed by erosion. The strata are gently dipping, generally at about 6-8 degrees. The Kootenai Formation extends to the highest elevation of the sampling area, so the soils should not be contaminated by material from other geologic formations.

Studies of Geology-Soil Relationships

That the properties of most soils are influenced, to some extent at least, by the nature of their parent materials has been recognized since systematic study of soils began (Stauffer, 1935). Parent material was included by Jenny (1941) as one of the five soil forming factors along with climate, organisms, topography and time, all of which are theoretically independent. The influence of parent material is particularly pronounced in more arid, cooler regions and in the initial stages of soil development (Birkland, 1974; Stauffer, 1935). However, even on the North Carolina Piedmont, one of the "oldest landscapes in the country", rock differences remain as the probable first cause of soil variability (Cady, 1950).

A review of many of the soil-geology studies reveals that soil texture and clay mineralogy are most commonly associated with parent material influences. In North Carolina, Cady (1950) found differences between soils formed on meta-gabbro and diorite to be mainly in texture and clay mineralogy. Variations in depth of weathering and certain aspects of soil structure were also found. In the Tuscahoma Sand Formation in Alabama, soils derived from a clayey geologic section had different textural and mineralogical classes at the family level of soil taxonomy than those of sandy sections (Hoyum and Hajek, 1979). Soils from the clayey strata also exhibited a higher cation exchange capacity. Gibbs and Perkins (1966) also documented the importance of bedrock parent material to soil genesis. They noted that lithologic variations in acid crystalline rocks in Georgia resulted in different fine sand contents and clay mineralogy between the Hayesville and Cecil series.

A soil-lithosequence study on granite, pyroclastics and schist was made by Parsons and Herriman (1975) in the mountains of southwestern Oregon. Soils developed from granite have coarse textures low cation exchange capacities and base saturations, and no argillic horizons. Soils from pyroclastic rocks are fine-textured, have the highest cation exchange capacities and base saturations, and generally have argillic horizons. The soils from schist may or may not have argillic horizons and have chemical and physical properties intermediate to those from granite and pyroclastic rocks. No clay mineralogy investigations were undertaken. A higher woodland productivity and fewer watershed management and construction problems were also associated with soils developed from pyroclastic rocks compared to soils developed from schist or granite. They noted that soil landscapes underlain by schist are especially susceptible to mass movement.

In California, Harradine and Jenny (1958) found that granodiorite weathers to a sandy, relatively light-colored mantle in contrast to basalt, which weathers to a dark-colored fine-textured soil. Due to the differences in weathering rates, they pointed out that one usually finds a deep soil mantle overlying basalt and a relatively shallow soil profile developed from granodiorite. The granodiorite, richer in quartz and potassium feldspar, decomposes more slowly the the basic igneous basalt, which contains calcium plagioclase and a higher percentage of less stable mafic minerals. The granodiorite derived soils have a higher texture variability than basalt soils and, in addition, a lower available soil moisture holding capacity and nitrogen content.

McCaleb (1959) compared soils formed from acid crystalline rocks and sandstones in North Carolina. Soils from the crystalline rocks had a higher clay content, lower cation exchange capacity and lower free iron oxide content than soils from sandstone. Clay mineralogy was dominantly kaolinite with minor vermiculite in soils of crystalline rocks, whereas soils from sandstone contained nearly equal amounts of kaolinite and vermiculite, with some illite. The higher vermiculite content in sandstone derived soils was suggested as a reason for the higher exchange capacity of these soils.

Ciolkosz et al.(1979) compared soils developed from colluvium of several geologic lithologies in the Ridge and Valley area of Pennsylvania. Contrasting soil properties include texture, color, base saturation, clay mineralogy and coarse fragment content.

In Manitoba, Canada, Ehrlich et al.(1955) found that the carbonate content of parent materials of glacial till soils has a profound effect on the type of soil profile formed. They concluded that increasing amounts of inorganic carbonate restricted profile development and inhibited the decomposition of noncalcareous rock fragments. Also, the carbonate content in the parent material was a major factor in determining and differentiating certain great soil groups. Thickness of the solum typically decreased with increasing carbonate content. They emphasized that although pH or soil reaction is usually considered to be a reflection of the degree of base saturation of the soil absorption complex, it also may be a reflection of the composition of the material from which the soil is derived.

In a lithosequence study in Greece, Yussoglou et al. (1969) conclude that soils with argillic horizons were developed from parent material having higher initial base saturation when compared to soils with cambic horizons. Wells (1959), in New Zealand, concludes that soil developed from greywacke are low in fertility, "podzolized" with residual quartz, and are less productive when compared to soils derived from basalt.

Parent Material--Clay Mineralogy Relationships

Birkland (1974) states that all other factors being equal, the texture of the parent material has a great influence on the course of soil development. He proposes that clay formation would probably proceed more rapidly in the finer-textured material because of the greater surface area available for weathering. Constituents for clay formation are more often removed by leaching in soils of coarsetextured parent materials resulting in lower base status. Because of this, he theorizes that soils of finer-textured material might have a higher ratio of 2:1 layer clays to 1:1 clays than those of more permeable material.

Clay minerals in a soil may originate by means of three different mechanisms: (1) inheritance from parent material, (2) alteration and degradation of primary mineral, and (3) synthesis by precipitation (Grim, 1968). For sedimentary rocks, inheritance from parent material is believed to be the predominant mode of clay origin (Grim, 1968; Van Houten, 1953). In 35 of 43 paired samples of soils and their sedimentary parent bedrock from Wyoming to Pennsylvania, Van Houten (1953) found the predominant clay mineral groups in the soils to be the same as those of their parent rock. Specific examples of soil clay mineralogy relationships to their parent materials include: montmorillonite White Store soil from montmorillonite-rich Triassic shale in North Carolina; kaolinitic Gasport soil from kaolinite-rich Pennsylvania shale in Ohio; and illitic Miami soil from illite-rich Late Wisconsin glacial till, also in Ohio.

