Hydrologic Properties of Soil and Subsoil on a Steep, Forested Slope

by

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A THESIS

submitted to

Oregon State University

in partial fulfillment of
the requirements for the
degree of

Master of Science

June 1974

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Date thesis is presented March 22, 1974

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AN ABSTRACT OF THE THESIS OF

Darrel Wesley Ranken for the Master of Science
(Name) (Degree)

in Forest Engineering presented on
(Major) (Date)

Title: HYDROLOGIC PROPERTIES OF SOIL AND SUBSOIL ON A STEEP, FORESTED SLOPE

Abstract approved: R Dennis Harr
R. Dennis Harr

The purpose of this study was to examine certain hydrologic properties of the soil and subsoil on a steep forested slope and relate these properties to the movement of water via subsurface routes. The hydrologic properties examined were bulk density, soil texture, total porosity, pore size distribution, saturated hydraulic conductivity, and soil moisture-tension relationships.

Soil samples were taken from a 2.5 ha study slope on watershed 10 of the H. J. Andrews Experimental Forest near Blue River, Oregon. Eleven soil pits were excavated and six soil core samples were taken at depths of 10, 30, 70, 110, 130, 150, and 200 cm where soil conditions permitted. Laboratory analyses were conducted to determine the hydrologic properties of each sample. The extreme permeability and high porosities of the samples necessitated the use of specially designed apparatus to measure the saturated weights and

hydraulic conductivities.

Particle size distribution changed only slightly with depth. The A and B horizons were predominately clay loams and the C horizons were classified as clays. Total porosities also varied little with depth. The porosity of the soil (A and B horizons) averaged nearly 65% while the porosity of the subsoil (C horizons) averaged nearly 55%. Bulk density also varied little with depth. Soil bulk densities averaged .825 gm/cm^3 and subsoil bulk densities averaged 1.180 gm/cm^3.

The hydraulic conductivity and pore size distribution of the soil and subsoil were well correlated and changed considerably with depth. Significant decreases in the hydraulic conductivities occurred between the 30 cm and 70 cm depths as well as between the 110 cm and 130 cm depths in some of the soil pits. At most soil pits the surface soil had conductivities greater than 400 cm/hr while the soil at the 70 cm and 110 cm depths had conductivities near 200 cm/hr. Subsoils had much lower conductivities, less than 60 cm/hr in most soil pits and less than 10 cm/hr in some pits. A power curve regression analysis was used to relate the hydraulic conductivity ($\hat{Y}$) and the mean percentage of pores greater than .294 mm in diameter (X) according to the equa- $\hat{Y} = 10,040X^{2.997}$. The resulting r^2 was .945. The percentage of pores greater than .294 mm in diameter was also found to change abruptly between the 30 cm and 70 cm depths in most soil pits and between the 110 cm and 130 cm depths in some soil pits.

The hydrologic properties were used to discuss the possible nature of water movement through the soil and subsoil of the study slope. The soil hydrologic properties and antecedent moisture conditions were predicted to be conducive to vertical unsaturated translatory flow. A zone of saturation was predicted to occur during winter rainfall events above the subsoil horizon having extremely low conductivity rates (above the 130 cm depth near soil pit 1). This zone of saturation was predicted to be the most probable zone of lateral water movement in the form of saturated translatory flow.

Data from a soil pit known to have saturated flow over the subsoil and from tensiometers installed near the soil pits were presented as evidence that a zone of saturation does exist within the subsoil during some rainfall events and that the soil and subsoil moisture conditions are conducive to translatory flow during the winter rainy season.

ACKNOWLEDGEMENTS

My gratitude is expressed to the U.S. International Biological Program for the financial assistance that made this study possible.

Appreciation is extended to my major professor, R. Dennis Harr, who provided substantial advice, guidance, and assistance throughout the research and preparation of this thesis.

Gratitude is expressed to my fellow graduate students for their help and guidance.

I also wish to extend special appreciation to my parents and brother, Randy, who have provided the constant faith and encouragement that were necessary for the completion of my graduate work.

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INTRODUCTION

The rising demand for the Northwest's high quality water has increased the necessity for intense management of our water resources. Knowing where the water is and where and how fast it is moving is required for the proper management of this valuable resource. In the western Cascades of Oregon the prevailing hydrologic process is the subsurface movement of water commonly called interflow or subsurface flow. A greater understanding of this important hydrologic phenomenon would aid water resource management.

The combined effects of the Northwest's geologic history, climate, and vegetation have produced a unique soil-water relationship. Long duration, low intensity rainstorms result in the rapid rise of stream hydrographs with the maximum rate of runoff in some instances approaching 80 percent of the average rate of precipitation for the preceding 12 to 24 hours (Rothacher, Dyrness, and Fredriksen, 1967). Although streams respond quickly to precipitation virtually no overland flow of water has been observed on undisturbed slopes. Nearly all of the precipitation reaching the soil surface passes through the soil and subsoil before entering the open channel drainage system. Considering the percentage of rainfall that has been accounted for in the storm hydrographs, the flashy response of the streams to the onset of winter storm events must be attributed to some form of rapid subsurface flow.

The subsurface movement of water also influences other physical phenomena in the forest ecosystem. In addition to contributing to storm flow subsurface water seepage is also responsible for sustained flow of streams during the summer. Subsurface flow is the mechanism by which dissolved chemical constituents are removed from the soil and the forest nutrient cycling system (Borman and Likens, 1967). During storm periods in some regions the concentration of water above a relatively impervious layer within the soil profile has resulted in pore water pressures sufficient to reduce total effective weight of a soil mass. This reduction decreases the shear strength of the soil and can trigger mass movements of soil on steep slopes (Swanston, 1970).

The precise nature of the mechanism of the subsurface movement of water through the soil and subsoil of the western Cascades is not fully understood. Rothacher et al. (1967) have suggested a shallow and rapid lateral movement of water through the soils and subsoils on the steep slopes of this region. Quantitative information concerning the physical properties of the soil and subsoil is insufficient to describe the nature of this movement.

The objectives of this study were to examine certain hydrologic properties of soil and subsoil on a steep forested slope and relate these properties to the movement of water via subsurface routes. Specific hydrologic properties examined included: bulk density, soil texture, total porosity, pore size distribution, saturated hydraulic conductivity, and soil moisture-tension relationships.

LITERATURE REVIEW

Energy Conditions Affecting Soil Water Movement

Total Potential Energy

The movement of water through soil and subsoil is controlled by its energy state. The potential energy of soil water often varies from one point in the soil to another. Like all matter soil water moves from points of higher potential energy to points of lower potential energy. The difference between two unequal potential energies divided by the distance between them is the primary moving force of soil water. This change in potential energy with distance is known as the negative potential gradient. The gradient is negative because the force is acting in the direction of decreasing potential (Hillel, 1971).

The total potential is the sum of three separate potentials: the gravitational potential, the pressure (or matric) potential, and the osmotic potential. An osmotic potential gradient requires a semi-permeable membrane for the movement of a liquid to occur. Because this condition is not applicable to rapid subsurface movement of water through the soil the osmotic potential will not be discussed further.

The gravitational potential of the soil water at any point is dependent on the elevation of that point above an arbitrary reference level. The reference level is usually established at a level allowing the gravitational potential to be positive or zero. The magnitude of this potential energy is dependent only on the density of water, the volume it occupies, the force of gravity, and the elevation of the water above the reference plane.

The pressure potential of any point can be either positive or negative depending on the position of the point relative to a free-water surface. A point below the surface will have a positive pressure potential, a point at the surface will have a zero pressure potential, and a point above the surface will have a negative pressure potential known also as "matric potential," "matric suction," or "tension." The potential energy of a submerged unit volume is given by the equation

$$p = \rho g h$$

where ρ is the density of water, g is the acceleration due to gravity, and h is the distance of the point below the free water surface. A point above a free-water surface will have a potential energy below that of bulk water because of the capillary and adsorptive forces binding the water to the soil particles. The magnitude of these forces is dependent on the surface area of the soil particles, the surface tension of water, and the contact angle between the water and the particles. In equation form the negative pressure potential (P) is given by the equation of capillarity:

$$p_o - p_c = P = \gamma\left(\frac{1}{R_1} + \frac{1}{R_2}\right)$$

where p_o is atmospheric pressure, conventionally taken as zero; p_c is the pressure of the soil water, which can be negative; γ is the surface tension of water; and R_1 and R_2 are the principal radii of curvature of a point on the meniscus (Hillel, 1971).

Darcy's Equation

The movement of soil water, as discussed above, is the result of a potential energy gradient between points or zones within the soil matrix. The quantitative relationship between the movement of a liquid and the potential energy gradient was first formalized into an equation by Henri Darcy in 1856. The equation reads:

$$q = -K\Delta H$$

where q is the volume of liquid flowing through a cross-sectional unit area per unit time (flux), K is the conductivity of the porous medium, and $-\Delta H$ is the negative potential energy gradient called the "hydraulic gradient" here. In terms of the soil-water medium the movement of soil water is proportional to the conductivity of the soil, and is also proportional to and in the direction of the hydraulic gradient (Hillel, 1971).

Hydraulic Gradient

Hydraulic gradient is the same as the gradient of total potential when osmotic forces are ignored. The hydraulic gradient between two points in the soil is the difference between the hydraulic heads of the two points divided by the distance between the two points. The total hydraulic head is synonymous with total potential. Where total potential is the sum of the pressure potential and the gravitational potential the total hydraulic head is the sum of the pressure head and the gravitational head

The gravitational potential gradient will effect only the movement of soil water in the vertical direction. The pressure potential of the soil water at any one point in a soil of uniform texture and structure is determined primarily by the moisture content of the soil at that point. A saturated soil will have a positive pressure potential and an unsaturated soil will have a negative pressure potential (tension). Thus in a horizontal column of soil, where the gravitational potential gradient is zero, water flows from a saturated zone (higher potential) to an unsaturated zone (lower potential). Once the water leaves a zone of positive pressure potential it is believed to flow along in films and menisci formed between the soil particles and not through the soil pores. Water moves from thick water films and less curved capillary menisci where the tension is low, to thin water films and more highly curved capillary menisci where the tension is high (Hillel, 1971).

The rate of water movement in both saturated and unsaturated soils is influenced by the hydraulic gradient. The steepness of the total hydraulic gradient will help determine the rate of water movement. A large difference in pressure potentials or tensions over a short distance will mean a more rapid movement of water than a smaller difference in both saturated and unsaturated soils. However, the rate of water movement is affected more by the hydraulic conductivity than by the steepness of the hydraulic gradient (Hillel, 1971).

Hydraulic Conductivity

The conductivity of a soil mass is simply the conductance of the soil pore system. From the Darcy equation conductivity is defined as the ratio of the flux to the hydraulic gradient. The dimensions of conductivity are expressed in length per unit time, such as cm/hr or cm/sec. Conductivity in the soil-water medium is affected by the soil moisture content, the nature of the soil pores, and to a lesser extent, the viscosity of the soil water.

The hydraulic conductivity of a soil is at its highest when the soil is saturated. With a reduction of soil moisture below saturation the conductivity decreases sharply. The area of flow is greatly reduced and the water is forced to find pathways around air gaps created in the soil pores. A soil with large pores has a high saturated conductivity but the large pores are the first to be drained and the conductivity of such a soil decreases rapidly when unsaturated. A soil with many small pores has a higher conductivity under the same unsaturated conditions since the smaller pores will not have drained and will still be conducting. Thus soils that are more capable of a rapid saturated flow may not be conducive to rapid unsaturated flow (Hillel, 1971).

Nature of Soil Water Movement

Studies of subsurface flow have had to include an investigation of the porous media through which the water passes. Hydrologists have often examined the processes of subsurface flow by studying flow processes in an experimental plot on an undisturbed slope. This technique supplies results for only the soils tested and are not necessarily applicable for other soil conditions. Horton and Hawkins (1965) attempted to define some basic flow phenomena that could be applied to all soil conditions. They conducted laboratory experiments on the nature of vertical water movement from the soil surface to a zone of saturation. Using artificially packed soil columns, they found that when the soil is near the field capacity state, water that infiltrates large pores will flow toward smaller pores due to the potential gradient between them. Using a radioactive tracer they found that water moved through a vertical column by displacement. A volume of water containing the tracer was added to the top of a column of soil near field capacity. Additional volumes of water were added each day and the effluent analyzed for the tracer. Only after 87% of the original water in the column had been displaced did water containing the tracer emerge.

Studies of the nature of subsurface water movement and its contribution to the rapid rise of the stream hydrograph have been conducted only within the last ten years. Very few of the investigators have been able to agree on the flow path of subsurface water or the extent to which subsurface water contributes to storm hydrographs. Perhaps one reason for some of the differing opinions is the differing soil conditions with which the investigators have worked. The flow path of water and its flow rate are controlled by the hydrologic properties of the soil which can vary with physical and biological characteristics of the soil.

Whipkey (1965, 1967, 1968) studied the nature of subsurface flow in the Allegheny-Cumberland plateau of the eastern United States. He found that subsurface flow of water through macropores in the soil matrix was responsible for stormflow. These macropores were formed by biological activity and consisted of worm holes, old root channels, or structural openings in the soil matrix. Whipkey stated that the hydraulic conductivities of the soil matrix in finer textured soils were too low to account for the rapid movement of water as mass interflow. However, the flow through coarse textured soils was found to be a function of the hydraulic conductivity. He concluded that the finer textured soils contribute to rapid subsurface stormflow through interconnecting cracks and channels, and that the coarse textured soils did not contribute to storm runoff.

Aubertin (1971), working on one of the plots used by Whipkey (1965), related rapid lateral movement of subsurface water to the presence of macropores formed by animal burrowing and root penetration into the soil. He noted that on his study plots the soil mass was not saturated before water entered and flowed through the large macropores. He attributed this to the moderately fine texture of the soil mass and the funneling of the water into the interconnecting system of subsurface channels. The hydraulic conductivity of the surface six inches of the soil was not exceeded by the rate of application of artificial rainfall. The conductivity of the soil mass below six inches was far less, due primarily to the texture and structure of the soil. As a result rainfall penetrated the surface six inches of soil rapidly and then flowed into the macropores and channels of the underlying soil.

Studies conducted in the Southeast have found that flow processes other than flow through large macropores are operative in this area. Wilson and Ligon (1973) investigated the interflow process on a gently inclined slope in the South Carolina Piedmont. Interflow (subsurface flow) and runoff from a 0.21 acre plot were measured during natural and artificial storms. Soil moisture and soil moisture tension were also monitored. Analyses indicated the surface horizon had an average saturated hydraulic conductivity more than seven times that of the underlying "B" horizon. The volume of interflow found to occur was only 10% of the volume of surface runoff occurring during the same storm events. The authors concluded that subsurface movement of water occurred only after the highly permeable surface horizons of the soil profile were at or near field capacity. They noted that these soil moisture conditions were representative of the watershed in which they worked and of any Piedmont watershed in general.

Hewlett and Hibbert (1967) investigated processes of subsurface flow through the forested soils of the southern Appalachians. They attributed the rapid rise of the storm hydrograph to contributions from a variable source area. They state that "quick flow", that portion of the stream hydrograph containing the storm peak, is due to precipitation falling directly into the stream along with rapid subsurface movement of water in soil close to the stream channel. The extent of the area contributing subsurface quick flow varies with rainfall amount and antecedent soil moisture conditions. This subsurface flow is termed "translatory flow." Rain from an on-going storm will displace water already in the soil if the soil is near field capacity. Water that enters the soil will cause a pulse-like displacement of antecedent soil water, and the same volume of water will enter the open channel as quick response subsurface flow. This type of flow was found to occur primarily on the lower and mid-slopes of the small watershed. Thus water from an on-going storm does not have to travel the entire length of the slope to contribute to the rapid rise of the stream hydrograph.

Other investigations of subsurface flow have discounted its importance as a substantial contributor to the stormflow portion of the hydrograph. These hydrologists feel that subsurface flow is slow and non-responsive to precipitation, and that some form of surface runoff is responsible for the rapid rise of the hydrograph during storm events. Dunne and Black (1970a, 1970b) investigated the runoff processes on a brown podzolic soil in northeastern Vermont. These authors considered the soils and topography of this region to be conducive to subsurface stormflow. However, despite favorable soil physical and moisture conditions, subsurface movement of water was too small, too slow, and too insensitive to changes in precipitation to contribute to the stormflow portion of the stream hydrograph. They concluded that runoff on small watersheds was influenced primarily by stream channel interception and overland flow near the stream channel.

Freeze (1972) used a mathematical model to simulate streamflow and subsurface flow which provided theoretical support for the field studies conducted by Dunne and Black. The simulation model showed that only with the special conditions of high saturated conductivity and steep convex slopes would subsurface flow contribute to the storm hydrograph. The simulations carried out by Freeze showed that most storm runoff is the result of "direct runoff through very short overland flow paths from precipitation on transient near-channel wetlands."

In a study conducted in England Weyman (1973) attempted to discover the existence of subsurface lateral flow and the extent to which lateral flow contributed to the storm hydrograph. A system of trenches and lateral troughs, similar to that used by Whipkey (1965), was used to collect subsurface flow from a gently sloping (2^{o}-23^{o}) study slope. Weyman found that after the initiation of rainfall the precipitation traveled vertically through the soil profile in the unsaturated condition. After continued rainfall, a zone of saturation moved upwards from the base of the slope. Saturated lateral flow occurred through this saturated zone. He concluded that a distinct break in soil horizons or an impermeable layer was necessary for the zone of saturation to form and lateral movement of water to occur. He also concluded that the time required for vertical percolation of water under unsaturated conditions and the low saturated hydraulic conductivity in the lower levels of the soil prevented the possibility of rapid subsurface flow that could contribute to the storm hydrograph. Weyman stated that the primary contributor to the storm hydrograph is most probably some form of rapid surface runoff, or in some instances a non-Darcian rapid flow of water through soil macropores.

Soil Characteristics Affecting Water Movement

Investigations of the subsurface movement of water began with the study of the process of water infiltration and percolation through the soil. L. D. Baver was one of the first to study soil characteristics influencing the movement of soil moisture. In an early study (Baver, 1936) he found that the principal soil characteristics influencing the downward movement of water under gravitational forces were the volume and continuity of the non-capillary soil pores as influenced by soil texture, soil structure, and biological channels. Non-capillary porosity was understood to mean the proportion of the total porosity in pores that would permit the percolation of water by gravitational forces. Antecedent soil moisture and soil air resistance also influenced this water movement.

Later, Baver (1938) studied the effects of the non-capillary pore space at various soil moisture tensions on the rate of water movement. He found that the permeability of a soil is directly related to the tension required to drain the pores. In this study he used what was later to be termed the "soil moisture characteristic curve" (Childs, 1940) to determine the tension at the lower limit of the non-capillary pores--the "flex point." The flex point is the point on the soil moisture characteristic curve having the least slope. This point was used to separate the tensions holding water in non-capillary pores from tensions holding water in capillary pores.

Nelson and Baver (1940) continued the study of the relation between the nature of soil pores and water movement. They also concluded that the permeability of the soil is related to the volume of non-capillary pores, the tension required to drain these pores, and the continuity of pore space. However, they went on to claim that the pore size distribution of the non-capillary pore space was also a controlling factor. The relation between the size distribution of the soil pores and permeability was later refined by Marshall (1958). He found that data from the soil moisture characteristic curve could be used in an empirical equation to calculate permeability.

The hydrologic properties of the soil that are most pertinent to the study of subsurface water movement were outlined for the Southern Appalachians by Hursh and Hoover (1941). These authors stated that practically all of the physical characteristics affecting water movement and storage in a soil profile can be defined in terms of pore space. These characteristics included hydraulic conductivity and the capillary and non-capillary pore size distributions. Detention and retention storage capacities were important properties calculated from the pore size data. The authors also recognized the importance of biological pathways, the antecedent soil moisture, and the continuity of the soil pore system. They suggested that non-capillary porosity and retention storage opportunity (a measurement of the volume of pores available to retain water by capillary forces) were the two most important soil profile characteristics that could be measured for use in hydrologic studies.

