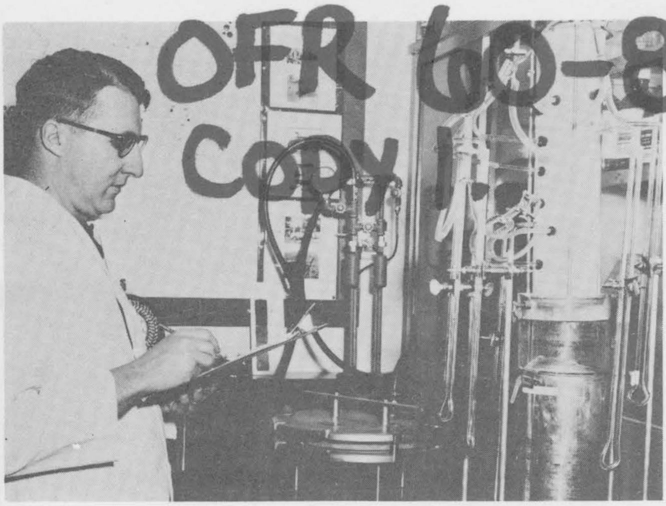


UNITED STATES

DEPARTMENT OF THE INTERIOR

GEOLOGICAL SURVEY

Johnson, A.I.

Specific Yield And Related Properties

An Annotated Bibliography

HYDROLOGIC LABORATORY

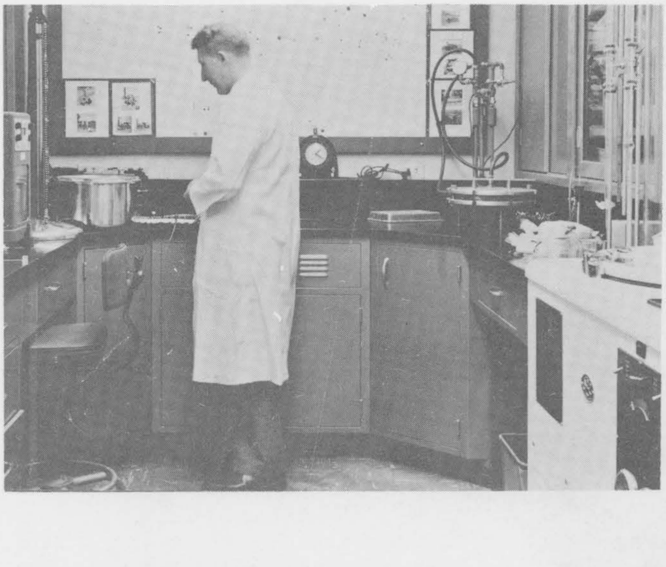
BIBLIOGRAPHY NO. 2

PREPARED IN COOPERATION

WITH THE CALIFORNIA

DEPARTMENT OF

WATER RESOURCES



UNITED STATES
DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

SPECIFIC YIELD AND RELATED PROPERTIES

AN ANNOTATED BIBLIOGRAPHY

By A. I. Johnson, D. A. Morris, and R. C. Prill

PREPARED IN COOPERATION WITH THE
CALIFORNIA DEPARTMENT OF WATER RESOURCES

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SPECIFIC YIELD AND RELATED PROPERTIES
AN ANNOTATED BIBLIOGRAPHY

PART I

A. I. Johnson, D. A. Morris, and R. C. Prill

INTRODUCTION

Purpose and Scope of Bibliography

The increasing use of ground water from many major aquifers in the United States has emphasized the need for a more thorough understanding of the important hydrologic property called "specific yield." Early in 1957 a cooperative agreement was made with the California Department of Water Resources to conduct laboratory and field research on specific yield. One purpose of this 5-year research project is to prepare a bibliography summarizing important literature on specific yield and methods for its determination. This report represents an initial phase of the bibliography and covers primarily literature from the United States for the period from 1900 through the first half of 1958.

The bibliography was prepared under the supervision of Joseph F. Poland, research geologist, G. F. Wortz, Jr., and Harry D. Wilson, Jr., successive district chiefs, Ground Water Branch, Sacramento, Calif.

Organization of Bibliography

This annotated bibliography summarizes much of the important literature on specific yield and related properties but does not pretend to be complete. Because of the complexity of the subject it has been necessary to investigate other technical fields related to specific yield. These include the fields of hydraulic and hydrologic engineering, petroleum geology, and soil science. Principal technical terms are defined later in this report. Papers not directly applicable or those containing incidental information on specific yield and related properties have been omitted.