Grim (1968) states that kaolinite is frequently the most abundant clay mineral of soils formed in sands and sandstones because they typically have a low base status and high leaching environment that is conducive to kaolinite formation. Also, kaolinite is the dominant clay mineral in ancient lacustrine sediments (such as in sections of the Kootenai Formation of this thesis) deposited in a low pH and active water environment. In alkaline, less active waters, illite and montmorillonite are usually the dominant clay minerals.

Johnson (1970) found kaolinite as the dominant clay mineral in soil parent material from sandstones and metamorphics in Pennsylvania. He theorizes that in the sandstone derived soil, this is probably a manifestation of the inheritance mechanism. Illite was found to be the dominant clay mineral of 95 percent of the shales from Precambrian to Tertiary-Pliocene age and their derived soils.

One predication upon which inheritance of soil clay mineralogy from its parent material is based is the uniformity of the clay minerals with depth in the soil profile. Numerous recorded analyses, such as those in Iowa soils by Peterson (1946), and Russell and Haddock (1941), and those in Minnesota soils by Caldwell and Rost (1942), reveal no appreciable change in colloidal composition or clay mineralogy with depth in the profiles of soils developed on sedimentary rocks. In Manitoba, Canada, Ehrlich et al.(1955) found that montmorillonite, illite and kaloinite content of soils formed on Mankato-age glacial till is relatively uniform throughout the profile. They concluded that this indicates the clay minerals were not formed in situ by recent soil forming processes, but rather were inherited from pre-Mankato soils or geologic strata.

The clays in soils derived from crystalline rocks, both igneous and metamorphic, are not directly inherited from the parent rock but are the products of alteration and degradation of primary minerals and thus more often reflect the environmental constraints and weathering processes of the soil forming period (Van Houten, 1953). In California, Barshad (1966) found illite to be a dominant clay mineral of soils derived from felsic intrusive igneous rock (the Boulder batholith of this study is predominantly felsic igneous rock) and attributed this to the biotite mica content of the rock and the availability of potassium. On the granitic Idaho batholith of Idaho and western Montana, which is mainly of granodiorite composition (Boulder batholith of this study is of quartz monzonite and granodiorite composition) Clayton et al. (1979) concluded that the biotite commonly weathers to degraded mica, then to a smectite-iddingsite product, and eventually to illite. Sericitic weathering products and ultimately, some kaolinization of feldspars are common. They found mixed-layer clays to be notably absent in soils of the Idaho batholith. Grim (1968) points out that felsic igneous rocks containing considerable quantities of potassium as well as magnesium will yield illite and smectite as the alteration products under weathering conditions which permit the potassium and magnesium to remain in the weathering environment. If the content of magnesium is low, illite will be the main product and if potassium is low, smectite will be the main product. Rapid removal of both elements would lead to the formation of kaolinite. Loughon (1969) concluded that, in dryer environments, potassium feldspar in felsic crystalline rocks will generally weather to illite and montmorillonite instead of kaolinite. Birkland (1974) and Loughon (1969) point out that parent material exerts an important control on chemical weathering and the clay mineral formation. Weathering releases constituents essential to the formation of the various clays, and the inherent texture and permeability influences how fast those constituents may be removed by leaching.

In most cases clay mineralogy does vary somewhat with depth in the soil profile. Birkland (1974) suggests two possibilities for this variation. First, that clay mineral assemblages are not in equilibrium with present environmental conditions and are being altered to more stable clay minerals, such as is often the case with clay mineral inherited from the parent material. Second, some clay minerals could have formed in the past under different environmental conditions, and with reaction rates being so slow under surface conditions, these clay minerals are metastable in the present environment. He also attributed some variability to the diverse microenvironmental conditions within the soil. Johnson (1970) suggests that the three mechanisms of clay origin (inheritance, alteration and degradation of primary minerals, and synthesis) operating under different profile weathering conditions together with the process of the translocation of material results in soil clay mineral composition becoming, to some degree, a function of soil depth. He states that weathering is most intense at the soil surface and decreases in intensity with depth. This is the horizon depth function of Jackson et al.(1948) which leads, in many cases, to the development of a profile of weathering in which clay mineral distribution and composition changes somewhat with depth.

Determining Soil Variability

Part of the objective of this study is to compare the predictability and range of variability of properties and features of soils derived from the Kootenai Formation and Boulder batholith. The literature was reviewed for site selection methods, sample size and statistical procedures. However, none of the soil-parent material studies previously cited attempt such statistical comparisons. Typically, only a few "representative" soil sites were compared with those of contrasting substrates. Since these soil-parent material studies had small sample sizes and little, if any, statistical comparisons, studies evaluating soil mapping unit composition and variability were reviewed for statistical approaches and soil property variability.

In a study of variability of soil morphological properties within three mapping units of Ohio, Wilding et al. (1965) employed a grid method for site selection. They chose ten numbers from a random numbers table to designate sites on a numbered grid overlain on an aerial photo. This was repeated for 24 mapping unit delineations, eight delineations for each of the three mapping units, for a total of 240 observations. The density of observations per unit area was found to be relatively constant for all mapping unit delineations, thus allowing for variability comparisons between mapping units. The most variable soil properties were horizon thickness, depth of leaching of carbonates, loess thickness, depth of mottling, pH, and size class of soil structure. Clay content, soil structure grade (strength), and soil color were the least variable properties. The number of samples per delineation required to estimate the mean within 10 percent was 7-9 for all properties except depth of leaching, loess thickness and drainage class, which required, 12, 31, and 14 samples, respectively.