From here the permeability as affected by soil characteristics studies expanded to consider a wider range of soil conditions. Such were the studies on the effect of the least permeable layer on water flow through a soil profile (Swartzendruber, 1960), and on the nature of unsaturated flow through non-uniform soil profiles (Zaslavsky, 1963).

Although most of the studies of hydrologic properties of soils have been related to the nature of subsurface flow some investigators have used the examination of these soil characteristics for other objectives.

Hoover (1950) examined the effects of prolonged cultivation on the hydrologic properties of South Carolina Piedmont soils. The properties examined included infiltration rate, hydraulic conductivity, and retention and detention storage capacities. He used these values to illustrate the changes brought about by cultivation: a decrease in infiltration rate, a subsequent increase in erodibility, and a reduced water storage capacity in the soil profile.

Dyrness (1969) studied the hydrologic properties of the soils on the H. J. Andrews Experimental Forest in the western Cascades of Oregon. He attempted to find a better soil classification system that could be used for hydrologic interpretations. For each soil type he determined percolation rate, soil porosity characteristics including capillary and non-capillary proportions of total porosity, and the retention and detention storage capacities. After finding that the most important soil factor governing the variation of these hydrologic properties was stone content, Dyrness recommended that any attempt to classify or map the soil types of this area should include an estimation of the stone content of the soils.

Hydrologic properties of soils were put to a different use by Wang (1970) in soils of the South Carolina Piedmont. He studied the areal variation in the hydrologic properties of the Piedmont soils in order to develop a method of predicting interflow with the use of measurable soil characteristics. Soil characteristics examined included bulk density, saturated hydraulic conductivity, moisture content and tension relationship, particle size distribution, and unsaturated conductivity-tension relationships. The results indicated that there was no significant areal variation of these hydrologic properties. Because of the areal uniformity of the hydrologic properties Wang concluded that interflow volumes and rates could be accurately predicted.

DESCRIPTION OF THE STUDY AREA

Location

The study area is located in the H. J. Andrews Experimental Forest approximately 70 kilometers east of Eugene, Oregon, near the town of Blue River (Figure 1). A 2.5 ha portion of the south-facing slope of experimental watershed 10 was selected for study. This area is bounded by the stream at the bottom of the slope and by the ridge line at the top of the slope. Elevation of the study area varies from 440 m to 535 m. The slope at the bottom of the study area exceeds 100%, while the upper slope is 50 to 60%.

The climate, vegetation, geology, soils and general hydrologic properties of this watershed are representative of conditions found on other watersheds in the western Cascades (Rothacher *et al.*, 1967). The south aspect of watershed 10 was selected because of its accessibility by trail, its location relative to the stream gage, and its soil depth.

Climate

The climate of the western Cascades is influenced primarily by the presence of the Pacific Ocean only 160 kilometers to the west. The annual distribution of precipitation and the range in temperature are typical of the maritime climate. The region has wet, relatively

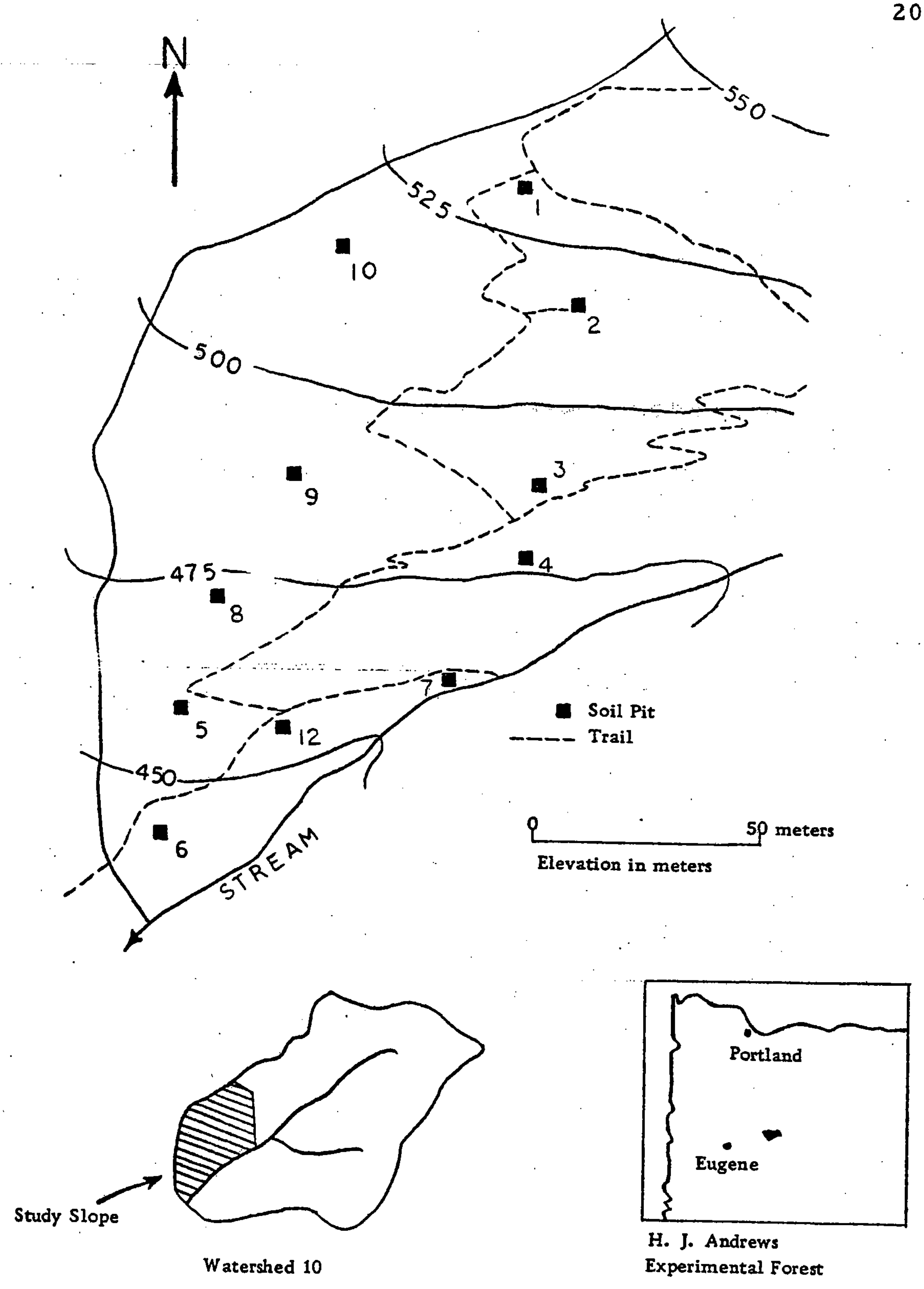

Figure 1. Map of study slope.

mild winters, with dry, cool summers. The temperature occasionally reaches extreme lows of -18° C in the winter and highs of 38° C for short periods of time in the summer. The mean annual temperature is 9.5° C. The mean January temperature is 1.7° C, while the mean July temperature is 20.7° C (Rothacher et al., 1967).

The mean annual precipitation of the H. J. Andrews Experimental Forest for the period 1952-1972, is 234.3 cm. About 87% of the annual precipitation falls during the rainy season from October to April (Rothacher et al., 1967). During this season warm, moist air masses move inland from the ocean. Most winter storms are of long duration, low to moderate intensity rainfall. The record storm recorded for this area produced more than 33 cm of rain over a four day period (Fredriksen, 1965). Normal storms consist of two or three days of almost constant low intensity rainfall and several additional days of intermittent rainy periods. Storm events of this nature often follow one another with few rainless days between events.

Although most of the winter precipitation falls in the form of rain, relatively light accumulations of snow are common on the higher elevations of watershed 10 each year. High peak flows have occasionally resulted when storms accompanied by warming temperatures have quickly melted large quantities of snow (Fredriksen, 1965).

Vegetation

The H. J. Andrews Experimental Forest lies within the western hemlock habitat (Franklin and Dyrness, 1969). Watershed 10 is vegetated with the tree and shrub species normal to this habitat type. The vegetation on the watershed is quite variable due to differences in radiation, elevation, and soil depth.

Overstory tree species include 450 year old Douglas-fir and western hemlock on all elevations, and scattered sugar pine on the upper slope. The understory of the lower slope consists primarily of vine maple, rhododendron, Oregon grape, and sword fern, while the mid and upper slopes have chinquapin, rhododendron, and bear grass as understory species.

Geology and Soils

The western Cascade mountains are the remnants of old volcanic flows and pyroclastic rock laid down during the stratigraphic period of the Oligocene Epoch and the great orogeny that began in the middle of the Miocene Epoch (McKee, 1972). Rock types found in these mountains include andesite, basalt, tuffs, and breccias.

There are three principal soil types that occur in the Blue River area. A residual clay loam soil generally found high on steep slopes and ridgetops was formed from andesite and basalt. Another residual soil, more characteristic of midslope positions, has a silty clay loam texture and is commonly unstable and sensitive to disturbance. This soil is formed from agglomerates, tuff and breccia. The third soil type is a colluvial clay loam common to gentle slopes and benches (Berntsen and Rothacher, 1959).

The soils of the study slope on watershed 10 are residual or colluvial Regosols, having developed primarily from red breccias. The soil type found on the study slope was identified as a Frissell by Fredriksen (1968). This series is characterized by gravelly clay loam textures with depth to bedrock varying from 0.5 m to over 3 m (Rothacher et al., 1967).

Soil profile development has been slow, and as a result, clear-cut horizons are difficult to distinguish. The soils are shallow from a pedological point of view. However, as a result of the extreme weatherability of the underlying breccia, varying depths of soft, highly porous saprolite underlie the soil profile. Because of the nature of this porous media the soils can be considered to be deep from a hydrologic point of view.

METHODS AND MATERIALS

Soil Sampling Methods

Soil Pits

Soil and subsoil samples were taken from 11 soil pits located on the study slope, and from a failure zone of a recent slope failure located on a nearby watershed. Samples were taken from the latter location to determine the nature of the hydrologic properties of subsoil over which saturated subsurface flow was occurring and on which failure had occurred. Hydrologic properties of this subsoil were compared with those of the study area to determine if the subsoil on the study area had hydrologic properties conducive to saturated subsurface flow. Soil pits located on the study slope were positioned in a rough grid pattern (Figure 1). Soil pits 5, 6, 7, and 12 were located near the bottom of the slope to provide a greater sampling of the more variable soil conditions found there.

The depth of each soil pit varied according to the conditions of the soil and subsoil. The original plan called for digging down to unweathered rock, but the depth of the subsoil and weathered saprolite proved to be beyond practical limits of excavation. Therefore, none of the soil pits were excavated deeper than 2.5 m. The amount of relatively unweathered rock within the soil profile made both digging and soil sampling difficult in several cases. Sampling was limited due to excessive rockiness at lower depths in soil pits 7, 9, and 10. Soil in pits 4 and 8 was too rocky to allow any sampling at all.

A profile description for each soil pit was made using the guidelines in the Soil Survey Manual (Soil Conservation Service, 1967) and existing soil profile descriptions for the study area (Fredriksen, 1968; Brown, 1973). The soil profile descriptions for each soil pit are given in Appendix C.

Soil Sampling

Two types of soil and subsoil samples were taken. First a bulk density core sampler (Blake, 1965) was used to provide undisturbed samples to be used for hydraulic conductivity, bulk density, pore space, and soil moisture-tension tests. The sampling instrument employed a brass retainer ring (6 cm x 5.4 cm in dia.) fitted inside a stainless steel cutting cylinder. Additional brass spacer rings fitted both above and below the soil retainer ring. This arrangement of retainer and spacer rings was used to minimize the disturbance of the soil sample. Although impossible to extract a completely undisturbed sample, this method has proven the most effective to date.

Soil and subsoil samples were taken as the soil pits were being excavated. The first samples were taken at the surface of the mineral soil directly below the organic horizons. The soil pit was then enlarged and deepened to the next sampling level, 30 cm below the mineral soil surface. After each depth had been sampled the soil pit was deepened to the next sampling depth. Subsequent sampling depths were at 60-70 cm, 100-110 cm, 140-150 cm, and every 50 cm beyond 150 cm. Preliminary laboratory analyses indicated a need for sampling at the 130 cm depth and for further sampling at the 0-10 cm and 20-30 cm depths in some pits. A total of 452 samples were taken from all soil pits.

The procedure used to take the soil samples was essentially the same as given by Blake (1965). The samples were taken in a vertical direction a measured distance below the soil surface. This was done to ensure that the hydraulic conductivity measured for each sample was in the direction of most probable water movement. To take an individual sample the cylinder containing the soil retainer ring was slowly hammered into the soil. The sample was removed from the soil by inserting a trowel beneath the open end of the cylinder. This procedure ensured that none of the soil would fall out of the retainer ring. The use of the trowel was necessary only for the samples taken from the surface 30 cm. Soil and subsoil below this depth were more cohesive and were adequately retained by the ring. The soil and subsoil was most easily sampled when the soil was moist. Dry soils were found to be extremely non-cohesive and difficult to sample using this method.

After the sampler was removed from the soil and the retainer ring was taken from the sampler, excess soil from the ends of the sample was removed using a large pocket knife. The soil was trimmed flush with the top and bottom of the retainer ring and roots extending from the soil sample were snipped with a fingernail clipper. When stones or large roots protruded beyond the ends of the retainer ring, the sample was discarded and replaced by another sample.

Next a double layer of cheese cloth was placed over the bottom end of the retainer ring and secured with a rubber band. The sample was then placed bottom down in a soil can whose lid was firmly held in place with another rubber band.

Six soil samples were taken from each sampling depth. At the time of sampling appropriate descriptive information concerning the soil pit and the depth of sampling was recorded.

The second type of soil sample taken was a grab sample of about 500 cm^3 from each depth from which core samples had been removed. These grab samples were later used to determine the particle size distribution of the soil and subsoil.

Soil Laboratory Analysis Methods

Sample Preparation

The samples were covered and placed in cold storage (4^{o} C) until ready for use. Cold storage retarded biological activity which might have altered certain hydrologic properties of the soil.

Time limitations and the capacity of the testing apparatus allowed only 24 samples to be analyzed together. In order to provide uniformity in the testing procedures, all samples from one depth were grouped together as the tests were conducted.

Excess cheese cloth was removed from each retainer ring, and the samples were placed in a large, deep-walled, stainless steel pan for saturation. Distilled water was added to the pan until the water level was approximately 1 cm below the top of the retainer rings. The samples were then allowed to stand in this position for 16 hours to ensure complete saturation and minimize the amount of entrapped air. The water level was then increased to a depth 10 cm above the retainer rings. The samples at this point were fully prepared for analysis.

Hydraulic Conductivity

The first hydrologic property determined was hydraulic conductivity. For these measurements a constant-head permeameter was constructed (Figure 2). This permeameter consisted of a support frame, inlet and outlet chambers, and a constant head reservoir. A screen was attached to each chamber to provide support for the soil samples. The support frame of the permeameter was placed underwater with the samples. The water inlet chamber was positioned in

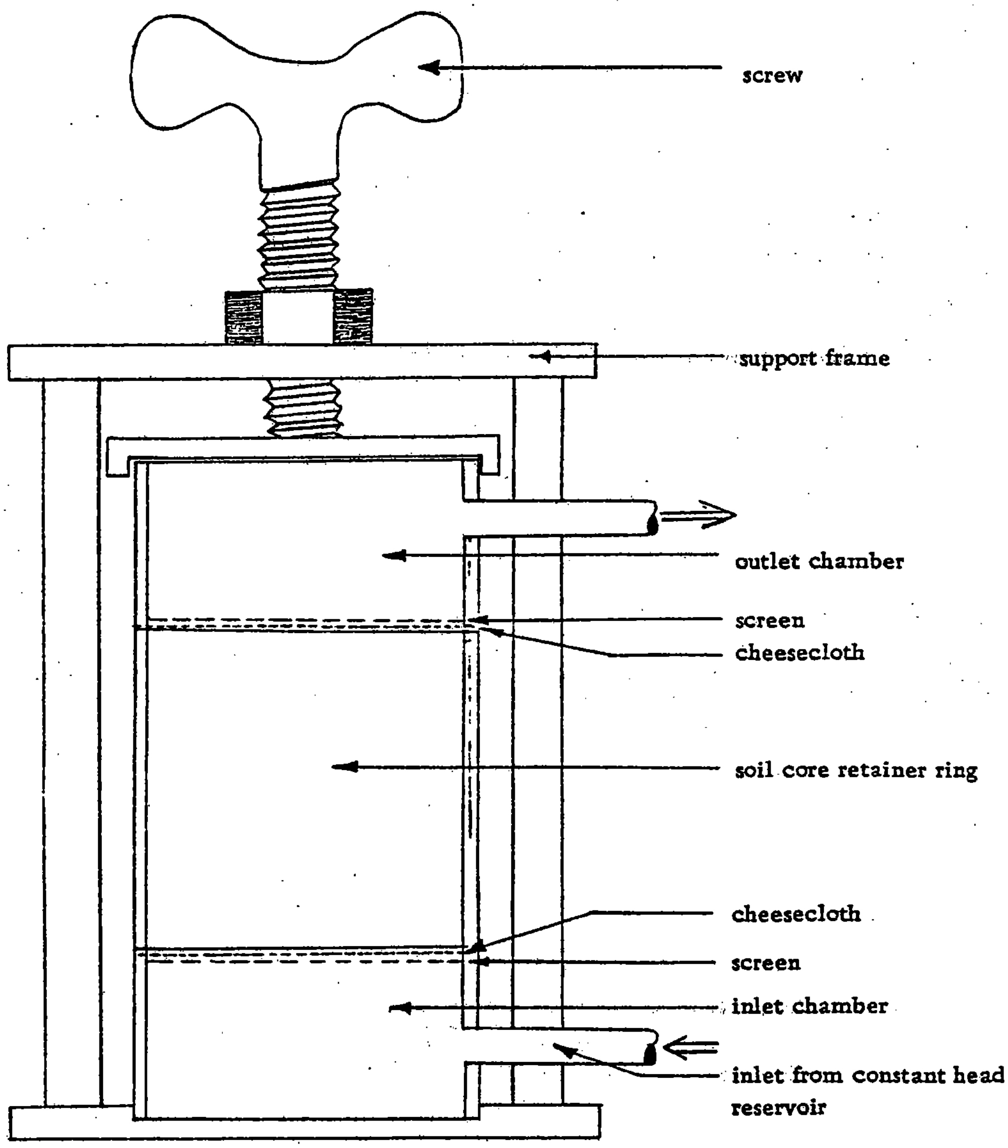

Figure 2. Constant head permeameter frame with soil core retainer ring in place.

the frame and purged of air bubbles. Next a rubber band was used to fasten a double layer of cheese cloth over the top of the sample to be tested. The sample was then placed in the permeameter frame in an inverted position so that the top of the sample was adjacent to the inlet chamber, and the bottom of the sample was adjacent to the outlet chamber. Thus the water flowed through the sample from top to bottom.

After the outflow chamber was placed above the soil retainer ring, the joints between the retainer ring and inlet and outlet chambers were sealed by forcing these three pieces together with the screw at the top of the frame. The permeameter assembly was then removed from the water, and the constant head reservoir was adjusted to provide hydraulic heads of 1 to 60 cm. The lowest head that would provide measureable outflows over a ten minute maximum period of time was used.

Low hydraulic heads were used to minimize alteration of the more fragile soil samples. Preliminary work showed clay and silt particles could be eluviated by using higher heads than were necessary. A hydraulic head of 1 cm was used for the 0-10 cm and 20-30 cm samples, while a 10 cm head was used for the other samples when feasible. Occasionally the subsoil samples would require a larger hydraulic head to allow collection of an adequate amount of outflow within a ten minute period.

Once the appropriate hydraulic head had been obtained, the inlet valve was opened. When the flow through the sample had become steady, outflow was collected and measured. While the length of collection time varied inversely with permeability the most frequent collection periods were 30 sec for more permeable samples, and 60 to 180 sec for the less permeable samples.

An equation based on Darcy's law was used to calculate the hydraulic conductivity of each sample tested. This equation is presented along with other calculations in Appendix A.