The annotations have been prepared to be informative rather than descriptive. The writers have attempted so to digest and summarize the papers that the reader can obtain the important points of the article without reading the entire paper. Sufficient information is included to enable the reader to evaluate the article and to assist him in deciding which papers to investigate further in reference to his particular problem. However, the reader probably will have to go to the original papers for many specific details and data.

Nearly all annotations are quoted or adapted from the text of the article and thus are in essentially the author's language. Exceptions are indicated by appropriate notes.

The abstracts are arranged alphabetically by author. Author and subject indexes are provided. If several papers are listed for the same author(s), the references are in chronological order.

Acknowledgments

Many individuals made contributions to this bibliography by giving advice or by furnishing information relating to source material. Valuable assistance was provided by Mr. L. B. James, chief geologist, and Mr. R. T. Bean, supervising engineering geologist, California Department of Water Resources.

Definitions

The following list presents definitions of the most important and the most frequently used technical terms used in this bibliography. (Synonyms and parallel terms are given in parentheses.) The table at the end of this definition is included for the convenience of the reader in converting the various pressure terms used in the annotations. Also for the convenience of the reader, the definitions were taken only from the following sources:

Aldrich, D. G., and others, 1956, Report of definitions approved by the committee on terminology, Soil Science Society of America: Soil Sci. Soc. America Proc., v. 20, no. 3, p. 430-440.

American Geological Institute, 1957, Glossary of geology and related sciences: Natl. Acad. Sci., Natl. Research Council Pub. 501, 325 p.

Calhoun, J. C., 1953, Fundamentals of reservoir engineering: Norman, Okla., Oklahoma Univ. Press, p. 97.

Meinzer, O. E., 1923, Outline of ground-water hydrology, with definitions: U.S. Geol. Survey Water-Supply Paper 494, 71 p.

Piper, C. S., 1950, Soil and plant analysis: New York, Interscience Publishers, Inc., p. 100.

Theis, C. V., 1938, The significance and nature of the cone of depression in ground-water bodies: Econ. Geology, v. 33, no. 8, p. 889-902.

<u>Adhesion potential</u>. See Adhesion tension.

<u>Adhesion tension</u>. The energy of attraction across the interface between a solid and liquid, or between two immiscible liquids. Numerically, it is equal to the sum of surface tensions of the substances singly, less the interfacial tension (American Geological Institute, 1957, p. 4).

<u>Capillarity</u>. The attractive force between two unlike molecules, illustrated by the rising of water in capillary tubes of hairlike diameters or the drawing-up of water in small interstices, as those between the grains of a rock (American Geological Institute, 1957, p. 44).

<u>Capillary conductivity</u>. The ratio of the water-flow velocity to the driving force in unsaturated soil. The calculation is valid under conditions where flow velocity is proportional to driving force. For example, in practical units when the driving force is expressed in terms of the hydraulic gradient, capillary conductivity is the ratio of flow velocity to hydraulic gradient and has the dimension of velocity. As saturation is approached capillary conductivity approaches the hydraulic conductivity (Aldrich and others, 1956, p. 431).

Capillary fringe. A zone, in which the pressure is less than atmospheric, overlying the zone of saturation and containing capillary interstices some or all of which are filled with water that is continuous with the water in the zone of saturation but is held above that zone by capillarity acting against gravity (American Geological Institute, 1957, p. 44).

Capillary potential. See Capillary pressure.

Capillary pressure. The difference in pressure existing between two phases, air-fluid, gas-fluid, or between fluids, measured at points of the interface and occurring in the interconnected phases in a rock (American Geological Institute, 1957, p. 44).

Coefficient of storage. The coefficient of storage of an aquifer is the volume of water it releases from or takes into storage per unit surface area of the aquifer per unit change in the component of head normal to that surface (modified by the Ground Water Branch from Theis, 1938, p. 894).

Connate water. See Retained water.

Diffusion. The spreading out of molecules, atoms, or ions into a vacuum, a fluid, or a porous medium, in a direction tending to equalize concentrations in all parts of a system (American Geological Institute, 1957, p. 82).