Crosson and Protz (1974) also used a grid system for site selection in a mapping unit composition study in Ontario, Canada. However, instead of random numbers, they sampled every 10 m on a 80 X 190 m grid. The calculated number of samples required to detect significant differences between the means of soil properties of the two mapping units at the .05 significance level was less than 60 for only six properties. These included chroma and hue of soil color, percent organic matter and percent sand, silt and clay in the Ap horizon. The number of samples required for other properties generally ranged from 118 to over 3000 (for pH).

In a statistical summary of physical and chemical properties of the Morley and Blount series in Ohio, Wilding et al. (1964) selected 59 total profile descriptions and lab analyses collected over the previous ten years in conjunction with the Ohio soil survey program. They felt that although the requirements of random sampling were not strictly satisfied, they were approximated because of the large scope of the survey program, the wide geographical scattering of sites, long time span during sampling and collection by different soil scientists. They also postulated that most characteristics would approximate a normal distribution and a range similar to those defined in each of the established series descriptions. For both series, the least variable properties were composite thickness, hue and value variables of color, depth of leaching of free carbonates, silt content and total clay content. These properties required less than ten samples to estimate the mean within 10 percent.

McCormack and Wilding (1969) investigated the variation of soil properties within eleven mapping units in Ohio. They randomly selected two delineations of each mapping unit from aerial photos and ten sampling sites in each delineation. The density of sampling sites per unit area was held relatively constant for all delineations, thus permitting reliable estimates of the aerial extent of included soil properties. Standard deviations, coefficients of variation and the number of samples required to estimate the population mean within

$\pm$ 10 percent indicated that thickness of the B1, B2 and IIB horizons, depth to mottling, soil structure grade and chroma are the most variable properties. The least variable soil properties were hue, value, pH of all horizons, texture of the IIB horizon and the size and type classes of soil structure.

Nelson and McCracken (1962) made a statistical summarization of soil properties of the Norfolk and Portsmouth series in North Carolina. Central tendencies, ranges and standard deviations of the soil properties were determined from 15 randomly selected sites of each series. The soil properties were analyzed for potential criteria for soil classification and characterization, and influence on corn yield. They concluded that a sample size of 15 is inadequate for estimating most soil properties. The only factor that showed promise as a differentiating criterion of corn yield potential was the 2:1 to 1:1 layer clay ratio.

In a soil variability study on a red pine plantation in Massachusetts, Mader (1962) found exchangeable Ca and Mg to be the most variable property. Only two samples were required to estimate the mean bulk density within $\pm$ 10 percent, whereas 20 samples were required for organic matter content and over 100 samples for exchangeable Ca and Mg. He attributed this to the greater natural variability of some soil properties than others.

Steers and Hajek (1979) evaluated the accuracy of randomly selected point intercept transects in determining map unit composition. The point intercept transect method is basically sampling at regular spacings on a straight line transect. The transect was distributed evenly throughout the mapping unit delineations for a uniform expression of variability. They concluded that many mapping units in a survey area can be characterized at an 80 percent confidence level by determining soil occurrence along less than ten transects with a minimum of ten point observations per transect. Points on the transect were located by pacing with compass orientation. The transect method was found to be particularly useful in wooded areas and in complex terrain with limited accessibility where soil boundaries and landmarks are not easily observed.

Ball and Williams (1968) also found the point intercept transect method applicable in evaluating the variability of soil chemical properties of mountain rangeland soils in North Wales. Points were located at 27 m intervals on parallel transects 260 m apart. In Michigan, Amos and Whiteside (1975) used the point intercept transect method for site selection in evaluating soil survey mapping accuracy in an urbanizing area. Randomly selected parallel transects were 152 m apart with points at 76 m intervals. A minimum of 50 samples were required to estimate the property means within ± 10 percent.

In evaluating mapping unit composition and precision in Georgia, Powell and Springer (1965) used the point intercept transect method for site selection. Point observations were determined by pacing and compass orientation. The transects were randomly selected and transect spacing varied with size and complexity of the mapping unit. Transects were about 1.2 km in length. The point intercept transect method was used because soil boundaries were not easily discernable in the wooded and hilly topography, making random site selection infeasible.

Wicherski (1980) analyzed the variability of soils in two forest land type mapping units in Oregon with the point intercept transect method. Eight delineations of each mapping unit were sampled with randomly located transects consisting of 40 point observations each. Variability was quantified in terms of estimates of required sample sizes, ranges and coefficients of variation. He found chemical properties to be more variable than physical or morphological properties, which were about equal. The most variable chemical properties were extractable bases and organic matter content. Coarse fragment content and thickness of the 0 horizon were the most variable physical and morophological properties. Least variable properties in general were pH, CEC, bulk density, texture, and color. A major source of variation was attributed to sudden changes in the character of the parent material and the variety of geomorphic surfaces of varying age and stability.

Chapter 3

MATERIALS AND METHODS

Study Area Descriptions

The Kootenai Formation study area is located in the Madison Range south of Big Sky in southwestern Montana (see Figure 5). This area has a mean annual precipitation of 46-66 cm (Ross and Hunter, 1976) and mean January, July and annual temperatures of -16, 25 and 5^{o} C, respectively (Cordell, 1960). Elevations of the Kootenai Formation exposure (outlined on the map in Figure 5) ranges from 1830 m in the northeastern portion to about 2745 m in the southwest. The Formation exposure is a smooth continuous dipslope with a northeast dip. This means that the topographic surface roughly parallels the inclined geologic strata which are plunging down to the northeast.