Sample Dyeing

In order to qualitatively describe the movement of water through the soil cores, a small amount of dye was introduced into the constant head permeameter system and allowed to pass through the sample cores. Two cm^3 of a 0.1% malachite green dye solution were injected into the inlet tubing with a hypodermic syringe. The water and dye were allowed to pass through the sample until all of the dye had entered the retainer ring, or traces of the dye were detected in the outlet tubing.

Aubertin (1971) successfully used this method to determine the major flow channels through soil core samples. He allowed his samples to air dry 24 to 48 hours before dissecting them. Preliminary tests showed this dye would also stain major passages in the soils used in this study. However, it was not possible to dissect these samples after dyeing them because the samples were subjected to other tests. Also, oven drying was necessary for certain computations; unfortunately, oven drying destroyed the malachite green dye. Therefore, the dyed samples offered no evidence of the water flow path. Because all of the samples had to be oven dried, it was impractical to continue the use of this dyeing method.

Saturated Weight

The saturated weight of each sample was needed to calculate total porosity and pore size distribution. A C-clamp apparatus was designed to hold the retainer ring and prevent water from escaping while the sample was being weighed (Figure 3).

After conductivity measurements were completed, soil samples were re-submerged in a pan of water. Each sample retainer ring was placed in the clamp underwater and sealed by clamping the apparatus. The clamp and sample were then removed and dried. By following a systematic method of removing the outside moisture from the clamp and retainer ring the apparatus could be dried to a constant level. Tests using a ring filled only with water showed saturated weight could be determined within an accuracy of $\pm$ 1 gm.

The saturated weight of the sample was recorded as the weight of the sample and the clamp. Later the tare weights of the retainer

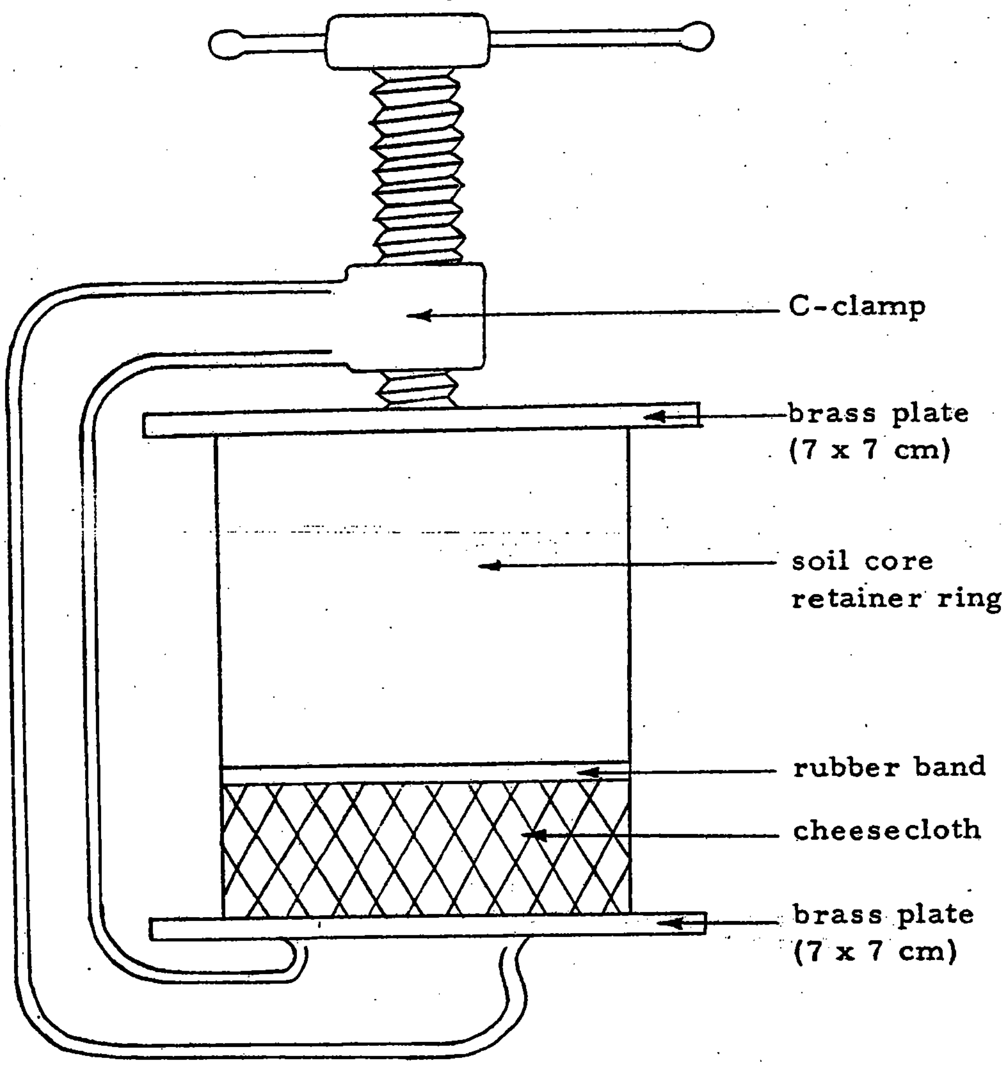

Figure 3. C-clamp apparatus.

ring and clamp were subtracted from the total weight to give the saturated weight of the soil.

Moisture Characteristics

The next step in the analysis was to determine the pore size distribution and moisture characteristic of each soil sample. Tension tables (Vomocil, 1965) were used for this purpose (Figure 4). Each table was equipped with a nylon screen 26 cm x 40 cm. Deaerated distilled water was added to cover the screen with a layer 1 cm deep. Next a 40 cm x 50 cm sheet of white blotter paper was carefully lowered into the water to prevent air bubbles from being trapped within the paper or between the paper and nylon screen.

After the blotter paper was in place a clamp on the outlet tubing was released, and the excess water on the table was allowed to drain. Next a hard rubber roller was used to smooth out the wrinkles in the blotter paper. This procedure was necessary to ensure a tight seal between the paper and the table around the outside edges of the nylon screen. A poor seal here would allow air to enter the system and break the column of water used in the tension extraction process.

In the systematic testing routine the tension tables were prepared prior to the determination of saturated weight. Then each sample was quickly transferred from the C-clamp to the blotter paper. When the clamp was released the water held in the soil at very low tension

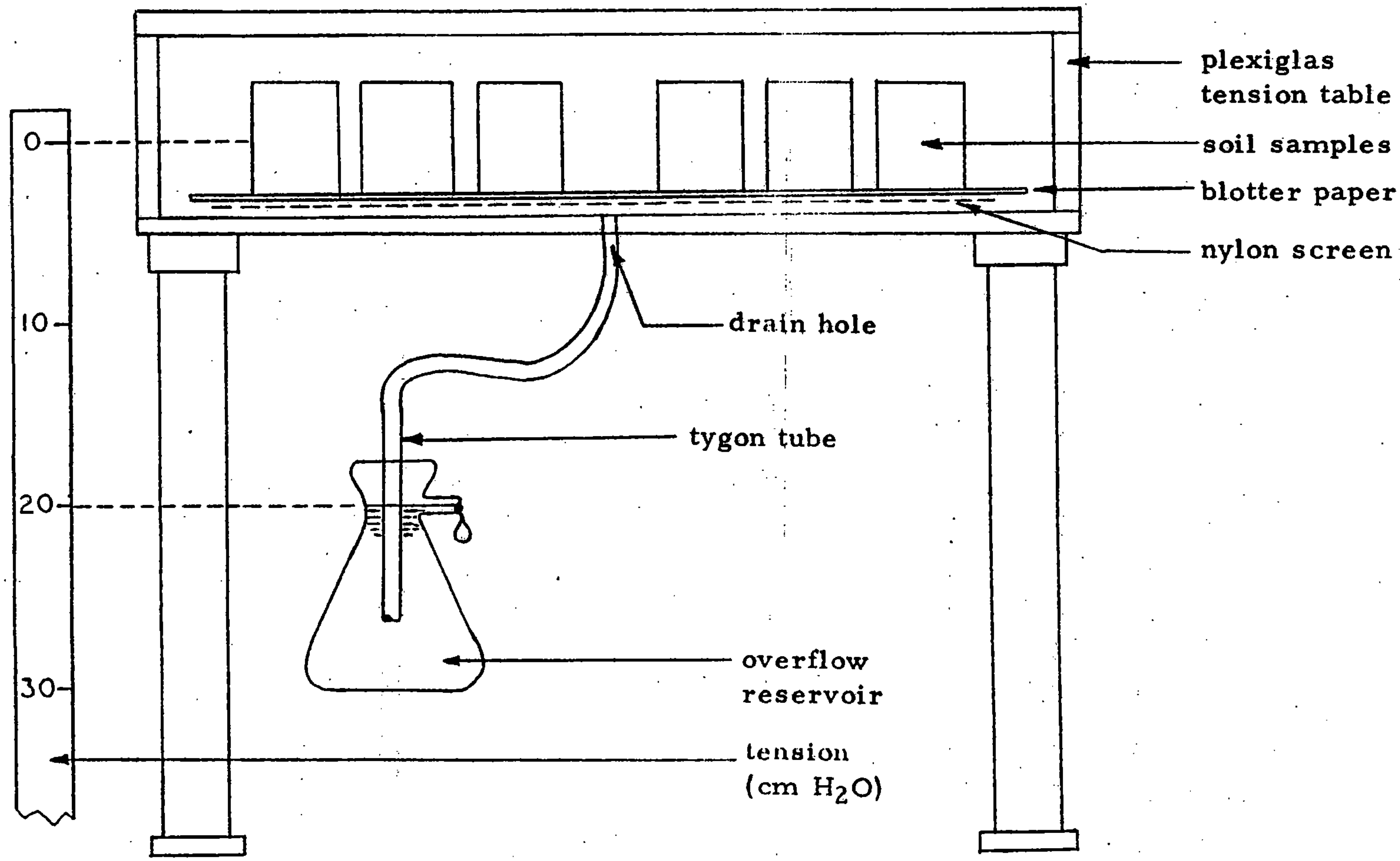

Figure 4. Tension table.

drained out of the sample. This transfer process was completed as quickly as possible.

Each tension table had a capacity of 24 retainer rings. Samples from similar depths in the profile were grouped together and placed on separate tension tables. This procedure was used to ensure that the samples would reach equilibrium with the tension being applied at approximately the same time. When all samples had been placed on the blotter paper the top of the tension table was sealed with tape to reduce evaporation of water from the samples.

The tension applied to the surface of the blotter paper was controlled by an overflow reservoir of water connected to the tension table with a length of tygon tubing. By lowering the reservoir below the level of the table, tensions of up to 100 cm of water were attained.

The outflow of the reservoir was first placed at 10 cm below the midpoint of the samples and the clamp on the tygon tubing was released. The samples were then allowed to equilibrate with the applied tension. Samples subjected to a tension equal to 10-30 cm of water required 48 hours to equilibrate. Up to 72 hours were allowed for equilibration at the higher tensions.

When apparent equilibrium had been reached the tygon tubing leading to the reservoir was clamped. Each sample was removed and any condensation on the ring was wiped off. The 10 cm weight of the sample was then determined. The samples were temporarily stored on moist paper towels while the other samples were being weighed and the table was prepared for the next tension. The paper towels prevented the cheese cloth from drying.

When new blotter paper was placed on the table and the table was prepared as stated previously, the samples were replaced, the top was sealed, and the reservoir was lowered to the 20 cm level. The cycle of increasing the tension and weighing the samples was repeated for the tensions of 30 cm, 40 cm, 60 cm, and 100 cm.

Oven Dry Weight

When the samples had been weighed after equilibration at a tension of 100 cm they were oven dried at 105°C for 48 hours. The samples were placed in soil cans for the oven drying process. After the samples were removed from the oven, the cheese cloth and rubber band were removed. The dry soil was easily separated from the cheese cloth before the tare weight of the cloth and rubber band was recorded. The weights of the soil, ring, and can were then recorded as the oven dry weight. The tare weights of the ring and can were subtracted later to calculate the oven dry weight of the soil. Typical series of calculations are given in Appendix A.

Soil Sample Structure

The validity of the foregoing measurements is dependent on a representative, undisturbed soil sample within the retainer ring. Therefore, after oven drying the samples were removed from the ring, dissected, and examined. This dissection allowed not only a check for ill-fitting samples but also an opportunity to examine each sample and record a complete description of the sample's structure. This description would provide evidence as to where the water might have passed through the sample and would help account for extreme values of hydraulic conductivity. The portion of the sample which had been in contact with the retainer ring was carefully examined for interconnecting spaces that could have allowed the free passage of water between the soil and the ring. Each sample found to have such a defect was discarded along with data pertaining to its hydrologic properties. Most of these samples were replaced by other samples taken at a later date.

Stone Volume

Before discarding the samples a No. 10 (2 mm mesh) soil sieve was used to separate stones from the soil. A 250 ml graduated cylinder containing 100 ml of water was used to measure the volume of the stones by the displacement method. This procedure worked satisfactorily for the samples taken from the upper soil depths where rock was relatively hard and unweathered. Samples below 70 cm, however, often contained varying amounts of saprolite which, when dry, were very hard and could not be broken to pass through the soil sieve. The saprolite was very porous and stored water. The problem was therefore one of defining what should be considered a stone. If the saprolite was to be considered a stone then 80% of some sample volumes would be stone. If the saprolite was not to be considered stone then less than 10% of the same sample would be defined as stones. The conclusion reached was that the stone volume of these samples could not be accurately determined. Stone volumes for samples taken from the sampling depths above 70 cm have been included in Appendix B, Table I.

Particle Size Distribution

Particle size distribution at each depth in each soil pit was determined using the grab samples collected in the field. A modified hydrometer method was used for this purpose. The samples were mixed with the dispersing agent (sodium hexametaphosphate) and distilled water in a 500 ml flask and then placed on a shaker table overnight. The standard hydrometer procedure (Day, 1965) was followed from this point. Two subsamples from each soil depth in the soil pits were analyzed. The values were averaged to determine the percentages of sand, silt, and clay. The texture class for each depth was determined using the standard soil texture triangle.

RESULTS AND DISCUSSION

Soil and Subsoil Characteristics

The complete soil profile descriptions made for each soil pit dug on watershed 10 and for the pit near watershed 9 are given in Appendix C. For the purpose of this thesis the soil was defined to be the portion of the pedon composed of the A and B horizons. The subsoil was defined to be the portion of the pedon lying beneath the A and B horizons and above bedrock. The profile descriptions resemble the description of the Frissell series as described by Dyrness (Rothacher et al., 1967). However, the soil profiles on watershed 10 include a B horizon and were generally deeper than the typical pit Dyrness described.

On the watershed 10 study slope soil depth varies from 50 cm to 110 cm. Subsoil depth varies considerably more. The subsoil is shallow or absent near the stream, but increases in depth further up the slope. Holes augered in the soil for installation of piezometers used in another study revealed that the soft weathered breccia composing most of the subsoil is several meters thick on the mid and upper slopes. However, an outcropping of unweathered bedrock does occur on the mid slope portion of the study area. Aside from localized variability the depth of the soil-subsoil profile generally increases with distance upslope from the stream.

The percentage of rocks within the soil profile also varied considerably between soil pits and with depth. The profile descriptions contain an estimate of the pebbles and cobbles exposed by the soil profile for each horizon. Generally the A horizons contain many small fragmented pebbles and many shot-like concretions. The B and C horizons contain larger fragments of unweathered rock as well as the soft, highly weathered breccia.

Two of the soil pits were found to have a very large amount of unweathered rock within their profiles. The volume of fragmented rock found in soil pits 4 and 8 was sufficiently high to prevent the use of the bulk density sampler. Soil pit 4 contained numerous large fragments of unweathered rock in a shallow soil while soil pit 8 was located in soil having 75% pebbles throughout most of the profile. Fragmented rocks were also found in all other soil pits, but did not hinder sampling.

All soil pits on the study slope have similar structural characteristics. The A horizons have predominately weak gravelly granular structure. The B horizons generally have a weak subangular blocky structure grading downward to the massive structure of the C horizons. Lithographic discontinuities found in several of the soil pits were most probably caused by a residual soil being overlain by colluvial material.

All soil pits with the exception of soil pit 6 were found to have numerous interstitial and tubular pores throughout the soil profile. The larger pores became less frequent with depth, with very few in the C horizons. Soil pit 6 had very few of the larger pores below 15 cm.

Roots were also abundant in all soil pits. The rooting zone was within the first meter of soil, but roots were found to penetrate the highly weathered C horizon to depths of 2 meters or more. Animal burrowings and old root channels, although not numerous, were observed in several of the soil pits.

Particle Size Distribution

The percentages of sand, silt, and clay for each sampling depth in the watershed 10 soil pits that were sampled are given in Table 1. There was relatively little change in the particle size distribution with depth or between soil pits. Generally the percentage of sand was reduced with depth with a subsequent increase in the finer fractions. The soil and subsoil samples from the soil pits were all within the finer texture classes. Most surface horizons were clay loams with B horizons generally classified as silty clays or silty clay loams. Subsoils were predominately classified as clays or clay loams. Sharp changes in particle size distribution did occur in some soil pits. These changes were a distinct shift from sand size particles to clay

Table 1. Mean values of bulk density, particle size distribution, total porosity, and saturated hydraulic conductivity.[1]

Soil Pit No.	Depth	Bulk Density	Sand	Silt	Clay	Total Porosity	Conductivity
	cm	gm/cm^3		percent		percent	cm/hr
1	10	.807	22.5	38.8	38.7	60.8	482
	30	.897	29.8	34.1	36.1	63.8	565
	70	1.015	28.5	33.8	37.7	60.3	196
	110	.981	28.8	33.0	38.2	63.1	206
	130	1.080	30.2	31.4	38.4	55.4	19
	150	1.053	34.0	31.3	34.7	57.6	6
	200	1.052	24.7	36.6	38.7	60.0	20
2	10	.840	39.9	29.4	30.7	65.6	444
	30	.909	29.2	34.4	36.4	62.4	458
	70	1.008	23.4	34.3	42.3	61.9	199
	110	.944	17.4	34.4	48.2	63.6	163
	130	1.018	19.7	35.1	45.2	57.9	51
	150	.991	20.8	31.2	48.0	61.0	26
	200	.943	27.2	34.0	38.8	62.1	21
	250	.974	25.6	39.1	35.3	61.4	55
3	10	.788	24.8	29.1	46.1	69.6	883
	30	.885	24.1	32.3	43.6	66.7	893
	70	.948	29.3	31.0	39.7	63.6	661
	110	.981	33.0	32.6	34.4	62.8	618
	160	1.023	4.9	43.7	51.4	59.1	55
5	10	.857	35.0	29.1	35.9	67.2	665
	30	.923	22.1	31.9	46.0	64.1	716
	70	1.050	29.4	29.8	40.8	59.0	137
	110	1.084	9.8	38.1	52.1	57.9	170
	150	1.210	16.7	35.7	47.6	52.5	68
	190	1.189	19.1	38.1	42.8	53.2	73

Table 1. (Continued)

Soil Pit No.	Depth	Bulk Density	Sand	Silt	Clay	Total Porosity	Conductivity
	cm	gm/cm^3		percent		percent	cm/hr
6	10	.722	12.6	38.7	48.7	70.7	706
	30	1.139	14.3	34.4	51.3	54.0	21
	70	1.108	12.0	37.7	50.3	53.5	8
	110	.954	3.4	41.2	54.6	58.0	14
	150	1.084	6.0	38.2	55.8	54.6	4
	200	1.158	9.4	39.6	51.0	52.4	2
7	10	.809	32.4	28.1	39.5	69.1	837
	30	.826	31.6	31.9	36.5	68.1	761
	70	1.008	31.8	33.3	34.9	54.2	60
9	10	.726	41.1	26.8	32.1	63.1	---
	30	.793	30.1	32.9	37.0	65.9	---
10	10	.860	37.6	31.4	31.0	58.1	615
	30	.930	28.5	37.2	34.3	60.1	431
	70	1.074	27.4	39.2	33.4	51.9	85
	110	.975	22.3	30.6	47.1	55.5	150
12	10	.822	21.3	45.6	33.1	65.1	918
	30	.918	17.2	46.1	36.7	63.6	716
	70	1.223	15.5	45.9	38.6	52.3	53
	110	1.188	7.7	44.2	48.1	53.0	19
	150	1.067	5.0	48.7	46.3	57.9	9

[1]Each value represents a mean of six samples.

size particles in the lower depths of soil pits 3, 5, and 6. The positions in the profile of the changes in particle size coincide with the lithographic discontinuities recorded in the soil profile descriptions. Thus some of the variability of particle size distribution with depth may be explained by the discontinuity of the soil profile.