The Kootenai Formation exposure is predominantly forest-covered with numerous scattered meadows comprising about 30 percent of the area. Dominant tree species include Abies lasiocarpa, Pinus contorta, Pinus albicaulis, Pinus flexilis and Pseudotsuga menziesii, with understory species of mainly Vaccinium scoparium, Physocarpus malvaceus, Calamagrostis rubescens, Spiraea betulifolia and Arnica cordifolia. Meadow vegetation includes Agropyron smithii, Festuca idahoensis, Artemisia tridentata, Balsamorhiza sagittata, Lupinus spp. and a wide variety of other forb species.

The exposure of much of the Boulder batholith is outlined on the

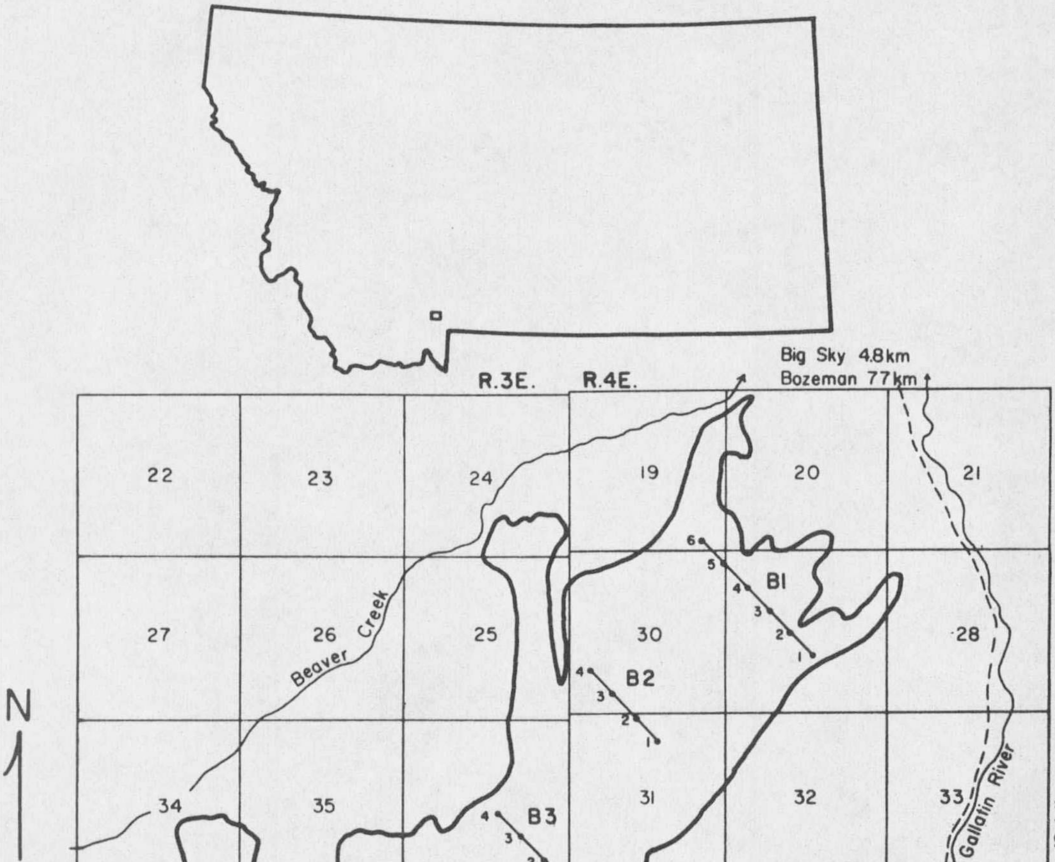

Figure 5. Kootenai Formation exposure with transect and point observation locations.

map in Figure 6. Smooth rounded mountain slopes with occasional to numerous rock outcrops called tor piles typify much of the batholith topography. The study areas on the batholith are located in portions of three mountain ranges: the Elkhorn Mountains in the northeast; Helena Mountains in the central and northwestern portions; and the Main Range in the south. Elevations of the study areas generally range between 1525 and 1920 m. Estimated mean annual precipitation is in the 36-56 cm range (Ross and Hunter, 1976). Mean January, July and annual temperatures in the study area are -14, 28 and 7^{o} C, respectively (Cordell, 1960).

The batholith study areas are largely forested with lower mountain slopes under grass cover. _Pseudotsuga menziesii_ and _Pinus ponderosa_ are the dominant tree species with understory species including _Agropyron spicatum_, _Festuca idahoensis_, _Berberis repens_ and _Arctostaphylos uvi-ursi_, though most forested sites do not support understory vegetation. Grassland and meadow species are mainly _Agropyron spicatum_, _Festuca idahoensis_ and _Artemisia tridentata_.

Site Selection

The soil site selection methods used on both parent materials reflect an attempt to encompass the widest range in soil variability within the constraints of limited access, time and manpower. Due to the difficulty of locating randomly selected or regular grid-selected