Bulk Density

The mean values of bulk density for each depth in the soil pits on the study slope are given in Table 1. The bulk density of the soil and subsoil samples were relatively low, ranging from .726 to 1.223 gm/cm^3. Generally the surface soil had the lowest bulk density with values increasing with depth through the B horizon and into the subsoil. However, this gradual increase was not continuous in all soil pits.

Soil properties that affect the value of bulk density the most are soil texture and structure. Fine textured, highly aggregated soils have low bulk density values, whereas sandy soils of massive structures have high values. The surface soil of the study slope is of fine texture and has a well aggregated structure. As previously discussed, the soil at most of the soil pits is fine textured throughout the soil profile. The structure of the soil, as given in the soil profile descriptions, did change from a fine granular structure in the surface horizons to a subangular blocky structure in the B horizons, to a massive structure in the subsoils. The structure, therefore, appears to account for the increase in bulk densities with depth more than a change in particle size. Although the bulk densities of the subsoil are higher than the surface soil, the values are still relatively low compared to most soils of massive structure (Hillel, 1971). One explanation for low bulk density values may be a low particle density for that soil series. Studies conducted to determine the particle density of several soil types found on the H. J. Andrews Experimental Forest have found that the Frissell soil has a lower mean particle density than most other soil series (Dyrness, 1973). The mean particle density for most soil series in this area is 2.65. The mean particle density for the Frissell series was found to be 2.45. Thus the low particle density of this soil might explain some of the low bulk densities reported.

Total Porosity

The mean total porosities for each depth sampled in the soil pits on the study slope are given in Table 1. The soil and subsoil samples were very porous. All samples had more than 50% air space by volume. In most soil pits the total pore space changed very little with depth. Where there was a change the surface soils had very high porosities of 65% of more, while the deepest subsurface soils had porosities of 50-55%.

Total porosity, like bulk density, is influenced primarily by soil texture and structure. Surface soils have a highly aggregated granular structure and a fine texture. There were many large pores between peds and many small pores between the silt and clay particles. As a result the surface horizons had very high pore volumes. In the B horizons, where the structure usually changes to subangular blocks, the larger pores between peds become less frequent. The massive structure of the subsoil has even fewer large pores. However, the total porosity does not drop below 50% at any depth, because texture becomes somewhat finer and the number of small pores between soil particles increases. The precise nature of the pore size distribution of the soil and subsoil will be discussed in a later section.

Hydraulic Conductivity

The mean values of the saturated hydraulic conductivity for each depth in the soil pits of watershed 10 are given in Table 1. The conductivities varied much more with depth than any of the other soil parameters previously discussed. The soil samples from the surface horizons had rapid conductivities of 400 to 1000 cm/hr. The conductivity below the first few decimeters decreased rapidly in every soil pit, but the decrease did not always occur between the same sampling depths. Soil pits 1, 2, and 5 all had high conductivities for the 10 cm and 30 cm depths. The conductivity then decreased by more than half at the 70 cm and 110 cm depths to values near 200 cm/hr. At the 130 cm depth in soil pits 1 and 2, and at the 150 cm depth in soil pit 5 the conductivity was again sharply reduced to a relatively low value. Soil pit 3 exhibited a similar reduction of conductivity from the 10 cm and 30 cm depths to the 70 cm and 110 cm depths; however, the reduction was not great. There was a considerable decrease in conductivity in this soil pit between the 110 cm sampling depth and the 150 cm depth. Soil pits 7, 10, and 12 all exhibited the same high conductivities in the 10 cm and 30 cm depths, but the sharp reduction in conductivity occurred at the 70 cm depth in these soil pits. Soil pit 6 was the only one not having a rapid conductivity at the 30 cm depth. Although the 10 cm depth did have a mean conductivity of more than 700 cm/hr, a sharp reduction to 21 cm/hr occurred at the 30 cm sampling depth. The lithographic discontinuity that was recorded in the profile description of this soil pit is the probable cause for the sharp reduction of conductivity at this shallow depth.

From the conductivity data given in Table 1 it is evident that the conductivity of the soil decreased with depth, and that this decrease was often sharply defined. In order to confirm the apparent reduction of mean conductivity between certain sampling depths in the soil profiles statistical analyses were conducted using the complete set of sample conductivities for the depths examined. The statistical analysis used was a two-sample comparison of samples having independent means. For each apparent sharp reduction of mean conductivity the sample conductivities of the shallower sampling depth were compared to the conductivities of the deeper sampling depth. Each two-sample comparison was made using the null hypothesis stating that the population means were equal (Ho: $\mu_1 = \mu_2$). The alternative hypothesis stated that one population mean was larger than the other (Ha: $\mu_1 > \mu_2$). The test statistic used was the t which for this analysis read:

$$t = \frac{\bar{x}_1 - \bar{x}_2}{S_{\bar{d}}^{2}}$$

where $\bar{x}_1$ and $\bar{x}_2$ are the sample means and the quantity $S_{\bar{d}}^{2}$ is the sample variance of the difference between means (Peterson, 1972). The critical region for each test depended on the level of significance, the nature of the alternative hypothesis and whether or not the population variances were equal. The level of significance was 95% and the alternative hypothesis was one-tailed for all analyses. The equality of the population variances was determined for each analysis using the F statistic. The calculations for determining the critical region and for the sample t are given in Appendix D for each comparison. The sample t, the table t, and the conclusions reached for each comparison are given in Table 2.

Table 2. Results of statistical tests comparing mean conductivities between two sampling depths at the 95% level of confidence.

Soil Pit No.	Depths Examined	$\bar{x}_1$	$\bar{x}_2$	Variance	Sample t	Table t	Conclusion
	dm	cm/hr	cm/hr				
1	3-7	565	196	$\sigma_1^2 \neq \sigma_2^2$	3.780	2.008	Ha: $\mu_1 > \mu_2$
	11-13	206	19	$\sigma_1^2 = \sigma_2^2$	10.371	1.812	Ha: $\mu_1 > \mu_2$
2	3-7	458	199	$\sigma_1^2 = \sigma_2^2$	3.025	1.812	Ha: $\mu_1 > \mu_2$
	11-13	163	51	$\sigma_1^2 \neq \sigma_2^2$	1.992	1.967	Ha: $\mu_1 > \mu_2$
3	3-7	893	661	$\sigma_1^2 \neq \sigma_2^2$	3.382	2.015	Ha: $\mu_1 > \mu_2$
	11-16	618	55	$\sigma_1^2 = \sigma_2^2$	19.140	1.812	Ha: $\mu_1 > \mu_2$
5	3-7	715	137	$\sigma_1^2 = \sigma_2^2$	11.010	1.796	Ha: $\mu_1 > \mu_2$
	11-15	170	68	$\sigma_1^2 = \sigma_2^2$	1.397	1.860	Ha: $\mu_1 = \mu_2$
6	1-3	706	21	$\sigma_1^2 \neq \sigma_2^2$	9.900	2.129	Ha: $\mu_1 > \mu_2$
	11-15	14	3.8	$\sigma_1^2 = \sigma_2^2$	2.680	1.860	Ha: $\mu_1 > \mu_2$
7	3-7	761	60	$\sigma_1^2 = \sigma_2^2$	13.110	1.943	Ha: $\mu_1 > \mu_2$
10	3-7	431	85	$\sigma_1^2 \neq \sigma_2^2$	6.210	2.015	Ha: $\mu_1 > \mu_2$
12	3-7	716	53	$\sigma_1^2 \neq \sigma_2^2$	4.810	2.895	Ha: $\mu_1 > \mu_2$
	11-15	19	9	$\sigma_1^2 = \sigma_2^2$	1.004	1.812	Ho: $\mu_1 = \mu_2$

The results of the analyses show that the conductivities of the shallower depths were significantly greater than the conductivities at the deeper depths for most of the comparisons that were made. The null hypothesis was not rejected for the 110-150 cm comparison in soil pit 5, and for the 110-150 cm comparison in soil pit 12. However, significant differences were found between conductivities of shallower depths in both of the soil pits. Thus all soil pits had a significant decrease in conductivity with depth between sampling levels. Most importantly, however, nearly all soil pits had very low conductivities at some level in the soil profile. Often the conductivity changed abruptly from a high rate to a low rate. Thus in soil pit 1 there was a change from 206 cm/hr at the 110 cm depth to only 19 cm/hr at the 130 cm depth. In soil pit 6 the change was from 706 cm/hr to 21 cm/hr between the 10 cm and 30 cm depths, and in soil pit 12 the sharp decrease in conductivity occurred between the 30 cm and 70 cm depths.

Pore Size Distribution

As discussed previously in the literature review the magnitude of the saturated hydraulic conductivity is determined primarily by the nature of the soil pores. The total porosity for the soil profiles was found to decrease gradually with depth; sharp decreases from high values to low values did not occur. Thus the change in total porosity does not completely explain the variability of the hydraulic conductivity. The change in the distribution of pore sizes offers a more complete explanation for the changes. Data for the pore size distribution is given in Table 3.

The size of the soil pores was found to be distributed unequally in that the majority of the pores were either less than .029 mm in diameter or larger than .294 mm in diameter. The remaining small portion of the total porosity was divided into the pore size classes in between. The total percentages of pores ranging from .029 mm to .294 mm in diameter remained nearly constant as the depth of the samples increased. However, the percentages of the small pores and very large pores did change considerably with depth as illustrated by Figure 5. In this figure the pore size classes between .029 mm and .294 mm have been lumped together so that there are only three pore size classes shown. Using the values from soil pit 1 as an example Figure 5 illustrates how the number of very large pores decreases with depth. A similar pattern of change in pore size distribution with depth occurs in several of the other soil pits.

An important feature that Figure 5 illustrates is that the pattern of change for the large size pores with depth is similar to the change in hydraulic conductivity with depth. The surface samples have a high proportion of large pores. At the 70 cm and 110 cm depths the percentage of large pores decreases sharply, and at the lower

Table 3. Mean values of pore size distribution as fractions of total porosity[1].

Soil Pit No.	Depth	<.029	.049	.073	.098	.147	.294	>.294
1	10	.460	.033	.011	.014	.026	.073	.384
	30	.472	.027	.009	.011	.020	.055	.407
	70	.652	.017	.015	.019	.021	.043	.234
	110	.693	.016	.016	.008	.015	.036	.214
	130	.806	.022	.018	.010	.012	.018	.113
	150	.832	.021	.016	.017	.012	.024	.079
	200	.767	.024	.021	.015	.011	.034	.128
2	10	.490	.022	.020	.014	.025	.052	.378
	30	.519	.018	.019	.014	.022	.050	.358
	70	.633	.016	.015	.017	.022	.052	.244
	110	.674	.017	.016	.019	.019	.040	.217
	130	.783	.019	.015	.008	.011	.025	.138
	150	.750	.014	.014	.016	.010	.022	.174
	200	.732	.021	.019	.019	.016	.054	.139
	250	.762	.024	.011	.008	.018	.016	.161
3	10	.383	.015	.013	.010	.016	.033	.531
	30	.430	.014	.013	.012	.017	.031	.482
	70	.455	.009	.011	.014	.021	.046	.442
	110	.473	.011	.012	.014	.024	.051	.414
	160	.768	.014	.007	.023	.016	.018	.154
5	10	.394	.022	.008	.008	.015	.032	.521
	30	.459	.021	.010	.009	.016	.035	.450
	70	.603	.018	.021	.009	.018	.041	.289
	110	.625	.013	.022	.008	.018	.036	.279
	150	.741	.012	.016	.010	.014	.026	.181
	190	.765	.026	.013	.006	.016	.016	.158

Table 3. (Continued)

Soil Pit No.		Diameter of Pores (mm)						
	Depth	$<$.029	.049	.073	.098	.147	.294	$>$.294
	cm							
6	10	.437	.028	.020	.020	.034	.066	.395
	30	.725	.023	.027	.013	.021	.034	.156
	70	.749	.029	.022	.018	.020	.027	.135
	110	.646	.005	.061	.048	.041	.062	.138
	150	.752	.001	.058	.032	.026	.032	.099
	200	.788	.003	.052	.028	.030	.033	.066
7	10	.406	.010	.008	.011	.016	.035	.513
	30	.404	.013	.006	.015	.017	.044	.501
	70	.588	.002	.062	.033	.033	.056	.225
9	10	.400	.022	.012	.018	.026	.059	.462
	30	.375	.019	.010	.019	.023	.058	.495
10	10	.411	.044	.011	.014	.043	.048	.429
	30	.433	.041	.012	.013	.049	.062	.390
	70	.628	.055	.012	.012	.046	.054	.193
	110	.630	.039	.009	.009	.031	.039	.245
12	10	.430	.013	.009	.009	.013	.032	.496
	30	.481	.017	.011	.011	.014	.037	.430
	70	.781	.018	.009	.008	.012	.026	.146
	110	.807	.016	.009	.007	.011	.021	.128
	150	.827	.010	.024	.004	.011	.017	.106

[1]Each value represents the mean of six samples.

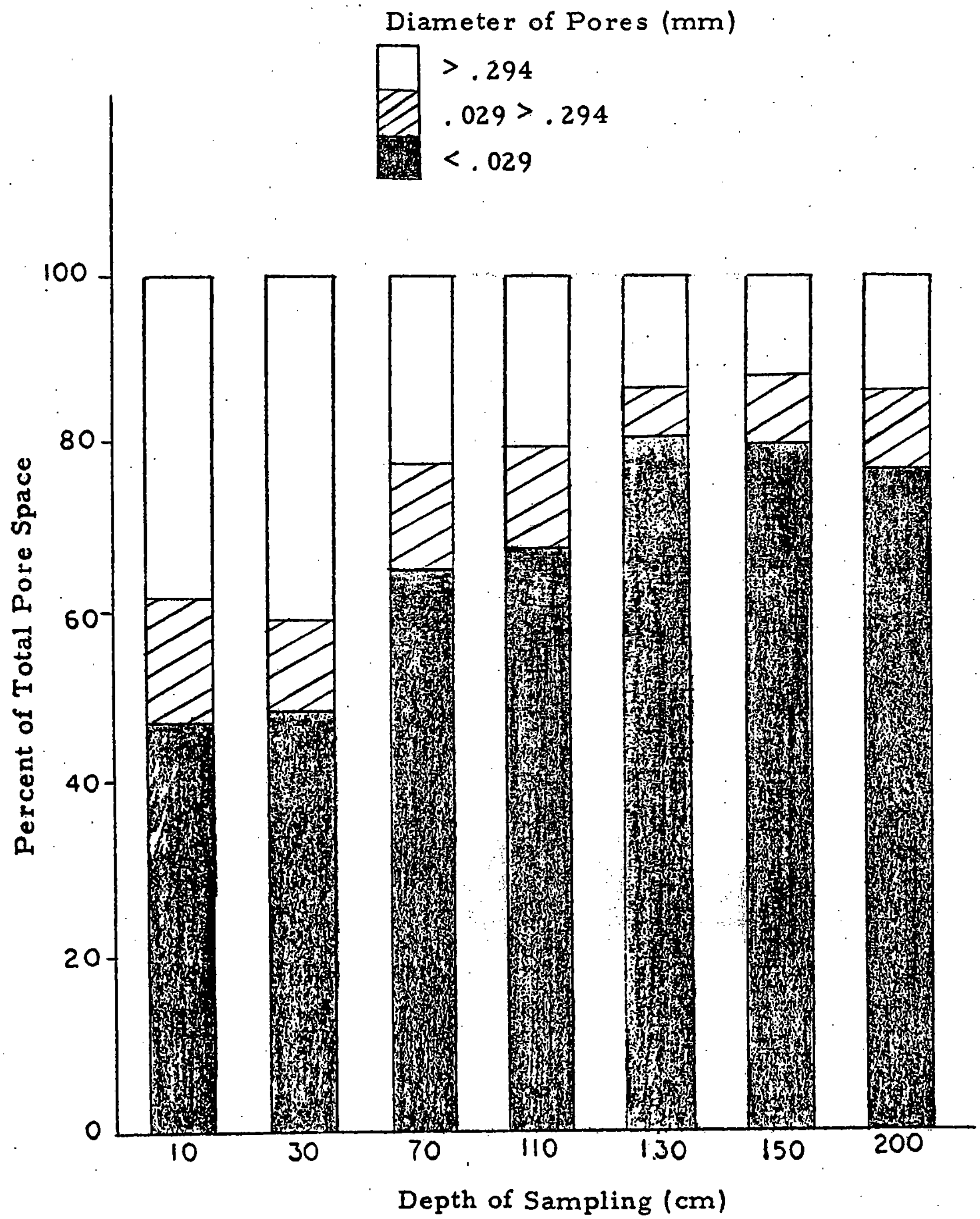

Figure 5. Change in pore size distribution with depth for soil pit 1.

horizons there is another sharp reduction. This pattern of change with depth suggests that there is a close correlation between large size pores and hydraulic conductivity. The relationship between conductivity and pore size distribution for all soil pits is illustrated in Figure 6. The mean hydraulic conductivity and mean percentage of pores greater than .294 mm in diameter are plotted against each other on a log-log scale. A linear regression, a power curve regression and an exponential curve regression were performed on the full set of mean conductivities and large pore size data. The power curve equation was found to represent the relationship more closely than the ~~others~~ and had an r^2 of .945. The equation for the line reads:

$$\hat{Y} = 10,040X^{2.997}$$

where $\hat{Y}$ is the value of hydraulic conductivity to be estimated and X is the value of the percentage of pores greater than .294 mm in diameter expressed as a decimal.

Because there is a strong correlation between conductivity and the percentage of large pores, the soil properties affecting the conductivity to the greatest extent must be those that affect the percentage of large size pores. As previously discussed these parameters are the soil texture and soil structure. The soil texture did not change considerably with depth and thus did not affect the change in the percentage of large size pores nor the hydraulic conductivity as much as the change in soil structure. Thus the change in soil structure

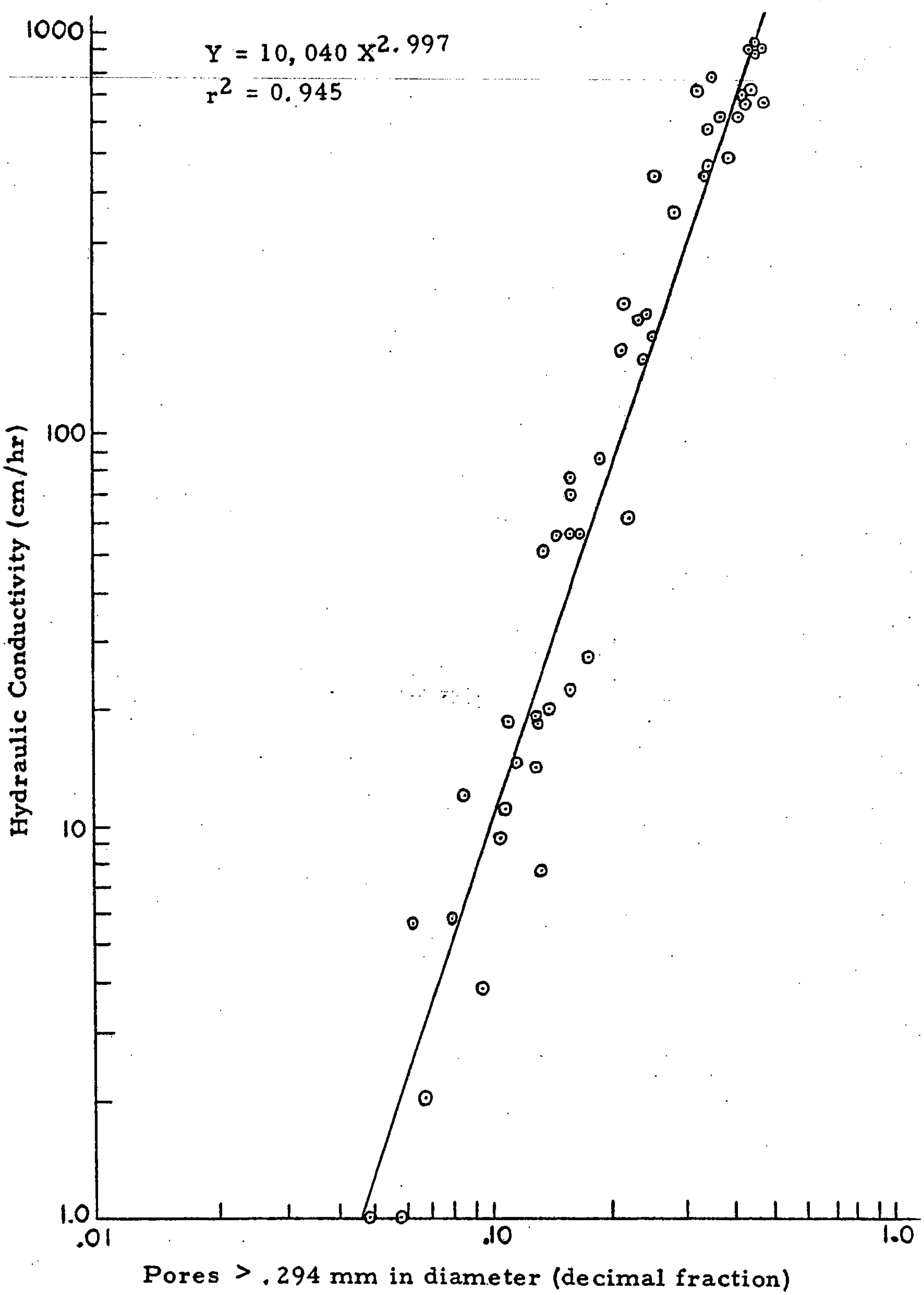

Figure 6. Hydraulic conductivity vs pores > .294 mm in diameter.

with depth apparently has the most influence on the hydrologic properties of the soil and subsoil. The changes in structure is primarily the result of differential weathering with depth. The soil profile was developed by the combined effects of root penetration, animal activity, organic matter, colluvial action, and chemical and physical weathering. The subsoil was developed primarily by the residual weathering of red breccia rock which resulted in a single-grained massive structure.

Moisture Characteristic

The moisture characteristics expressed as water content in percent by volume for each tension applied to the sample on the tension table, are given in Table II in Appendix B. All values are mean water contents for the soil pit and sampling depth indicated. Table II shows the percentage of the total sample volume consisting of water after each sample was held at the indicated tension for 48 hours.

A graphic representation of the change in water content with tension is known as a soil moisture characteristic curve. The characteristic curves for the sampling depths of soil pit 1 are given in Figure 7.

Figure 7 shows that the samples all have nearly the same percentage of water when saturated, but vary considerably in water content at the 100 cm tension. The difference arises at the 10 cm

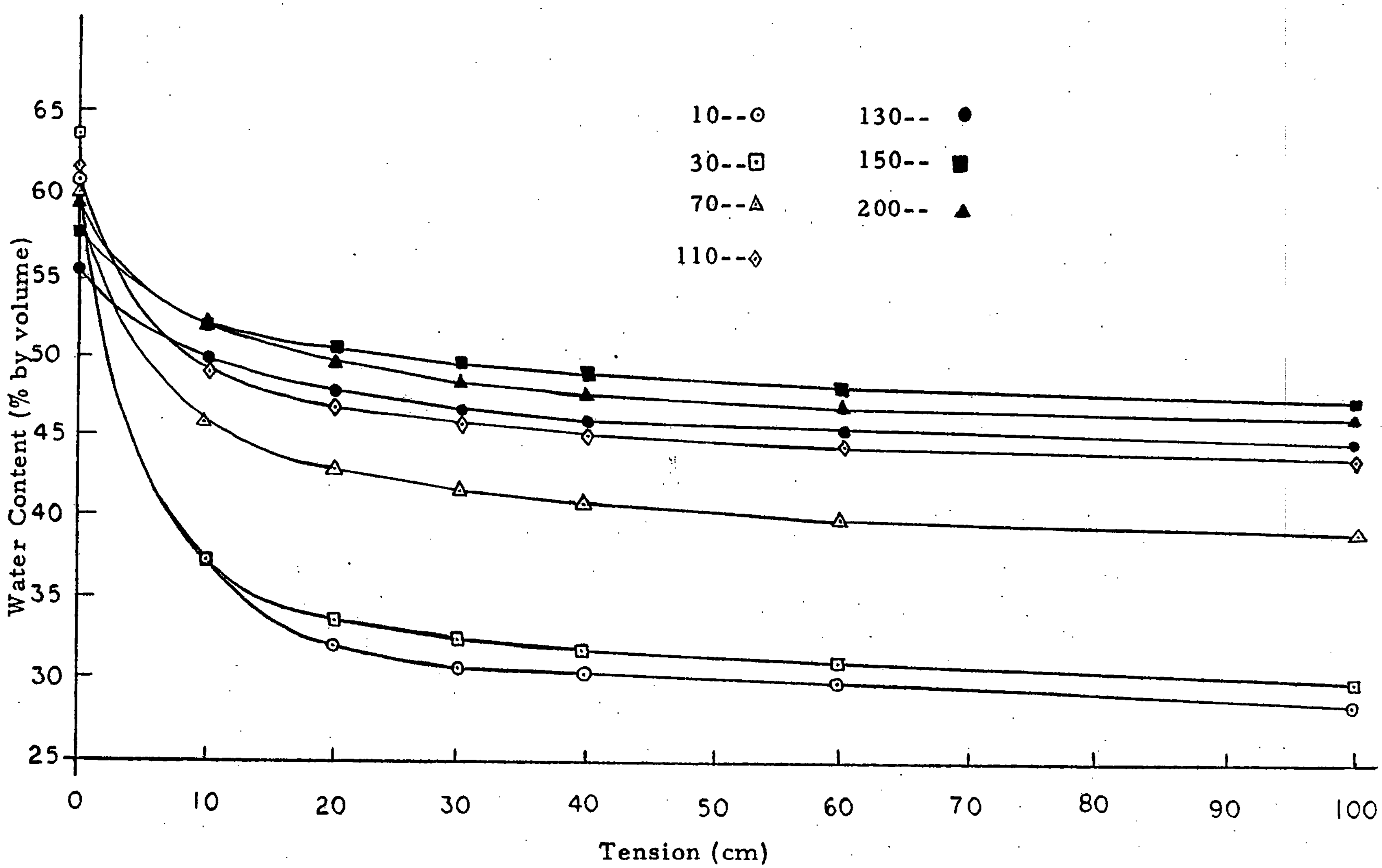

Figure 7. Soil moisture characteristic curves for soil pit 1.

tension where the largest proportion of water is lost in each of the sampling depths.

By examining the curves between saturation and the 10 cm tension, the influence of sample depth on water content can be better understood. The samples from the 10 cm and 30 cm depths lose more water than any other samples at the 10 cm tension, from approximately 62% to 37%, a difference of 25% or 40% of the water held at saturation. The next lower depths 70 cm and 110 cm lose less, from approximately 62% to 48%, a loss of 23% of the water held at saturation. The lower depths, 130 cm, 150 cm, and 200 cm, lose the least, from near 58% to 52%, a loss of only 10% of the water held at saturation. Thus for soil pit 1 the soil moisture-tension relationship changes with depth in a manner similar to the hydraulic conductivity, i.e., in steps or stages between the 30 cm and 70 cm depths, and the 110 cm and 130 cm depths.

The amount of water retained in the soil at each tension is dependent on the size of pores within the soil. Smaller pores have greater capillary potentials and can thus retain more water against tensions applied to the sample than larger pores. As previously discussed the pore size distribution with depth changes from many large pores in the surface soil to mostly small pores in the subsoil. Thus the subsoil samples retained more water than the surface soils when the tensions were applied. The shape of the soil moisture characteristic curves for each depth must therefore also depend more on the change in soil structure than the change in soil texture because the change in structure was found to control the percentage of large pores found at each sampling depth.

NATURE OF SOIL WATER MOVEMENT

The hydrologic properties of soil and subsoil can help explain the nature of subsurface movement of water. As previously discussed, streams of this area respond rapidly to precipitation. Maximum rates of runoff are as high as 80% of the average rate of precipitation for the preceding 12 to 24 hours (Rothacher et al., 1967). Considering that almost no surface runoff occurs in this area it must be concluded that some form of net subsurface movement of water is responsible for the rapid rise of storm hydrographs. Interpreting the hydrologic properties of the soil and subsoil on the study slope of watershed 10 in terms of soil physics will help provide a possible explanation for the nature of this subsurface movement.

Antecedent Moisture Conditions

Moisture conditions within the soil will be determined by the nature of the precipitation being added to the soil and the soil's pore size distribution. During the winter rainy season storm events typically occur every 2-4 days and are usually of low intensity and long duration. Rates of 6 mm/hr for 8-10 hours are not uncommon. The extremely porous, well aggregated soil found in the surface horizons of all soil pits has very high infiltration capacities. Table 3 shows that the soil also contains a large percentage of small diameter

(< .029 mm) pores. At the beginning of the rainy season the small pore spaces throughout the soil profile can be expected to become filled. Figure 7 shows that if the smallest pores (represented by the 100 cm tension) remained filled in the soil at soil pit 1, the moisture content would be as high as 40%, or 66% of saturation for the 70 cm depth. Because of the frequency and duration of the winter storm events the soil on the study slope will remain at a high moisture content for long periods of time, probably from early December until late March. Thus during the winter rainy season the soil will have a considerable retention storage of water, more than 28 cm in the surface 70 cm of soil.

The subsoil on the study slope consists primarily of massive, highly weathered breccia having pore size characteristics dissimilar to the shallower horizons. The subsoil pore space consists primarily of pores less than .029 mm in diameter (see Table 3 and Figure 5). These pores, like the small pores in the soil, would remain filled with water throughout the winter season. The moisture content of the subsoil (Figure 7) remains near 47% or 84% of saturation at the 100 cm tension for the 130 cm depth. The subsoil would remain much closer to saturation than the surface soil when the soil water tensions in the soil and subsoil are the same. I therefore conclude that the subsoil would be the most probable level to become saturated as additional water percolates down from the surface and water moves more-or-less downslope laterally from the subsoil upslope.

The position of the zone of saturation within the soil profile will vary with the hydrologic properties of the subsoil. The position of the zone of saturation in the subsoil near soil pit 1 should be near the 130 cm depth, the level where the pore size distribution abruptly changes to primarily pores less than .029 mm in diameter. The vertical extent of the saturated zone would be limited by the rate of water movement in the soil and by the rate of rainfall entering the soil. I would not expect the saturated zone to extend much further into the subsoil due to the very low conductivity rates in this zone. Although downward movement of water would continue, the saturated zone would most likely extend more rapidly upward as long as additional rainfall was being added to the soil. However, with the intensity of rainfall as low as it normally is, the zone of saturation probably would not extend more than a few centimeters upward into soil having a greater percentage of large diameter (> .294 mm) pores.

Unsaturated Flow Through the Soil

Because of the nature of rainfall and the physical properties of the soils on the study slope the soil moisture content will remain high and the soil water tension will remain low throughout the winter rainy season. The soil will not become saturated because of the sizeable volume of the total pore space in large diameter (.294 mm) pores and the low rates of precipitation. Therefore the primary mechanism of water movement through the soil is unsaturated flow.

The equation for flow through an unsaturated soil is a form of Darcy's law modified by the unsaturated hydraulic conductivity. The equation reads:

$$q = -K(\theta)\,\Delta H$$

where q is the flux, $K(\theta)$ is the unsaturated conductivity, a function of the soil moisture content, and $-\Delta H$ is the hydraulic gradient, which includes both pressure (tension) and gravitational components (Hillel, 1971).

The determination of unsaturated hydraulic conductivity of the soils and subsoils on the study slope was outside the scope of this thesis. Therefore, the values of the flux for varying moisture contents in the soil were not calculated for use in this thesis. Although the rate of movement was not determined, the direction of water movement is probably in the vertical direction toward the subsoil during rainfall events. The surface soil would increase in soil moisture and decrease in soil water tension after rain water began to enter. The vertical hydraulic gradient would increase between the surface soil and the soil at lower depths having higher tensions. Thus the tension as well as the gravitational component of the hydraulic gradient will cause a net downward movement of water. After rainfall has ceased the soil water tension in the surface soil and in the deeper soil would begin to approach equilibrium and the rate of downward movement of water would decrease. The next rainfall event would begin the cycle of water movement over again.

Saturated Flow in the Subsoil

It is within the potential saturated zone directly above the relatively impervious subsoil that the greatest probability for saturated lateral movement of water exists. With vertical movement impaired, the water in this zone would respond to the combined gradients of the gravitational potential and the pressure potential of the water in the saturated zone upslope. The result would be a net movement of water in a more lateral direction than in the soil zone above.

I do not envision the zone of saturation within the subsoil to be more than a few centimeters in depth. The flow of water both above and below this saturated zone will occur as unsaturated flow. Thus most of the water movement on the study slope is in the form of unsaturated flow.

Translatory Flow

With unsaturated flow predominant within the soil the question arises as to how the stream is able to respond so swiftly to the onset of precipitation. The answer is a net movement of water through the soil to the stream in the form of a displacement of antecedent water within the soil. This net water movement has been termed translatory flow by Hewlett and Hibbert (1967). With proper soil moisture conditions, water entering the soil is capable of sending a pulse through the soil water which displaces a similar volume of antecedent water at a point some distance away. Thus the water entering the stream during or just after a rainfall event need not be the precipitation from that event. Such displacement requires the soil to remain at low tensions and high moisture contents between rainfall events.

Because the soils and subsoils of the study slope do remain at a high moisture content the conditions for translatory flow exist. As water from a rainfall event infiltrated into the soil the moisture content would be increased further. Because the soil would initially be at a high moisture content the entire profile would respond to the rainfall event. The zone of saturation above the nearly impervious subsoil layer would form and translatory flow would occur through the unsaturated soil from the surface to the zone of saturation. Translatory flow would also occur through the zone of saturation, downslope toward the stream. Thus translatory flow would provide for a displacement of water that would be much faster than water could actually flow through the soil, saturated or unsaturated.

Distance from the stream will also determine how fast the rainfall on a particular area will contribute to storm runoff. The soils and subsoils nearest the stream would be expected to retain a high moisture content for a greater period of time due to the volume and duration of the subsurface flow passing down from the slope above. These soils and subsoils would then be expected to respond most quickly to the onset of rainfall events. Soils and subsoils of the mid-slope would contribute to the runoff more slowly and the upper portion of the slope may not even contribute to the storm runoff despite the translatory flow process. This type of response to rainfall events is the variable source area concept described by Hewlett and Hibbert (1967). Thus translatory flow coupled with the variable source area concept provides a feasible explanation for the quick response of stream hydrographs to the onset of rainfall events in this area.

The extreme variability of soil and subsoil hydrologic properties indicates that the subsurface flow process is perhaps more complicated. Water has been observed to enter the stream of watershed 10 at the base of the study slope by way of soil piping suggesting some form of channelization of the soil water before it reaches the stream. Undoubtedly, sheet flow over bedrock occurs in some areas. The investigation of such mechanisms within the soil-subsoil system of the study slope was also outside of the scope of this study.

SUPPORTING EVIDENCE OF PREDICTED SUBSURFACE FLOW

The preceding discussion used the measured hydrologic properties of the soils and subsoils to make a reasonable prediction of the nature of water movement and its response to winter rainfall events. To help verify these predictions concerning subsurface water movement further evidence was required. First, samples were taken from a soil pit where lateral movement of water had been observed to occur over the surface of a subsoil horizon. The hydrologic properties of this subsoil were compared to the hydrologic properties of the study slope subsoils to determine if the properties of the subsoils of the study slope were conducive to the same type of subsurface flow. To verify the predictions made for the nature of water movement through the entire soil profile, data from tensiometers installed on the study slope were used. The tensiometers installed near several of the soil pits provided soil moisture data during and between rainfall events. The data were used to help verify the predictions made concerning changing soil moisture conditions and subsequent water movement throughout the soil profile.

Soil Pit 11 Hydrologic Data

A mass movement of soil occurred above a road cut approximately one mile from the watershed 10 study slope. Following the event water was observed flowing out of the escarpment and over the surface of a saprolite subsoil horizon that had been exposed. Evidently the hydrologic properties of the soil and subsoil were conducive to saturated lateral movement of water over the top of a subsoil composed of the same saprolite material found within the study slope subsoils. The subsoil of the escarpment was sampled and the hydrologic properties were determined. The hydrologic properties of the subsoil at this location (labeled soil pit 11), were then compared to the properties of the soil pits on the study slope to determine if similarities existed. If so, then by way of logical inference the subsoil of the study slope could be considered capable of saturated lateral subsurface flow similar to the type of flow occurring in soil pit 11.

The profile description for soil pit 11 is given in Appendix C. The soil was found to be similar to the Frissell soil found on the study slope. The profile description resembles the profile descriptions for the soil pits on the upper portion of the study slope where the subsoils are deepest. The textures of soil pit 11 are also fine in the A and B horizons and very fine in the subsoil. The structure grades from a fine granular to a subangular blocky to massive in soil pit 11 as it does in the study slope soil pits. One notable difference is the stone content. Soil pit 11 had less unweathered rock in all horizons than most of the study slope soil pits. The subsoil of soil pit 11 also had fewer pores and roots. The subsoil was found to be composed almost entirely of highly weathered breccia with very few unweathered stones. Additional differences were discovered when examining the hydrologic properties.

Table 4 gives a listing of the subsoil hydrologic properties for soil pit 11. Because the properties of the subsoil were considered most important, more depths were sampled in this material and no surface soil samples were taken. By comparing the properties in soil pit 11 to the properties of similar depths in the study slope soil pits (Tables 1 and 3) some differences and similarities can be noted.

Bulk densities of the subsoil in soil pit 11 are greater than most bulk densities of the study slope subsoils. Very few of the subsoils on watershed 10 have densities greater than 1.200 gm/cm^3 and none have values consistently above 1.200. A greater particle density for the soil in this soil pit could provide a partial explanation for this difference. The particle size distribution appears to be similar in both areas with very fine textures in the lower depths. Total porosity figures for soil pit 11 are also approximately the same as or slightly lower than the study slope porosities.

The hydraulic conductivities of the depths sampled in soil pit 11 are all very low. The conductivities of the 180 and 200 cm depths were less than 1 cm/hr. The 180 cm level was the approximate level where lateral water movement had been observed. The low conductivities measured for all subsurface sampling levels appears to be the

Table 4. Mean values of hydrologic properties of soil pit 11.[1]

Depth	Density	Sand	Silt	Clay	Total Porosity	Conductivity	Pore Size Distribution <.029mm	.029mm $\leq$ .294mm	>.294mm
cm	gm/cm^3		percent		percent	cm/hr		decimal fraction	
90	1.184	20.6	41.8	37.6	54.2	11	.831	.061	.108
110	1.292	7.1	48.8	44.1	49.8	15	.843	.058	.099
130	1.273	---	---	---	51.4	12	.856	.059	.085
150	1.276	16.3	34.6	49.1	50.7	6	.884	.053	.063
180	1.176	15.4	31.2	53.4	54.9	<1	.901	.042	.057
200	1.173	---	---	---	55.1	<1	.913	.039	.048

[1]Each value represents the mean of six samples.

result of the pore size distribution. The fraction of pores greater than .294 mm in diameter is very small in soil pit 11, smaller than most sampling depths in the subsoils of the study slope soil pits. At the 180 and 200 cm depths only 5% of all pores are greater than .294 mm in diameter. With such a disproportionate distribution of pores sizes and the subsequent low conductivities vertical movement of water through this subsoil would be extremely slow to nonexistent. Conductivities of the subsoil in the soil pits on the study slope often were less than 20 cm/hr but were always found to be greater than 1 cm/hr. In addition these subsoils were overlain by soils having a much higher rate of water movement. The depth of the sharp break in conductivities and pore size distribution had been theorized to be the depth where saturation and lateral water movement would most likely occur. Because there appears to be very little difference in the conductivities of the subsoil directly above and below the failure plane in soil pit 11 it is not possible to conclude that water will move laterally in the same manner in the study slope subsoils. However, this does not preclude the possibility of lateral movement of water in the study slope. Soil pit 11 was in the only area where lateral water movement over the subsoil had been observed and soil and subsoil could be sampled. Due to limitations of sampling the results of a comparison between the hydrologic properties of this area and of the study slope were considered inconclusive in terms of subsurface flow.