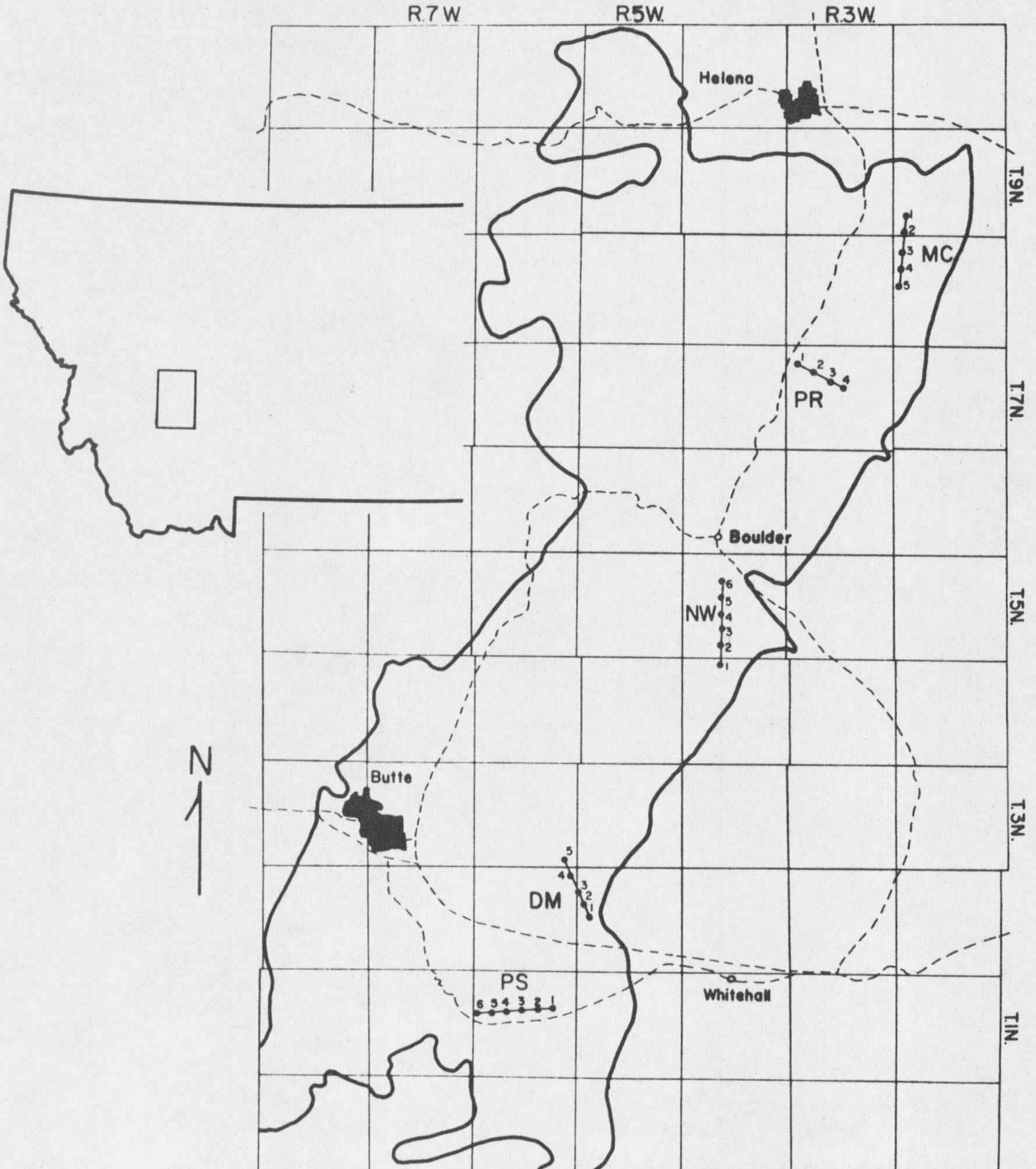

Figure 6. Boulder batholith exposure with transect and point observation locations.

points in forested terrain, a point intercept transect method was chosen. On the Kootenai Formation, five parallel transects were spaced 1.6 km apart with point observation every 322 m (5/mi). A total of 25 observations were described and sampled with 4-6 per transect. Transect lengths varied slightly with the width of the formation exposure. The transects were drawn on a topographic map and located in the field with an aerial photo. The first site on each transect was randomly selected and each successive site along the transect was located by pacing and compass orientation. Transects and point observation locations on the Kootenai Formation are shown in Figure 5.

The point intercept transect method of site selection was also used on the Boulder batholith; however, a different spacing was required. Due to the limited access on the batholith, transects were oriented along roads on the eastern half of the batholith. Transect spacing is 10-19 km. The transects were drawn on topographic maps and point observations were located every 1.6 km along the transects with spacing determined by car odometer and map reference points. Sample sites are at least 50 m from the roads. A total of 25 sites were described and sampled with 4-6 per transect. Transect and point observation locations on the Boulder batholith are displayed in Figure 6. The letter symbols near the transects indicate general points of reference; for example, PS refers to Pipestone Pass.

Soil Site Description and Sampling

All soil sites were excavated by hand to a depth of 100 cm or to bedrock. Approximately 2 kg samples were collected from each horizon for lab analyses. For each horizon, the following properties were measured by procedures of the Soil Survey Staff (1951): thickness; moist color (Munsell Color Chart); texture; structure type, class and grade; wet and moist consistence; effervescence (0.1 normal HCl); pH (Hellige-Truog field pH kit); and percent (by vol) coarse fragments > 20 mm. Slope and aspect of the sites were determined with a clinometer compass. Dominant vegetation was identified and rockiness and stoniness classes (Soil Survey Staff, 1951) were estimated within a circle with a radius of about 25 m from the point observation.

Physical Laboratory Analyses

Soil samples were air dried and measured for dry color (Munsell Color Chart) and dry consistence (Soil Survey Staff, 1951), and then ground with a flail-type grinder. Samples were hand sieved with 20 mm and 2 mm sieves for coarse fragment separation. The coarse fragment contents were estimated on a volume basis with measurements of dry total sample volumes and water displacement volumes of the coarse fragment fractions.

Soil particle size analysis was determined by the hydrometer method as described by Day (1965). Sand size fractions were determined by decanting the suspended sediment from the hydrometer analysis cylinders, oven drying the remaining sand and passing it through a sieve stack. The sieves were placed on a reciprocating sieve shaker for three minutes at 120 strokes per minute. Sieve sizes were: .0053, .0106, .0250, .0425, .05 and .1 mm. Weights of the sieved fractions were plotted on a semilog graph along with the < 2 mm hydrometer method data for percentage conversions of the sand size fractions. Coarse (very coarse and coarse) medium and fine (fine and very fine) sand size contents were determined (Soil Survey Staff, 1951).

Soil Water Holding Capacity

Five soils on each parent material, one from each transect, were analyzed for water holding capacity. The percent soil water at 1/3 and 15 atmospheres tension was determined gravimetrically with a ceramic plate apparatus (method 30) and pressure membrane apparatus (method 31), respectively, used in Agric. Handbook 60 (Salinity Laboratory Staff, 1954). The percent available water holding capacity was estimated as the percent soil water content between those tensions.

Clay Mineralogy

The same five soil profiles on each parent material were analyzed for clay mineralogy. Samples of the < 2 mm fraction from the A and C (or B2) horizon at 100 cm depth were prepared using the methods described by Thiesen and Harward (1962). X-ray analysis methods, using a

Cu target and Ni filter, and interpretation procedures are from Whittig (1965). The relative abundance of alumino-silicate minerals in the clay fraction was determined by a semiquantitative method comparing the areas under the diffraction peaks (Klages, 1980). In this method, area weighting factors are established for the various clay minerals. The peak areas are multiplied by the weighting factors, and the respective percentages of each clay mineral calculated.

Chemical Laboratory Analysis

Chemical analyses of the five selected soil profiles on each parent material were made by the Soil Testing Laboratory at Montana State University, Bozeman. Organic matter percentages were calculated from colormetric determination (Sims and Haby, 1970). A 2:1 water-soil mixture was used for pH and electrical conductivity measurements. Extractable cations (Bower et al., 1952) and cation exchange capacity from Method 19 of Agric. Handbook 60 (Salinity Laboratory Staff, 1954) were determined by atomic adsorption spectrophotometry

Statistical Comparisons

Statistical comparisons were made of a variety of quantitative soil properties in the A1, A2, B2t and C horizons of all soils sampled on each parent material. These properties include: horizon depth and thickness; pH; coarse fragment content; coarse, medium and fine sand content; and sand, silt and clay content. From Snedecor and Cochran (1973), the statistical comparisons included means, standard deviations, ranges, 90 percent confidence intervals and the number of samples (N) required to estimate the mean within a specific number of units (cm, percentage points or $\log_{10}$ units) about the mean. A .10 significance level was used for both the 90 percent confidence interval and to estimate N. Predictability of qualitative soil properties such as color, structure and consistence were estimated by frequency distributions.

The significance of differences in soil property means between parent materials was determined with the t statistic (Snedecor and Cochran, 1973) at the .10, .05 and .01 levels. Pooled variances were used since unequal sample sizes resulted with property analysis by horizon. A normal distribution was assumed. The assumption of equal variances was not completely met for all soil properties introducing possible error in significance of differences in some cases.

Chapter 4

RESULTS AND DISCUSSION

The contrasting lithologies of the Kootenai Formation and Boulder batholith, previously described, should be reflected in a variety of contrasting soil properties. Differences emphasized in this study are rockiness and stoniness; depth to bedrock; solum thickness; coarse fragment size and content; sand, silt and clay content; coarse, medium, and fine sand content; structure type, size and grade; inherited color; consistence; thickness of the B2t horizon; water holding capacity; clay mineralogy; cation exchange capacity; extractable Ca concentration; and pH as well as other soil properties. Greater soil variability in Kootenai Formation soils due to the variability of lithologies is also an important difference.

The following soil characterizations and comparison of selected soil properties indicate that there are some important differences between soils of the two parent materials. An awareness of these property and variability differences should help to improve the accuracy of soil inventories and management on the two parent material exposures.

Rockiness and Stoniness

Soil rockiness refers to the percent bedrock exposure, either as rock outcrops or soil areas too shallow over bedrock for agricultural use. Soil stoniness indicates the percent stones over 25 cm in diameter near or on the soil surface. A comparison of rockiness and stoniness classes of soil on the Kootenai Formation and Boulder batholith is shown in Table 1.

Only 16 percent of the Kootenai sites contained rock exposures compared to 78 percent on the batholith, with a majority of the batholith sites in rockiness classes 2 and 3. The difference in degree of stoniness, in contrast, is small, with sites on both parent materials having the same frequency of class 0 stoniness. High stoniness classes 4 and 5 include a larger percentage of Kootenai sites than those on the batholith, 20 percent as compared to 4 percent. This reflects the greater weathering resistance of some Kootenai Formation sandstones and limestones in contrast to the uniformily weathering batholith rocks.

All Kootenai sites are located on the dip slope exposures. On scarp slope exposures, where the topographic surface cuts across the bedding of the geologic strata, Kootenai soils would probably tend to have higher rockiness and stoniness classes.

Depth to Bedrock

Depth to bedrock (Table 2) in the > 100 cm class tends to be slightly greater in Kootenai soils (12 percent higher frequency). Batholith soils have an 8 percent higher frequency of being < 50 cm deep to bedrock. Frequencies of medium depth ranges are nearly the

Table 1. Frequency distribution of rockiness and stoniness classes[†] of soils on the Kootenai Formation and Boulder batholith.

Parent Material	Rockiness Classes[§]						Stoniness Classes[‡]					
	0	1	2	3	4	5	0	1	2	3	4	5
	%											
Kootenai Formation	84	8	8	—	—	—	36	20	24	—	12	8
Boulder batholith	12	16	32	36	4	—	36	12	28	20	4	—

† Soil Survey Staff (1951)

§ 0 = < 2%, 1 = 2-10%, 2 = 10-25%, 3 = 25-50%, 4 = 50-90%, 5 = > 90%.

‡ 0 = < 0.01%, 1 = 0.01-0.1%, 2 = 0.1-3.0%, 3 = 3-15%, 4 = 15-90%, 5 = > 90%.