Tensiometer Data

More substantial evidence indicating the nature of water movement through the soil and subsoil on the study slope was provided by measurements of soil water tension. Prior to the 1973-1974 rainy season tensiometers were installed in plots near several soil pits on the study slope as part of a broader investigation of subsurface water movement. A portion of the data collected for that study has been used here to provide information on the changes in soil water tension during two rainstorm events and to evaluate predictions of water movement based on soil characteristics. This data was collected between December 22 and 25, 1973. The tensiometers were read at four hour intervals during storm events and at eight hour intervals between events. Rainfall was also recorded during this four day period.

Tensiometers were installed in soil and subsoil that were considered to have hydrologic properties similar to those of the soil and subsoil sampled in the adjacent soil pits. For this study two tensiometer plots located about 10 m from soil pit 1 and one plot located about 5 m from soil pit 5 were used. Each plot consisted of several tensiometers located at depths corresponding to the sampling depths in the soil pits nearby. Plots 1 and 2, located near soil pit 1, contained six tensiometers which were at depths of 10, 30, 70, 110, 130, and 150 cm. Plot 3 near soil pit 5 contained tensiometers at the 10, 30, 70, and 110 cm depths. Other plots were located near some of the other soil pits, but several of the tensiometers did not function properly and a more complete set of data was not available.

Changes in soil water tension and rainfall were plotted for the four day period (Figures 8, 9, and 10). Tension measurements at each tensiometer depth have been connected with straight lines to aid in following the changes in tension with time. The soil water tension data were analyzed to determine what the soil and subsoil moisture conditions were during and between rainfall events. The data were also used to help describe the nature of water flow through the soil and subsoil.

Antecedent Moisture Conditions

Soil moisture characteristic curves for soil pit 1 (Figure 7) and soil pit 5 were used to prepare plots of the change in soil moisture with time for the three tensiometer plots (Figures 11, 12, and 13). The effect of hysteresis during the wetting and drying of the soil and subsoil from one rainfall event to the next was ignored in the preparation of these plots. Most tensiometers registered low tensions within a small range near saturation (zero tension). Therefore hysteresis is assumed to have induced only a very small error in the determination of soil moisture contents here. The soil water tension and soil moisture content figures show that the soil and subsoil had low tensions

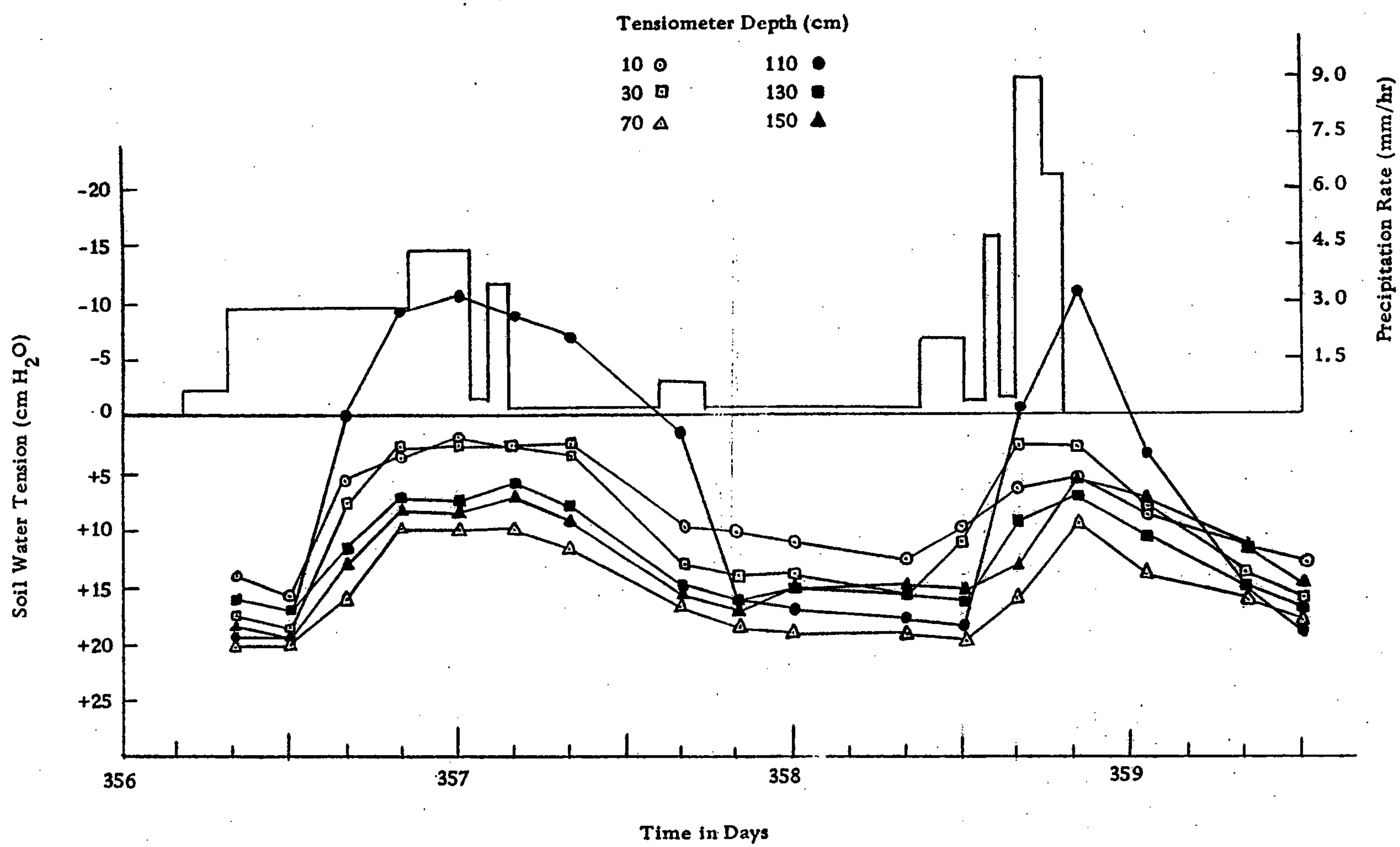

Figure 9. Soil water tension and rainfall intensity data at tensiometer plot 2.

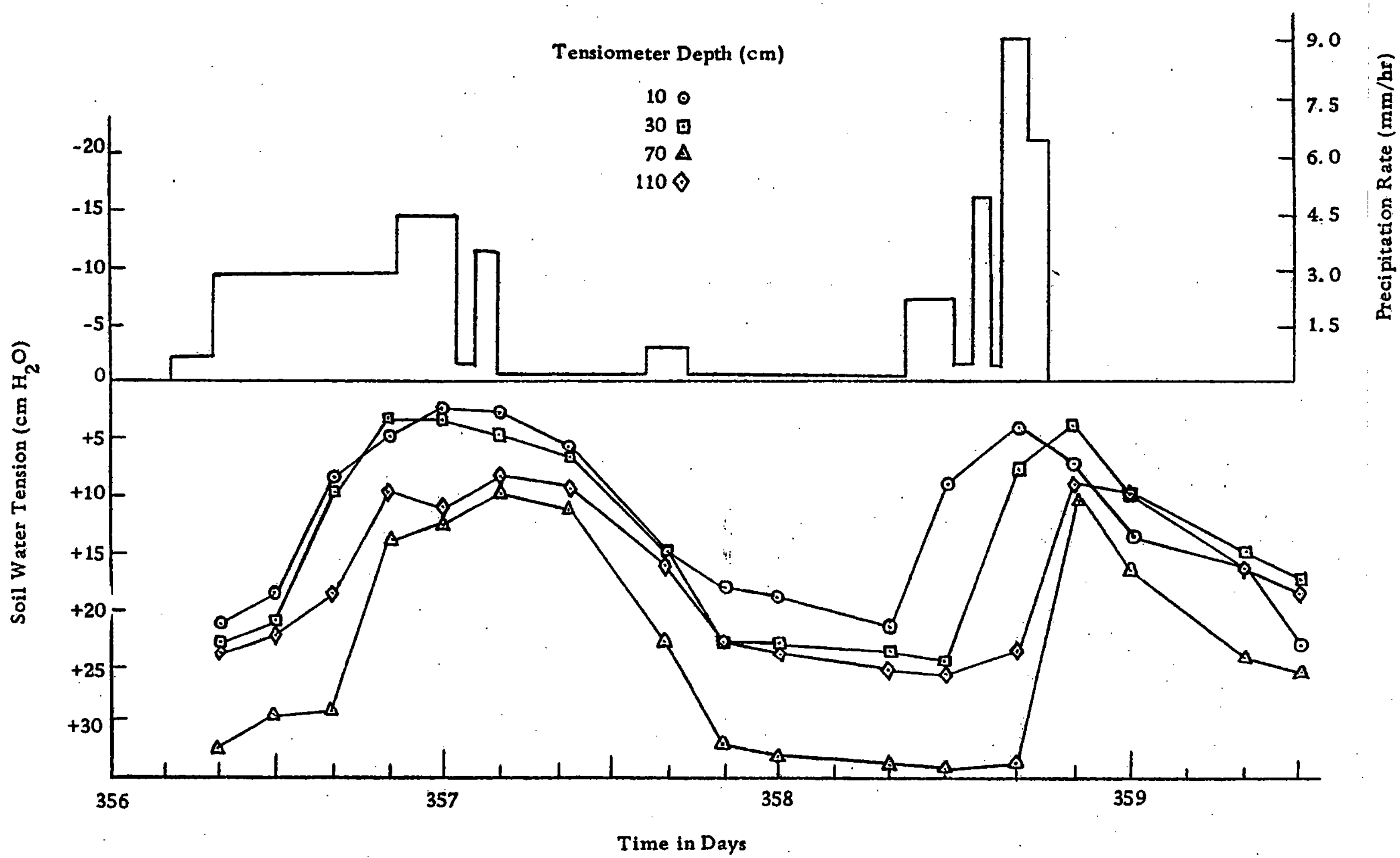

Figure 10. Soil water tension and rainfall intensity data at tensiometer plot 3.

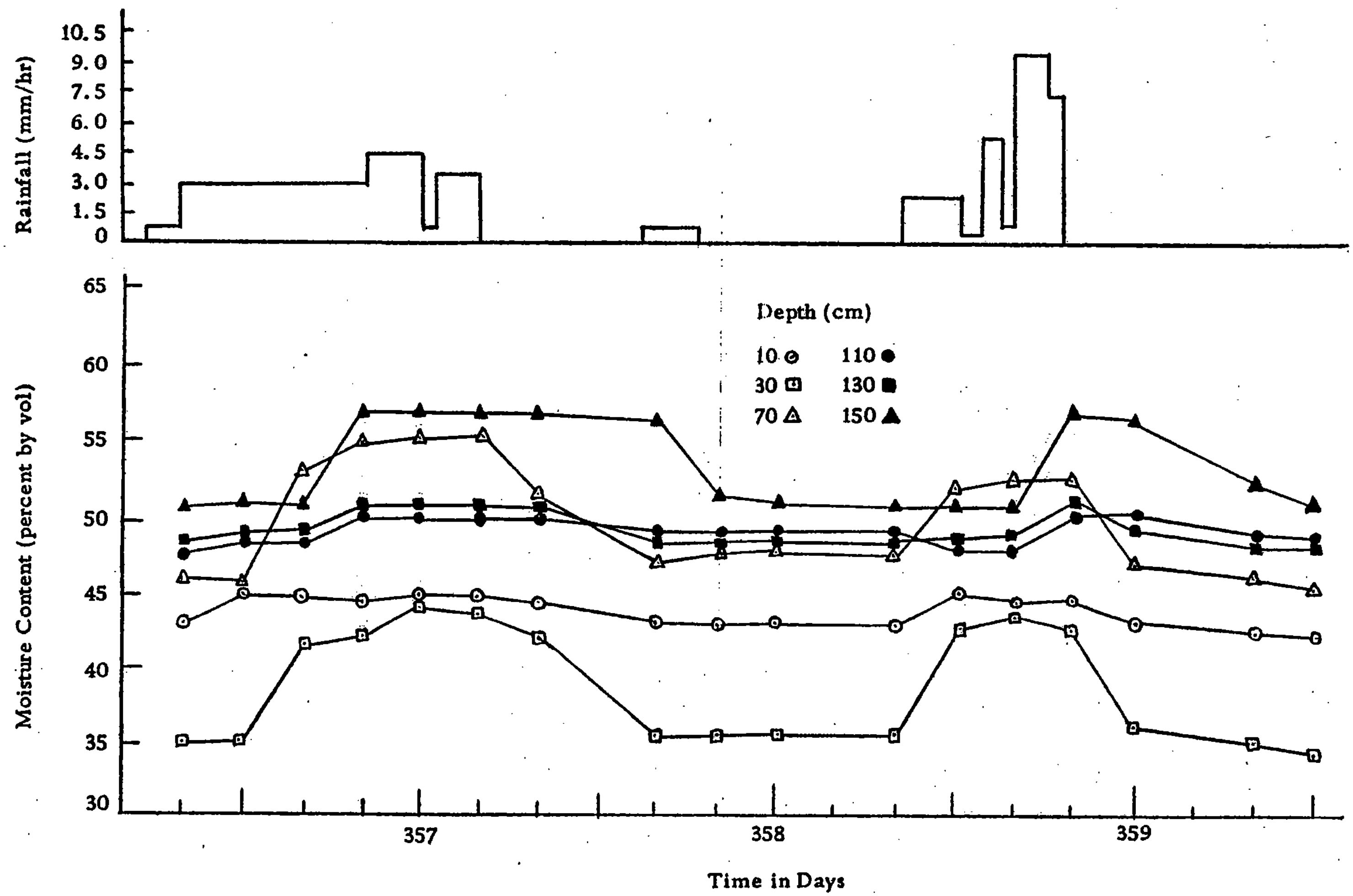

Figure 11. Soil moisture content and rainfall intensity data at tensiometer plot 1.

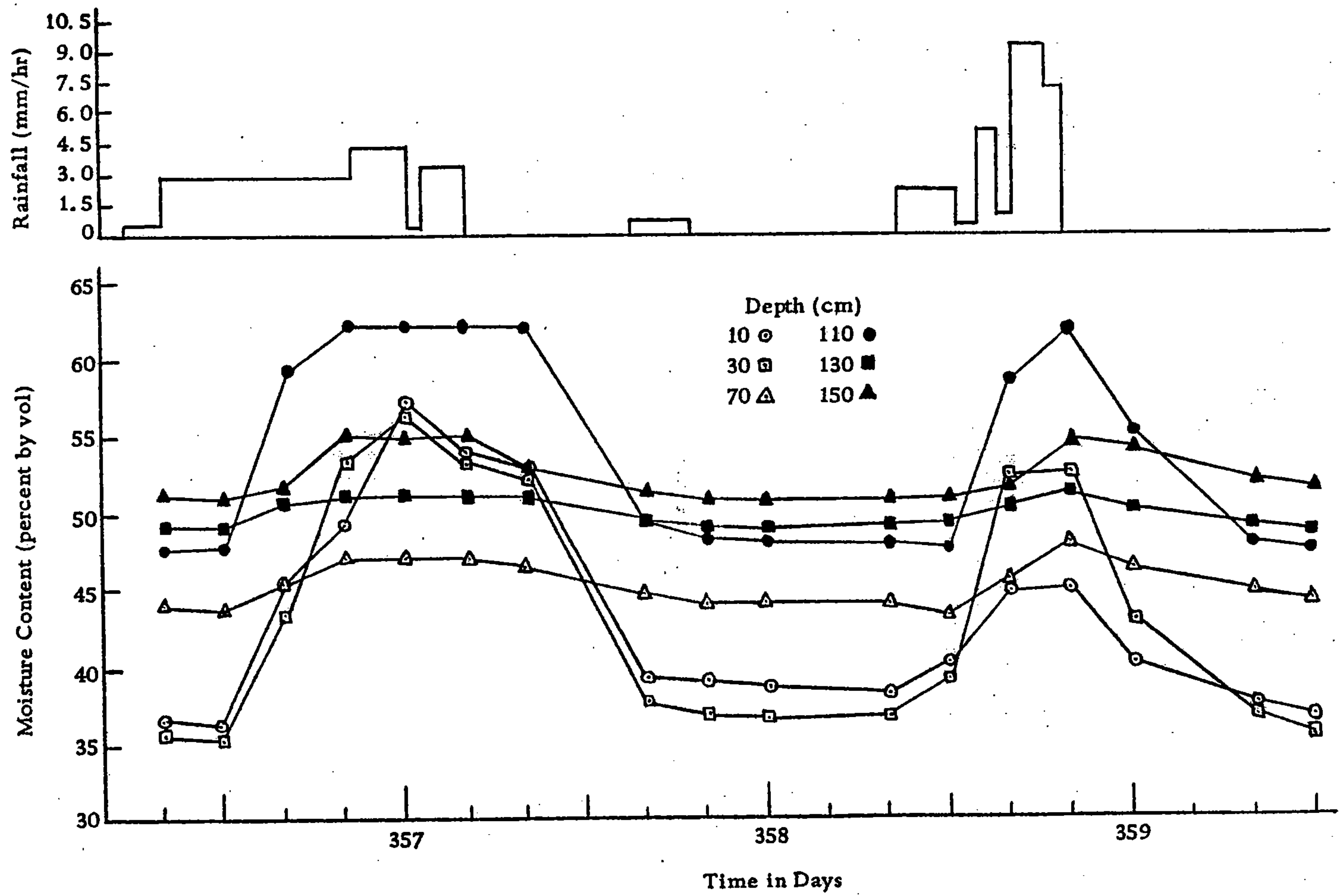

Figure 12. Soil moisture content and rainfall intensity data at tensiometer plot 2.

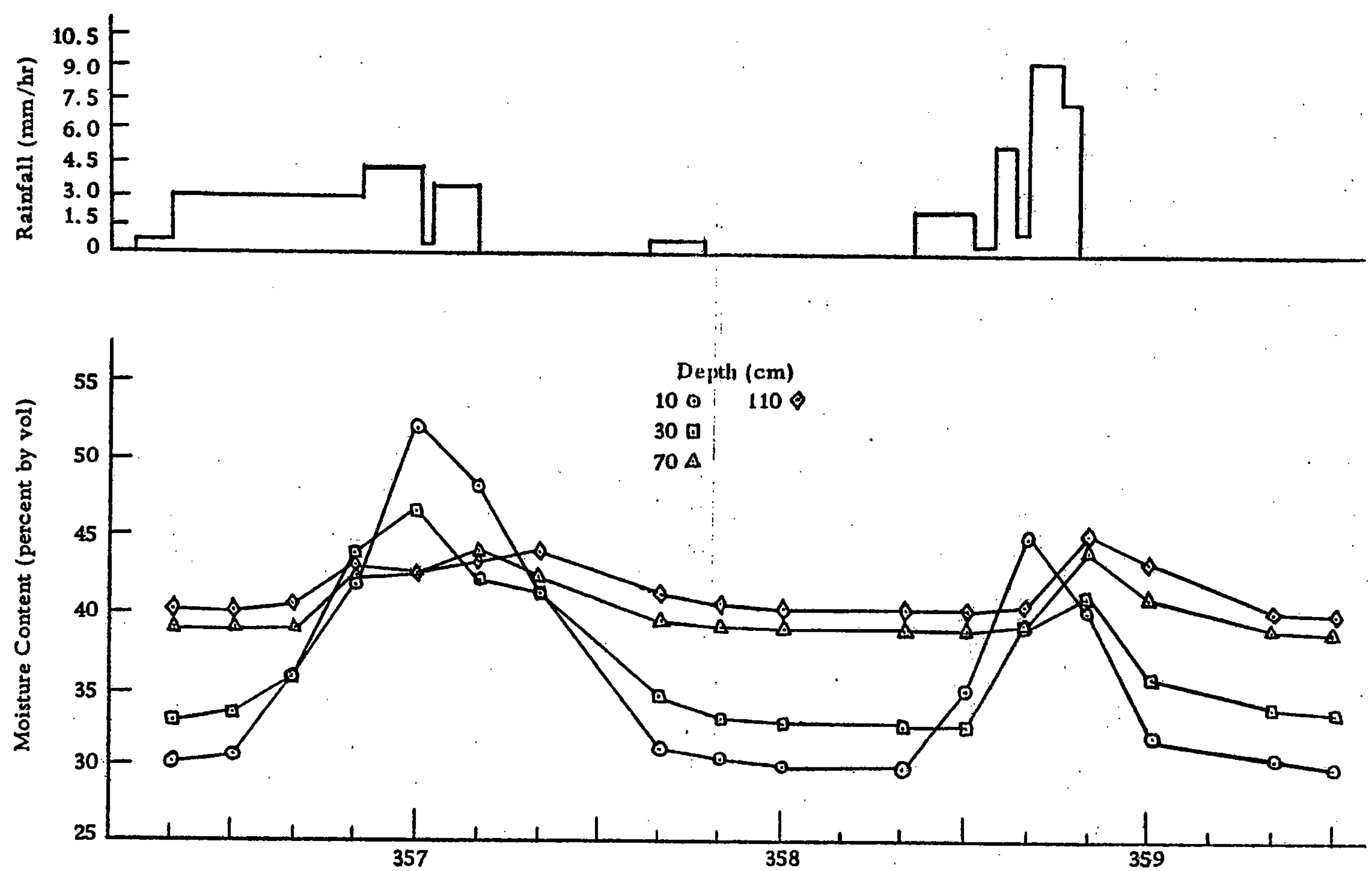

Figure 13. Soil moisture content and rainfall intensity data at tensiometer plot 3.

and high moisture contents throughout the four day period. Figures 8, 9, and 10 show that the soil water in the soil (represented by the 10, 30, and 70 cm tensiometer data) decreased in tension during the rainfall events but never reached zero tension. The soil moisture content figures show that the soil generally fluctuated between 30% and 55% moisture content or between 50% and 90% of saturation from one storm event to the next. All three figures also show that the subsoil had higher moisture contents and remained closer to saturation than the surface soil for most of the four day period as had been previously predicted.

As discussed earlier a zone of saturation was thought probable above the 130 cm depth in the subsoil near soil pit 1. Figures 8 and 9, representing the soil water tension data for the soil near soil pit 1, show that a zone of saturation did form in the subsoil. However, positive pressures (negative tensions) were recorded during both storms at the 150 cm depth in tensiometer plot 1 and at the 110 cm depth in plot 2 rather than at the predicted 130 cm depth. Undoubtedly, local variation in the vertical location of the abrupt change in the pore size properties of the subsoil will account for these differences. The shape of the 110 cm curve in plot 2 and the 150 cm curve in plot 1 are quite similar, indicating that the zone of saturation formed and responded to the rainfall events in a similar manner.

None of the tensiometers in plot 3 near soil pit 5 indicated saturated soil conditions. Tensiometers were not installed at depths greater than 110 cm in this plot so that if a saturation zone did form at a lower level it was not detected.

Nature of Water Flow Through the Soil and Subsoil

Of particular interest on the plots of moisture content (Figures 11, 12, and 13) is the shape of the curves during and between the rainfall events. As the rainfall events began, the moisture content for each depth increased. The increase continued as the rainfall event continued, but ~~then~~ the ~~moisture~~ contents reached a maximum level and stopped increasing. Each depth remained at this nearly constant level until the rainfall event ceased. The moisture contents then dropped to a lower level and remained close to that level until another rainfall event occurred. In terms of water movement this pattern of change in soil moisture content would appear to indicate that the water entering each soil layer increased the moisture content by filling a certain volume of pores before passing through to lower depths. As the rainfall event continued, the water passed through the soil and subsoil without increasing their moisture contents. The mechanism by which this would most probably be accomplished is translatory flow. Once the soil and subsoil have reached their maximum moisture contents the volume of water entering the profile as rainfall could displace the antecedent soil water causing an equal volume of water to enter the saturated zone in the subsoil. This zone of saturation was at the 110 cm level in the subsoil near plot 2 and at the 150 cm level in the subsoil near plot 1.

The tensiometer data for the soil and subsoil near soil pit 1 have provided evidence that have supported predictions of saturated and unsaturated translatory flow through the profile near this soil pit. Although no zone of saturation was found near soil pit 5 the tensiometer data do indicate that the soil moisture conditions near this soil pit were also conducive to translatory flow. Thus information provided by the tensiometer plots has supported the predictions of the nature of subsurface flow through the study slope that were based on the analyses of the hydrologic properties of its soil and subsoil.

CONCLUSIONS

Major conclusions regarding the hydrologic properties of the study slope and subsoil are as follows:

1. Soil and subsoil structural characteristics had a greater influence than the textural characteristics on the hydrologic properties. The surface soils were well aggregated and usually contained more than 30% pebbles and shot-like concretions. Lower soil depths had subangular blocky structure with less aggregation of soil particles. The subsoil on the study slope has a massive structure containing highly weathered breccia rock.

2. The particle size distribution changed very little from the soil to the subsoil. Most soils were classified as clay loams and the subsoils as clays.

3. Bulk density values were low for both the soil and subsoil on the study slope. The surface soil had values averaging near .800 gm/cm^3 while the subsoil bulk densities increased to near 1.100 gm/cm^3. The bulk densities may be this low partially because of the low particle density of the Frissell soil.

4. Total porosity decreased with depth on the study slope but remained high even at the deepest of sampling points. Surface soils had total porosities of 60-70%, while the subsoils had porosities of 50-60%.

5. Although the total volume of pore space did not decrease much with depth the distribution of the pore sizes did change. The changes involved a shift from pores of large diameter (>.294 mm) in the soil to pores of small diameter (< .029 mm) in the subsoil. The changes occurred in most soil pits between the 30 cm and 70 cm depths as well as between the 110 cm and 130 cm depths in some soil pits. The soil structure change with depth caused the change in pore size.

6. Saturated hydraulic conductivity was closely correlated with the percentage of large diameter (> .294 mm) pores within the soil and subsoil. A power curve equation showed the relation to be: $\hat{Y} = 10,040X^{2.997}$ where $\hat{Y}$ is the conductivity and X is the decimal fraction of pores greater than .294 mm in diameter. Conductivities changed with depth in stages. The 10 and 30 cm sampling depths usually had conductivities in excess of 400 cm/hr. The 70 and 110 cm depths had conductivities closer to 200 cm/hr, while subsoil conductivities at depths of 130 cm, 150 cm, and deeper dropped to values less than 60 cm/hr in most soil pits and less than 10 cm/hr in some pits.

7. Soil moisture-tension relationships indicated that most of the water held in the soil and subsoil between the tensions of 0 and 100 cm of water were held at tensions of 10 cm or less. It was also shown that the subsoils retained a greater percentage of water held at

saturation than the soil when subjected to a tension of only 10 cm. At this tension the samples from the 10 and 30 cm depths lost 41% of the water held at saturation, the 70 and 110 cm samples lost 23% of the water held at saturation, while the 130 and 150 cm samples lost only 6%. Thus the subsoil will be considerably closer to complete saturation than the soil for the same tension.

8. The data collected from a soil pit located on a nearby watershed indicated that the hydrologic properties of the subsoil at this location were somewhat different than the hydrologic properties of the study slope subsoil. The values of the hydraulic conductivity, bulk density, and percentage of large diameter pores for this soil pit were all below the values for the study slope soil pits. The results of this data proved to be inconclusive toward describing the nature of subsurface flow on the study slope soil pits.

The results of this study indicate that the hydrologic properties and the antecedent soil moisture conditions were conducive to translatory flow through the soil and subsoil on a steep, forested slope. The hydrologic properties of the soil indicate that the movement of water occurs primarily in a vertical direction as unsaturated flow. The hydrologic properties of the subsoil at several soil pits indicate that a zone of saturation would probably form within these subsoils. The subsurface movement of water was concluded to be in the form of translatory flow if soil moisture conditions permitted. Tensiometer data for the study slope indicated that unsaturated flow did occur through the soil, a zone of saturation did form near one soil pit, and that the soil moisture conditions were conducive to translatory flow.

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APPENDICES

APPENDIX A

CALCULATIONS FOR HYDROLOGIC PROPERTIES

The following calculations were conducted for the indicated hydrologic property using values obtained in the lab. For most cases these calculations were performed for every soil sample taken. The values given in Tables 1 and 2 are mean values obtained from several individual soil samples. Following each equation is one example using values from soil sample number 1 in most instances.

Bulk Density

$$\text{Bulk Density (gm/cm}^3\text{)} = \frac{\text{mass of oven dry soil}}{\text{volume of oven dry soil}}$$

$$\text{Bulk Density} = \frac{\text{oven dry wt} - (\text{ring wt} + \text{can wt})}{\text{volume of soil retainer ring}}$$

Example for spl 1:

$$\frac{294.8\text{ gm} - (152.1\text{ gm} + 40.1\text{ gm})}{135.6\text{ cm}^3} = .757\text{ gm/cm}^3$$

Total Porosity

$$\text{Total Porosity (fraction of total volume)} = \frac{\text{volume of pore space}^{1}}{\text{volume of soil ring}}$$

[1] The volume of pore space is taken to be the volume of water in the soil between the saturated and oven dry conditions. Since the density of water is 1 gm/cm^3 the mass of the water removed is equal to the volume removed.

$$\text{Total Porosity} = \frac{\begin{array}{l}\text{saturated wt - (drying clamp wt + wt of} \\ \text{cheese cloth and rubber band) - ring wt} \\ \text{+ (ring wt + can wt) - oven dry wt}\end{array}}{\text{volume of soil ring}}$$

Example for spl 1:

$$\frac{\begin{array}{l}1036.1 \text{ gm} - (702.4 \text{ gm} + 2.0 \text{ gm}) \\ - 152.1 \text{ gm} + (152.1 \text{ gm} + 40.1 \text{ gm}) \\ - 294.8 \text{ gm}\end{array}}{135.6 \text{ cm}^3} = .567$$

Hydraulic Conductivity

The equation used to calculate the hydraulic conductivity of each sample was derived from Darcy's law which reads:

$$q = K \frac{\Delta H}{L}$$

where q is the flux, K is the hydraulic conductivity, and $\Delta H/L$ is the hydraulic gradient (Hillel, 1971).

The flux (q) can also be written as:

$$q = V/At$$

where V is the volume of water passing through the soil, A is the cross-sectional area it flows through, and t is the time the volume of water takes to flow through.

The hydraulic gradient (H/L) can also be written as:

$$H/L = \frac{Hi - Ho}{L}$$

where Hi is the hydraulic head at the inflow point, Ho is the hydraulic head at the outflow point, and L is the length of the soil sample.

By combining the new terms the following form of Darcy's law is obtained:

$$V/At = K\left(\frac{Hi - Ho}{L}\right)$$

Rearranging terms the equation can be solved for the hydraulic conductivity.

$$K = \frac{VL}{At(Hi - Ho)}$$

The hydraulic conductivity for each sample was found by measuring the volume of water (V) that passed through a soil sample having a cross-sectional area (A), and length (L), over a measured period of time (t). The hydraulic head at the inlet could be varied, while the head at the outlet was kept at zero. Thus the equation used to calculate hydraulic conductivity for each sample was as follows:

$$K = (V)(L/AtHi)$$

Example spl 264:

$$(85\ \text{cm}^3)\,\frac{(6\ \text{cm})(3600\ \text{sec/hr})}{(22.56\ \text{cm}^2)(30\ \text{sec})(10\ \text{cm})} = 270\ \text{cm/hr}$$

Water Content

The following equation was used to calculate the water content of the soil samples after each level of tension had been applied. It is similar to the equation used for total porosity except that the weight of the sample after each level of tension is used instead of the saturated weight.

$$\text{water content} = \frac{\begin{array}{l}\text{tension wt - ring wt + clamp wt -} \\ \text{(cheese cloth and rubber band wt +} \\ \text{clamp tare) - (ring wt + can wt) -} \\ \text{oven dry wt}\end{array}}{\text{volume of soil ring}}$$

$$\begin{array}{l}\text{example for spl} \\ \text{1, 10 cm tension}\end{array} = \frac{\begin{array}{l}\text{299.5 gm - 152.2 gm + 702.4 gm -} \\ \text{(2.0 gm + 702.4 gm) - (152.2 gm +} \\ \text{40.1 gm) - 294.8 gm}\end{array}}{135.6\ \text{cm}^3}$$

water content = .315

Pore Diameter Size

The pore diameter size classes given in Table 3 were found using an equation derived from the height of capillary rise equation which reads:

$$h = \frac{2\gamma \cos a}{g(\rho l - \rho g)r}$$

where h is the height of capillary rise, ρ_g is the density of gas (which is generally neglected), ρ_1 the density of the liquid, g the acceleration of gravity, r the capillary or pore radius, α the contact angle, and γ the surface tension between the liquid and the air (Hillel, 1971).

Solving for the radius of the capillary tube the equation reads:

$$r = \frac{2\gamma \cos\alpha}{h\rho g}$$

For the purposes of this thesis the contact angle α is considered zero, so $\cos\alpha = 1$. The surface tension of water at 20°C is 72.5 gm/cm^2, the acceleration of gravity is 981 cm/sec^2, and the density of water is 1 gm/cm^3. The value of the height of rise (h) is equal to the tension applied to the tension table: 10, 20, 30, 40, 60, and 100 cm of water.

Thus the minimum radius of the pores drained at 10 cm of tension was found by using the above values in the equation.

$$r = \frac{(2)(72.5\ \text{gm/sec}^2)(1)}{10\ \text{cm}(1\ \text{gm/cm}^3)(981\ \text{cm/sec}^2)}$$

$$r = .0147\ \text{cm}$$

$$\text{dia} = .294\ \text{mm}$$

The pores draining at 10 cm of tension will be .294 mm in diameter or larger. The range of diameters drained by the other tensions was found using the other values of tension applied to the tension tables.

<u>Pore Size Distribution</u>

The decimal fraction of the total porosity in each of the pore size classes was calculated using the mass of the water drained at each tension. Because the density of water is 1 gm/cm^3 the volume of water drained at each tension will equal the volume of the pores drained at each tension. Equations and examples follow for the first two pore size diameter classes.

decimal fraction of pores:

$$.294 \text{ mm in dia} = \frac{\text{spl sat. wt - 10 cm wt}}{\text{spl sat. wt - oven dry wt}}$$

Example spl 1:

$$\frac{333.7 \text{ gm} - 299.5 \text{ gm}}{333.7 \text{ gm} - 256.8 \text{ gm}} = .445$$

decimal fraction of pores:

$$.147 \text{ mm---}.294 \text{ mm in dia} = \frac{\text{10 cm wt - 20 cm wt}}{\text{spl sat. wt - oven dry wt}}$$

Example spl 1:

$$\frac{299.5 \text{ gm} - 293.7 \text{ gm}}{333.7 \text{ gm} - 256.8 \text{ gm}} = .075$$

APPENDIX B

Table I. Mean values of stone content.

Soil Pit No.	Depth	Stone Content: Volume	Stone Content: Percent by Volume
	cm	cm^3	percent
1	10	45.5	33.6
	30	48.5	35.8
	70	46.2	34.1
2	10	36.5	26.9
	30	42.3	31.2
	70	31.0	22.9
3	10	46.3	34.1
	30	49.5	36.5
	70	58.3	43.0
5	10	59.7	44.0
	30	58.3	43.0
	70	52.0	38.3
6	10	46.7	34.4
	30	59.5	43.9
	70	56.2	41.4
7	10	58.8	43.4
	30	53.3	39.3
	70	46.8	34.5
9	10	52.7	38.8
	30	57.3	42.3
10	10	65.8	48.6
	30	61.5	45.3
	70	78.8	58.1

Table II. Soil moisture percent of total volume for increasing tensions.[1]

Soil Pit No.	Depth	Tension, cm of water						
		0	10	20	30	40	60	100
	cm							
1	10	60.8	37.5	33.1	31.5	30.7	30.0	28.0
	30	63.8	37.5	34.0	32.7	32.0	31.4	29.7
	70	60.3	46.1	43.5	42.2	41.1	40.2	39.2
	110	63.1	49.6	47.3	46.4	45.7	44.7	43.7
	130	55.4	49.2	48.2	47.5	47.0	45.9	44.7
	150	57.6	53.0	51.6	51.0	50.1	49.1	47.9
	200	60.0	52.3	50.3	49.6	48.7	47.4	46.0
2	10	65.6	40.6	37.2	35.6	34.7	33.4	32.0
	30	62.4	40.0	36.8	35.4	34.5	33.4	32.2
	70	61.9	46.4	43.1	41.8	40.7	39.7	38.7
	110	63.6	49.5	47.0	45.8	44.6	43.6	42.5
	130	57.9	50.1	48.7	48.0	47.5	46.5	45.4
	150	61.0	50.0	48.6	48.0	47.0	46.2	45.3
	200	62.1	53.4	50.1	49.1	48.0	46.8	45.5
	250	61.4	51.4	50.4	49.3	48.8	48.2	46.6
3	10	69.6	32.5	30.2	29.1	28.4	27.6	26.5
	30	66.7	34.4	32.2	30.9	30.3	29.4	28.4
	70	63.6	35.4	32.4	31.1	30.2	29.4	28.8
	110	62.8	36.7	33.5	32.0	31.1	30.3	29.6
	160	59.1	50.0	49.0	48.0	46.7	46.2	45.4
5	10	67.2	32.3	30.1	29.1	28.5	28.0	26.5
	30	64.1	35.2	32.9	31.9	31.3	30.7	29.4
	70	59.0	41.8	39.4	38.3	37.8	36.5	35.4
	110	57.9	42.9	41.1	40.2	39.9	38.4	37.6
	150	52.5	44.0	42.9	42.3	41.8	41.0	40.4
	190	53.5	44.8	44.0	43.1	42.9	42.1	40.9

Table II. (Continued)

Soil Pit No.	Depth	Tension, cm of water						
		0	10	20	30	40	60	100
	cm							
6	10	70.7	45.3	38.3	36.0	33.9	31.1	30.0
	30	54.0	45.4	43.7	42.6	41.8	40.4	39.1
	70	53.5	46.1	44.6	43.6	42.7	41.5	39.9
	110	58.0	50.1	46.2	43.8	40.9	37.2	36.9
	150	54.6	49.4	47.6	46.2	44.4	41.3	41.1
7	10	69.1	34.9	31.6	30.4	29.6	29.0	28.2
	30	68.1	37.5	33.6	32.1	31.1	30.4	29.5
	70	54.2	42.4	39.3	37.5	35.7	32.1	32.0
9	10	63.1	34.0	30.2	28.6	27.4	26.6	25.2
	30	65.9	33.2	29.3	27.8	26.5	25.8	24.6
10	10	58.1	34.5	31.1	29.1	28.2	27.3	25.6
	30	60.1	40.4	36.4	34.1	33.2	32.4	30.8
	70	51.9	41.9	39.1	36.7	36.1	35.5	32.6
	110	55.5	41.8	39.6	38.0	37.5	37.0	34.8
12	10	65.1	33.4	30.5	29.6	28.8	27.9	27.1
	30	63.6	36.2	33.3	32.3	31.6	30.6	29.6
	70	52.3	44.5	43.1	42.5	42.1	41.6	40.6
	110	53.0	46.1	45.0	44.4	44.0	43.5	42.7
	150	57.9	51.8	50.8	50.1	50.0	48.5	47.9

[1] Each value represents the mean of six samples.