Table 2. Frequency distribution of the depth to bedrock in soils on the Kootenai Formation and the Boulder batholith

	Depth Intervals†		
	---cm---		
Parent Material	< 50	50-100	> 100
	---%---		
Kootenai Formation	4	24	72
Boulder batholith	12	28	60

† Intervals for shallow, moderating deep and deep soils (Soil Survey Staff, 1951).

same in soils of both parent materials. The greater frequency of deeper soils on the Kootenai Formation may reflect a predominance of soils derived from soft shale and an overall lower weathering resistance of Kootenai Formation strata in comparison to the crystalline batholith rocks. Some Kootenai Formation sandstones and limestones, however, are also quite resistant to weathering, resulting in soils with higher stoniness classes (previous section) and higher coarse fragment contents, which are discussed later.

Solum Thickness

Solum thickness (Table 3) is greater in Kootenai soils ($\overline{X}$ = 70.8 cm) than batholith soils ($\overline{X}$ = 50.6 cm) with the difference between means significant at the .05 level. Variability of solum thickness, however, is very similar between soils of the two parent materials with similar standard deviations and ranges. The 90 percent confidence intervals are nearly the same width, but they do not overlap.

Table 4 shows that a similar number of samples are required to estimate the mean solum thickness for soils of both parent materials. A confidence interval width of $\pm$ 5 cm about the mean appears to be the most feasible.

Horizon Frequency

A frequency distribution of the presence of selected soil horizons in the upper 100 cm of the profiles is shown in Table 5. The horizon

Table 3. Statistical data on solum thickness in the upper 100 cm of soils on the Kootenai Formation and Boulder batholith.

Parent Material	$\bar{X}^{*}$	SD	Range	90% CI†
	cm			
Kootenai Formation	70.8	28.8	19-100+	61.0-80.1
Boulder batholith	50.6	27.2	11-100+	41.0-59.9

† Confidence interval at the .10 significance level.
* Difference between means significant at the .05 level.

Table 4. Number of samples (N)# required to estimate the mean solum thickness of soils on the Kootenai Formation and Boulder batholith.

	Confidence Interval Width		
	cm		
Parent Material	±2	±5	±10
Kootenai Formation	24	4	1
Boulder batholith	22	4	1

Estimated at the .10 significance level.

Table 5. Frequency distribution of the presence of selected horizons in the upper 100 cm of soils on the Kootenai Formation and Boulder batholith.

Parent Material	Horizons†						
	0	A1	A2	B2t	C	Cr	R
	- - - - - - - - - - -			%	- - - - - - - - - - -		
Kootenai Formation	40	84	36	84	48	16	28
Boulder batholith	52	64	52	76	88	28	40

† Soil Survey Staff (1951).

with the lowest frequency in soils of both parent materials is the paralithic Cr horizon, which has a 12 percent higher frequency in batholith soils. The frequency of an R horizon or lithic contact is also fairly low (28 percent) in Kootenai soils, but includes 40 percent of the batholith soils. Almost twice the frequency (88 percent as compared to 48 percent) of a C horizon in the upper 100 cm is found in batholith soils as in Kootenai soils. This difference may reflect, in part, the lower availability of clay for development of thick B2t horizons in batholith soils. There is a high frequency of B2t horizons in soils of both parent materials, with the Kootenai soils having a slightly greater frequency, likely again because of the higher clay content of the parent material. Although there is a similar frequency of B2t horizons, large differences in clay content and thickness exist between soils of the two parent materials, as will be discussed later. Kootenai soils have a higher frequency of A1 horizons and slightly lower frequency of O and A2 horizons than batholith soils. This is likely a reflection more of the vegetative cover than parent material influences. There are 14 meadow-grassland sites on the Kootenai Formation and 10 on the Boulder batholith, which may explain, in part, the higher A1 horizon frequency and lower A2 and O horizon frequencies in Kootenai soils.

Horizon Thickness and Depth

The only significant difference (.05 level) between mean horizon thicknesses in the upper 100 cm of Kootenai and batholith soils (Table 6) is in the A1 and B2t horizons. Kootenai soils averaged 3.5 cm thicker A1 horizons and 15.2 cm thicker B2t horizons. The 90 percent confidence intervals for the mean thicknesses of these two horizons do not overlap, whereas all other horizon confidence intervals have large overlaps. Mean depths to the top of the B2t and C horizons are larger for Kootenai soils though the difference is not significant at the .10 level.

In Table 7, the O and A1 horizon thicknesses are the only ones that require a low number of samples to estimate the means within ± 2 cm about the mean in both Kootenai and batholith soils. The number of samples required to estimate the mean thickness and depth to the top of the other horizons are very similar for soils of both parent materials and they are generally only practical at the confidence interval of ± 10 cm about the mean. The variability in horizon thickness, particularly in surface horizons, may be more a result of changes in other soil forming factors such as vegetative cover, slope, aspect, etc. rather than parent material influences.

Soil Color

In the frequency distribution of dry soil color (Table 8), the

Table 6. Statistical data on thickness and depth to the top of selected horizons in the upper 100 cm of soils on the Kootenai Formation (KF) and the Boulder batholith (BB).

		$\overline{X}$		SD		Range		90% CI†	
Soil Property	Horizon	KF	BB	KF	BB	KF	BB	KF	BB
		cm							
Thickness	O	4.8	3.8	1.0	1.3	3-7	2-7	4.2-5.3	2.4-5.2
	A1*	10.3	6.8	3.4	1.8	7-17	4-10	9.0-11.6	6.0-7.6
	A2	23.0	15.2	13.5	15.2	10-45	4-40	14.8-31.2	8.9-21.5
	B2t*	49.2	34.0	22.0	20.6	9-94	8-86	40.9-57.5	25.8-42.2
	C	37.0	43.0	22.3	19.7	8-60	15-76	25.4-48.6	35.8-50.2
Depth to the top	B2t	20.8	14.5	13.8	11.9	6-53	4-38	15.6-26.0	9.8-19.2
	C	47.8	38.8	16.6	17.5	19.68	11-72	39.2-56.4	32.4-45.2

† Confidence interval at the .10 significance level.