APPENDIX C

SOIL PROFILE DESCRIPTIONS

Soil Pit 1

Horizon	Depth, cm	Description
0_1	5-3	Leaves, needles, twigs, cones, etc.
0_2	3-0	Partially decomposed leaves, needles, twigs, cones, etc.
A_1	0-18	Dark-brown (7.5 YR 3/2 moist) gravelly clay loam of weak fine and very fine granular structure; slightly hard when dry, and slightly sticky and slightly plastic when wet; 30% pebbles, many medium, common coarse and very fine, and fine roots; many fine and very fine interstitial pores; common fine and very fine tubular pores; pH 6.0; fragments of basalt and andesite; clear wavy boundary.
A_3	18-42	Dark-brown (7.5 YR 3/2 moist) gravelly clay loam of weak fine and very fine granular structure; friable when moist, and slightly sticky and slightly plastic when wet; 35% pebbles; common medium and fine, many very fine roots; many fine and very fine interstitial pores; common fine, very fine, and medium, and few coarse tubular pores; pH 6.1; coarse fragments of basalt; clear wavy boundary.
B_2	42-76	Brown (7.5 YR 4/4 moist) very gravelly clay loam of weak very fine sub-angular blocky structure; friable when moist, and slightly sticky and slightly plastic when wet; 50% pebbles and cobbles; common medium and very fine, and few coarse and fine roots; common fine and very fine interstitial pores; common very fine and medium, and few fine tubular pores; pH 6.0; coarse fragments weathered and highly weathered breccia; clear wavy boundary.

Horizon	Depth, cm	Description
C	76-150+	Dark yellowish-brown (10 YR 4/4 moist) very gravelly clay loam of massive structure; friable when moist, and slightly sticky and slightly plastic when wet; 65% pebbles and cobbles; common medium, and few coarse, fine, and very fine roots; common very fine and very few medium and fine tubular pores; pH 5.7; coarse fragments weathered and highly weathered breccia.

Soil Pit 2

Horizon	Depth, cm	Description
O_1	3-2	Leaves, needles, twigs, cones, etc.
O_2	2-0	Partially decomposed leaves, needles, twigs, cones, etc.
A_1	0-20	Dark brown (7.5 YR 3/2 moist) clay loam of weak fine and very fine granular structure; slightly hard when dry, friable when moist, and slightly sticky and plastic when wet; 15% pebbles and shot; common coarse, medium, and very fine, and few fine roots; many fine and very fine interstitial pores; common very fine and few fine tubular pores; pH 6.0; coarse fragments andesite and breccia; clear wavy boundary.
B_2	20-78	Brown (7.5 YR 4/4 moist) clay loam of weak very fine subangular blocky structure; slightly hard when dry, firm when moist, and slightly sticky and plastic when wet; 15% pebbles; many coarse, common medium, few very fine, very few fine roots; many very fine and common fine interstitial and tubular pores; pH 5.8; coarse fragments weathered and highly weathered breccia and andesite; gradual smooth boundary.

Horizon	Depth, cm	Description
IIC_1	78-139	Brown (7.5 YR 4/4 moist) gravelly clay of massive structure; firm when moist, sticky and plastic when wet; 45% pebbles; common medium and very few fine and very fine roots; few medium and fine and common very fine tubular pores; few fine charcoal chips; pH 5.6; coarse fragments weathered and highly weathered breccia and andesite; gradual smooth boundary.
IIC_2	139-217+	Dark yellowish-brown (10 YR 4/4 moist) very gravelly clay loam of massive structure; friable when moist, slightly sticky and plastic when wet; 65% pebbles and cobbles; very few coarse, medium, and very fine roots; few very fine and very few medium fine tubular pores; pH 5.4; coarse fragments weathered and highly weathered breccia and andesite.

Soil Pit 3

0_1	4-3	Leaves, needles, twigs, cones, etc.
0_2	3-0	Partially decomposed leaves, needles, twigs, cones, etc.
A_1	0-22	Dark brown (7.5 YR 3/2 moist) gravelly clay of weak fine and very fine granular structure, soft when dry, slightly sticky and slightly plastic when wet; 45% pebbles and cobbles; many medium, common coarse and very fine, and few fine roots; many fine and very fine interstitial pores; few coarse, medium, fine and very fine tubular pores; pH 6.6; few fine charcoal chips; coarse fragments weathered breccia, tuffs, and andesite; clear wavy boundary.

Horizon	Depth, cm	Description
B_1	22-42	Brown (7.5 YR 4/4 moist) gravelly clay loam of moderate fine granular and moderate very fine subangular blocky structure; slightly hard when dry, slightly sticky and slightly plastic when wet; 45% pebbles and cobbles; common coarse and medium, few very fine, and very few fine roots; many fine and very fine interstitial pores; many medium, common coarse and fine, and few very fine tubular pores; pH 6.4; coarse fragments weathered breccia and andesite; clear smooth boundary.
B_2	42-89	Brown (7.5 YR 4/4 moist) very gravelly loam; moderate subangular blocky structure; firm when moist, slightly sticky and slightly plastic when wet; 50% pebbles and cobbles, common coarse, and medium, few very fine, and very few fine roots; many very fine interstitial pores, common fine and very fine tubular pores; pH 6.1; coarse fragments weathered breccia and andesite; clear wavy boundary.
C_1	89-159	Brown (7.5 YR 4/4 moist) very gravelly clay loam of massive structure; friable when moist, slightly sticky and slightly plastic when wet; 60% pebbles and cobbles; common medium and fine, few coarse and very fine roots; common medium, fine and very fine, and few coarse tubular pores; pH 6.0; coarse fragments weathered breccia, tuff, and basalt; approximately 10% of horizon consists of old root channel, 15 cm in diameter, abrupt smooth boundary.
IIC_2	159-238+	Yellowish brown (10 YR 5/8 moist) gravelly silty clay of massive structure; very friable when moist; slightly sticky and plastic when wet; 45% pebbles and cobbles; common coarse, few medium and fine, very few very fine roots; common very fine and very few fine tubular pores; pH 5.9; coarse fragments weathered and highly weathered breccia and tuff.

Horizon	Depth, cm	Description
		Soil Pit 4
0_1	4-2	Leaves, needles, twigs, cones, etc.
0_2	2-0	Partially decomposed leaves, needles, twigs, cones, etc.
A_1	2-0	Brown (10 YR 4/3 moist) gravelly clay loam of weak fine granular structure; 45% pebbles and cobbles; friable when moist, slightly sticky and slightly plastic when wet; many medium, common fine and very fine roots; many medium, and common and fine and very fine interstitial pores, many fine, common medium and very fine tubular pores; coarse fragments basalt; gradual smooth boundary.
B_2	31-60	Brown (10 YR 4/3 moist) very gravelly clay loam of moderate, fine subangular blocky structure; 55% pebbles and cobbles; firm when moist, slightly sticky and slightly plastic when wet; few fine and very fine, and very few medium roots; common medium and fine, few very fine interstitial pores, common medium, fine and very fine tubular pores; many coarse fragments of basalt; gradual smooth boundary.
C	60-75+	Brown (10 YR 4/3 moist) very gravelly clay loam of massive structure; 60% cobbles; firm when moist, slightly sticky and slightly plastic when wet; few fine and very few very fine roots; few fine and very fine interstitial pores, few very fine and very few fine tubular pores; many coarse fragments of basalt.
		Soil Pit 5
0_1	5-4	Leaves, needles, twigs, cones, etc.
0_2	4-0	Partially decomposed leaves, needles, twigs, cones, etc.

Horizon	Depth, cm	Description
A_1	0-40	Dark brown (7.5 YR 3/2 moist) very gravelly clay of medium very fine granular structure; soft when dry, slightly sticky and slightly plastic when wet; 55% pebbles; many coarse, medium, common fine and very fine roots; many fine and very fine interstitial pores; pH 6.2; coarse fragments of andesite; clear wavy boundary.
IIB_2	40-106	Brown (7.5 YR 4/4 moist) gravelly clay of medium fine and very fine subangular blocky structure; friable when moist, sticky and plastic when wet; 25% pebbles; common medium, few coarse, and very few fine and very few fine and very fine roots; many very fine interstitial pores; common medium and very fine, and few fine tubular pores; pH 5.8; coarse fragments of weathered andesite and basalt; gradual smooth boundary.
IIC	106-172+	Brown (7.5 YR 4/4 moist) gravelly clay of massive structure; firm when moist, sticky and plastic when wet; 30% pebbles and cobbles; few medium and very few fine and very fine roots; common very fine and few fine tubular pores; pH 5.8; coarse fragments of weathered andesite and basalt.

Soil Pit 6

Horizon	Depth, cm	Description
0_1	4-2	Leaves, needles, twigs, cones, etc.
0_2	2-0	Partially decomposed leaves, needles, cones, twigs, etc.
A_1	0-15	Brown (7.5 YR 4/4 moist) gravelly clay of moderate fine granular structure; friable when moist, nonsticky and nonplastic when wet; many fine and common very fine roots; many fine and very fine, common medium interstitial pores; coarse fragments of andesite; clear smooth boundary.

Horizon	Depth, cm	Description
IIB_2	15-70	Strong brown (7.5 YR 5/6 moist) clay of moderate fine subangular blocky structure; firm when moist, sticky and plastic when wet; few very fine and very few fine roots; many fine and very fine interstitial pores; common fine and very fine tubular pores; clear, smooth boundary.
IIC	70-210+	Brown (7.5 YR 4/4 moist) clay of massive structure; firm when moist, slightly sticky and slightly plastic when wet; few coarse, very few medium, fine, and very fine roots; common medium and very fine, few fine interstitial pores, common fine, few medium and very fine tubular pores; coarse fragments of weathered and highly weathered breccia.

Soil Pit 7

Horizon	Depth, cm	Description
0_1	4-2	Leaves, needles, twigs, cones, etc.
0_2	2-0	Partially decomposed leaves, needles, twigs, cones, etc.
A_1	0-35	Brown (10 YR 4/3 moist) gravelly clay loam of weak fine granular structure; 40% pebbles; friable when moist, slightly sticky and slightly plastic when wet; many medium and fine, common very fine interstitial pores; many medium, and common fine and very fine roots; coarse fragments andesite and basalt; gradual smooth boundary.
B_2	35-70	Brown (10 YR 4/3 moist) gravelly clay loam of moderate, fine subangular blocky structure; firm when moist, slightly sticky and slightly plastic when wet; 35% pebbles; few fine and very fine, and very few medium roots; common fine and very fine, and few medium interstitial pores; coarse fragments of basalt, andesite, and weathered breccia; clear smooth boundary.

Horizon	Depth, cm	Description
C	70-120+	Brown (10 YR 4/3 moist) gravelly clay of massive structure; firm when moist, slightly sticky and slightly plastic when wet; 30% pebbles and cobbles; few medium and fine, and very few coarse and very fine roots; few fine and very fine interstitial pores; coarse fragments weathered and highly weathered breccia.

Soil Pit 8

Horizon	Depth, cm	Description
0_1	5-3	Leaves, needles, twigs, cones, etc.
0_2	3-0	Partially decomposed leaves, needles, twigs, cones, etc.
A_1	0-25	Brown (7.5 YR 4/4 moist) very gravelly clay loam of moderate medium granular structure; firm when moist, nonsticky and nonplastic when wet; many medium, fine and very fine roots; many medium and fine common coarse and very fine interstitial pores; 75% pebbles and shotty concretions; coarse fragments andesite and basalt; gradual smooth boundary.
A_{32}	25-46	Brown (7.5 YR 4/4 moist) very gravelly clay loam of moderate medium granular structure; 75% pebbles; firm when moist, sticky and plastic when wet; common medium, fine and very fine roots; many medium and fine, common very fine interstitial pores, coarse fragments andesite and basalt; gradual smooth boundary.
C	46-60+	Brown (7.5 YR moist) gravelly clay of massive structure; 40% pebbles and cobbles, firm when moist, sticky and plastic when wet; common fine and very fine, very few coarse roots, common fine and very fine interstitial pores, common fine and very fine, and few medium tubular pores; coarse fragments of basalt and weathered breccia.

Horizon	Depth, cm	Description
		Soil Pit 9
O_1	5-2	Leaves, needles, twigs, cones, etc.
O_2	2-0	Partially decomposed leaves, needles, twigs, cones, etc.
A_1	0-20	Brown (7.5 YR 4/4 moist) gravelly clay loam of weak fine granular structure; firm when moist, slightly sticky and slightly plastic when wet; 40% pebbles; many fine, common medium and very fine roots; common coarse, medium, and very fine, few fine interstitial pores; coarse fragments andesite and basalt; gradual smooth boundary.
B_2	20-55	Brown (7.5 YR 4/4 moist) gravelly clay loam of moderate medium subangular blocky structure; firm when moist, sticky and plastic when wet; 45% pebbles and cobbles; common medium, few fine and very fine roots; many medium and fine, common very fine interstitial pores, common fine, and few medium and very fine tubular pores; coarse fragments basalt and weathered breccia; clear smooth boundary.
C	55-72+	Brown (10 YR 4/3 moist) clay loam of massive structure; very firm when moist, sticky and plastic when wet; 40% cobbles; few coarse, very few medium fine and very fine roots; few fine and very fine interstitial pores, few medium and fine tubular pores; coarse fragments basalt and weathered breccia.
		Soil Pit 10
O_1	4-2	Leaves, needles, twigs, cones, etc.
O_2	2-0	Partially decomposed leaves, needles, twigs, cones, etc.

Medium	Depth, cm	Description
A_1	0-24	Dark brown (10 YR 3/3 moist) gravelly clay loam of moderate fine granular structure; firm when moist, slightly sticky and slightly plastic when wet; 50% pebbles and shotty concretions; common medium, fine, and very fine roots; many medium and fine interstitial pores, coarse fragments andesite and basalt; clear smooth boundary.
B_2	24-66	Brown (7.5 YR 4/4 moist) gravelly clay loam of moderate medium subangular blocky structure; firm when moist, slightly sticky and slightly plastic when wet; 55% pebbles and cobbles; common medium, and few fine and very fine roots; common medium and fine interstitial pores, common medium and few fine tubular pores; coarse fragments of basalt and andesite, clear smooth boundary.
C	66-81+	Brown (7.5 YR 4/4 moist) rocky clay loam of massive structure; firm when moist, slightly sticky and slightly plastic when wet; 50% cobbles; few coarse and very few fine and very fine roots; common very fine, and few fine tubular pores; coarse fragments of basalt and breccia.

Soil Pit 11

0_1	5-3	Leaves, needles, twigs, cones, etc.
0_2	3-0	Partially decomposed leaves, needles, twigs, cones, etc.
A_1	0-12	Brown (10 YR 4/3 moist) silty clay loam of weak medium granular structure; friable when moist, slightly sticky and slightly plastic when wet; 20% pebbles and shotty concretions; many fine, common medium and very fine roots; many medium and common fine and very fine interstitial pores; coarse fragments of andesite and basalt; gradual smooth boundary.

Medium	Depth, cm	Description
B_2	12-78	Brown (10 YR 4/3 moist) clay loam of moderate medium granular structure; firm when moist, slightly sticky and slightly plastic when wet; 20% pebbles and cobbles; many fine and common very fine roots; few medium and fine interstitial pores, common fine and very fine tubular pores; coarse fragments of basalt and weathered breccia; gradual smooth boundary.
C_1	78-180	Dark yellowish brown (10 YR 4/4 moist) clay of massive structure, firm when moist, slightly sticky and slightly plastic when wet; 10% cobbles; few coarse and fine roots; common medium and few fine tubular pores; coarse fragments of weathered and highly weathered breccia; gradual smooth boundary.
C_2	180-210+	Light yellowish brown (10 YR 6/4 moist) clay of massive structure; firm when moist, sticky and plastic when wet; very few very fine interstitial pores, very few very fine tubular pores; coarse fragments of weathered and highly weathered breccia.

Soil Pit 12

0_1	4-2	Leaves, needles, twigs, cones, etc.
0_2	2-0	Partially decomposed leaves, needles, twigs, cones, etc.
A_1	0-30	Dark yellowish brown (10 YR 4/4 moist) very gravelly clay loam of moderate medium granular structure; friable when moist, slightly sticky and slightly plastic when wet; 50% pebbles and shotty concretions; many medium and common fine and very fine roots; many coarse, medium, and fine interstitial pores; coarse fragments of basalt and andesite; clear smooth boundary.

Medium	Depth, cm	Description
B_2	30-51	Dark yellowish brown (10 YR 4/4 moist) gravelly silty clay loam of weak medium subangular blocky structure; firm when moist, sticky and slightly plastic when wet; common fine and very fine roots; common medium, fine and very fine interstitial pores, coarse fragments of basalt and weathered breccia, clear, smooth, boundary.
C	51-135+	Yellowish brown (10 YR 5/6 moist) silty clay loam of massive structure; firm when moist; sticky and slightly plastic when wet; very few very fine roots; few fine interstitial pores; few fine and very fine tubular pores; coarse fragments of weathered and highly weathered breccia.

APPENDIX D

STATISTICAL CALCULATIONS

Sample of Statistical Tests Used to Compare Mean Conductivities Between Two Sampling Depths

Question: Is there a significant (95%) difference between the mean hydraulic conductivities of samples from the 30 cm depth and the samples from the 70 cm depth?

1. (a) First it must be determined if the population variances are equal. The F statistic is used, which reads:

$$F = s_1^2/s_2^2$$

where s_1^2 and s_2^2 are the sample variances for populations 1 and 2.

(b) The null hypothesis states that the population variances are equal. Ho: $\sigma_1^2 = \sigma_2^2$

(c) The alternative hypothesis states that the population variances are not equal. Ha: $\sigma_1^2 \neq \sigma_1^2$

(d) Sample conductivity data for soil pit 1 (cm/hr).

30 cm . . . 677, 760, 383, 546, 811, 211

70 cm . . . 236, 258, 239, 166, 185, 134, 83, 268

(e) Statistical manipulation of data.

	30 cm	70 cm
n	6	8
Σx	3388	1569
$\bar{x}$	565	196
Σx^2	2, 182, 976	337, 831
$(\Sigma x)^2/n$	1, 913, 091	307, 831
SS	269, 885	30, 111
s^2	53, 977	4, 302

(f) Tabular F for 5 and 7 degrees of freedom from the 5% points table is 3.97.

The calculated F is: $F = s_1^2/s_2^2 = \dfrac{53,977}{4,302} = 12.54$

(g) Conclusion: Since $12.54 > 3.97$, reject Ho and conclude that there is a difference in variability of conductivities at this significant level.

2. (a) To test for differences in mean conductivity between sampling depths the null hypothesis states that there is no difference, Ho: $\mu_1 = \mu_2$. The alternative hypothesis states that one of the mean conductivities is greater than the other, Ha: $\mu_1 > \mu_2$.

(b) The statistic used to test the hypothesis is the t statistic, which for this purpose reads:

$$t = \frac{\bar{x}_1 - \bar{x}_2}{s_{\bar{d}}^2}$$

where $\bar{x}_1$ and $\bar{x}_2$ are sample means and $s_{\bar{d}}^2$ is the sample variance of the difference between the means.

(c) Since $\sigma_1^2 \neq \sigma_2^2$ and $n_1 \neq n_2$, $s_{\bar{d}}^2 = \dfrac{n_2 s_1^2 + n_1 s_2^2}{n_1 n_2}$

so that $s_{\bar{d}}^2 = \dfrac{8(53,977) + 6(4,302)}{6(8)}$

$$s_{\bar{d}}^2 = 9,534$$

(d) The sample t can now be calculated:

$$t = \frac{565 - 196}{9,534} = 3.78$$

(e) Since $\sigma_1^2 \neq \sigma_2^2$ and $n_1 \neq n_2$, the tabular t is found by calculating an average between the tabular t for 5 degrees of freedom (t_1) and the tabular t for 7 degrees of freedom (t_2). This value is known as t* and is calculated as follows:

$$t^* = \frac{n_2 s_1^2 t_1 + n_1 s_2^2 t_2}{n_2 s_1^2 + n_1 s_2^2}$$

$$t^* = \frac{(8)(53,977)(2.015) + (6)(4302)(1.895)}{8(53,977) + 6(4302)}$$

$$t^* = 2.008$$

(f) Conclusion: Since 3.780 > 2.008, reject Ho and conclude that the 30 cm mean conductivity is significantly (95%) greater than the 70 cm mean conductivity.