* Difference between means significant at the .05 level.

Table 7. Number of samples (N)# required to estimate the mean thickness and depth to the top of selected horizons in the upper 100 cm of soils on the Kootenai Formation (KF) and the Boulder batholith (BB).

		Confidence Interval Width					
		cm					
		±2		±5		±10	
Property	Horizon	KF	BB	KF	BB	KF	BB
Thickness	0	1	1	1	1	1	1
	A1	9	3	1	1	1	1
	A2	153	130	25	21	6	5
	B2t	360	318	50	50	14	13
	C	401	287	64	46	16	12
Depth to the top	B2t	142	106	22	17	6	4
	C	222	227	36	36	9	9

\# Estimated at the .10 significance level.

Table 8. Frequency distribution of dry soil color† by horizon in soils on the Kootenai Formation (KF) and Boulder batholith (BB).

Horizon	Parent Material	Hue			Values								Chroma							
		5 YR	7.5 YR	10 YR	1	2	3	4	5	6	7	8	1	2	3	4	5	6	7	8
		%																		
A1	KF	24	33	43	—	—	—	19	19	29	29	5	—	38	14	43	—	5	—	—
	BB			100	—	—	25	38	19	17	—	—	6	62	25	6	—	—	—	—
A2	KF		11	89	—	—	—	—	—	—	67	33	11	44	11	22	—	11	—	—
	BB			100	—	—	—	15	54	31	—	—	—	31	31	38	—	—	—	—
B2t	KF	23	29	48	—	—	—	4	29	29	29	9	9	—	19	29	—	43	—	—
	BB			100	—	—	—	11	32	47	—	—	—	5	26	37	—	32	—	—
C	KF	50	17	33	—	—	—	8	17	8	33	33	—	42	—	25	—	33	—	—
	BB			100	—	—	—		18	59	14	9	—	14	23	32	4	18	—	9

† Munsell Color Chart

only predictable difference is in the hue variable of color. All of the batholith dry soil colors are of hue 10 YR. Kootenai dry soil colors range from 5 YR and 7.5 YR to 10 YR in all horizons except the A2, where no hue of 5 YR is found and 89 percent of the hues are 10 YR. The red color predominant in the Kootenai Formation strata is most pronounced in the C horizon with a 50 percent frequency of the reddish 5 YR hue. No striking different or consistent pattern can be found in chroma.

Kootenai soils have higher frequencies of light color values of 7 and 8 than batholith soils. One might expect to find lighter values in forest soils that are subjected to higher leaching and lower pH environments in contrast to meadow and grassland soils. However, it does not explain the lighter values in this case, since there are four more meadow sites on the Kootenai Formation than on the Boulder batholith. The higher precipitation range and associated leaching potential of the Kootenai study area may be a factor. Since the C horizon of Kootenai soils also has a higher frequency of light values (7 and 8) than batholith soils, it appears to be an inherited color difference, possibly accentuated by climatic factors in the upper horizons.

Like dry soil color, the main difference in moist color (Table 9) between soils on the two parent materials is the hue variable. Batholith soils again are 100 percent hue 10 YR, except 95 percent in

Table 9. Frequency distribution of moist soil color† by horizon in soils on the Kootenai Formation (KF) and Boulder batholith (BB).

Horizon	Parent Material	Hue			Values								Chroma							
		5 YR	7.5 YR	10 YR	1	2	3	4	5	6	7	8	1	2	3	4	5	6	7	8
		%																		
A1	KF	24	43	33	—	9	62	19	—	—	—	—	—	33	29	38	—	—	—	—
	BB	—	—	100	—	63	38	—	—	—	—	—	56	31	13	—	—	—	—	—
A2	KF	—	22	78	—	—	—	11	49	45	—	—	—	—	44	45	—	11	—	—
	BB	—	—	100	—	—	69	23	8	—	—	—	15	31	54	—	—	—	—	—
B2t	KF	29	33	38	—	—	10	32	48	10	—	—	—	—	9	38	—	48	—	5
	BB	—	5	95	—	—	16	68	11	5	—	—	—	5	37	53	—	5	—	—
C	KF	25	42	33	—	—	—	17	25	33	25	—	—	—	50	42	—	—	—	8
	BB	—	—	100	—	—	—	27	59	9	5	—	—	5	18	63	5	9	—	—

† Munsell Color Chart

the B2t horizon. Kootenai soils are fairly evenly distributed between hues 5 YR, 7.5 YR and 10 YR. Moist soils are redder than dry soils with a higher frequency of 5 YR and 7.5 YR hues. The consistent moist and dry hues in batholith soils reflects the greater uniformity of batholith rock mineralogy in contrast to the variable Kootenai Formation lithologies.

Soil Structure

Table 10 shows the frequency distribution of soil structure grade, size and type. Structure grade distribution in the A1 horizon is roughly 2/3 moderate and 1/3 weak in Kootenai soils, whereas in batholith soils it is about 2/3 weak and 1/3 moderate. In the A2 horizon of soils of both parent materials structure grade is predominantly weak. Moderate and strong structure grades dominate in the B2t horizon of Kootenai soils, whereas batholith soils have 100 percent moderate structure grade. The C horizon of both Kootenai and batholith soils have a structureless grade because the structure type is massive. The stronger structure grade of Kootenai soils reflects the higher mean clay content, which will be discussed later.

Structure size classes for Kootenai soils are very fine, fine and medium in the A1 horizon; fine and medium in the A2; and medium and fine in the B2t. Batholith soils have medium and fine classes in the A1 horizon; very coarse and medium in the A2; and dominantly medium