Diffusion coefficient. A parameter in diffusion calculations, having the dimensions distance-squared divided by time. It varies with the nature of the particles diffusing, the nature of the diffusion medium, and temperature. It is expressed by the formula $l^2 t^{-1}$

where l = length, t = time (American Geological Institute, 1957, p. 82).

Field capacity. The amount of water held in a soil by capillary action after gravitational water has percolated downward and drained away; expressed as the ratio of the weight of water retained to the weight of dry soil (Aldrich and others, 1956, p. 433).

Free energy. The capacity of a system to perform work, a change in free energy being measured by the maximum work obtainable from a given process (American Geological Institute, 1957, p. 116).

Freezing point. The temperature at which a liquid solidifies; especially applied to pure water which has freezing point at 0°C or 32°F under normal atmospheric pressure (American Geological Institute, 1957, p. 116).

Heat of wetting. The heat of wetting is the heat evolved when dry soil is wetted. It is expressed in calories per gram of dry soil (Piper, 1950, p. 100).

Hydraulic gradient. Pressure gradient. As applied to an aquifer it is the rate of change of pressure head per unit of distance of flow at a given point and in a given direction (Meinzer, 1923, p. 38).

Hydrostatic pressure. The pressure exerted by the water at any given point in a body of water at rest. That of ground water is generally due to the weight of water at higher levels in the same zone of saturation (American Geological Institute, 1957, p. 143).

Hygroscopic coefficient. The hygroscopic coefficient of a soil is the ratio of (1) the weight of water which at that temperature the soil will absorb if, after completely dry, it is placed in free contact with a saturated atmosphere until equilibrium is established, to (2) the weight of the soil when dry. This ratio is expressed as a percentage (Meinzer, 1923, p. 24).

Imbibition. The tendency of granular rock, or any porous medium, to "imbibe" a fluid, usually water under the force of capillary attraction, and in the absence of any pressure (American Geological Institute, 1957, p. 146).

Infiltration. The downward entry of water into soil (Aldrich and others, 1956).

Interfacial tension. The force tending to reduce the area of contact between two liquids or between a liquid and a solid (American Geological Institute, 1957, p. 151).

Irreducible saturation. See Retained water.

Irreducible water. See Retained water.

Moisture equivalent. The ratio of (1) the weight of water which the soil, after saturation, will retain against a centrifugal force 1,000 times the force of gravity to (2) the weight of soil when dry. The ratio is stated as a percentage (Meinzer, 1923, p. 25).

Moisture potential. See Capillary pressure.

Moisture tension. The equivalent negative gauge pressure, or suction, in the soil moisture. Soil-moisture tension is equal to the equivalent negative or gauge pressure to which water must be subjected in order to be in hydraulic equilibrium, through a porous permeable wall or membrane, with the water in the soil (Aldrich and others, 1956, p. 435).

Moisture retention. See Specific retention.

Osmotic pressure. If a pure solvent is separated from a solution by a membrane permeable only to molecules of the solvent, the extra pressure which muct be applied to the solution in order to prevent flow of solvent into it by osmosis is known as the osmotic pressure of the solution (American Geological Institute, 1957, p. 207).

Permeability. The permeability (or perviousness) of rock is its capacity for transmitting a fluid. Degree of permeability depends upon the size and shape of the pores, the size and shape of their interconnections, and the extent of the latter. It is measured by the rate at which a fluid of standard viscosity can move a given distance through a given interval of time. The unit of permeability is the Darcy (American Geological Institute, 1957, p. 217).

Porosity. The ratio of the aggregate volume of interstices in a rock or soil to its total volume. It is usually stated as a percentage (Meinzer, 1923, p. 19).

Pressure membrane. A membrane, permeable to water and only very slightly permeable to gas when wet, through which water can escape from a soil sample in response to pressure gradients (Aldrich and others, 1956, p. 436).

Pressure potential. See Capillary pressure.

Relative permeability. The ratio of the effective permeability to a given fluid at a definite saturation to the permeability at 100 percent saturation (Calhoun, 1953, p. 97).

Retained water. Interstitial water held by molecular attraction against gravity, in isolated interstices or as water vapor occupying interstices from which liquid water has been withdrawn (American Geological Institute, 1957, p. 245).

Specific retention. As applied to a rock or soil it is the ratio of (1) the volume of water which, after being saturated, it will retain against the pull of gravity to (2) its own volume. It is stated as a percentage (Meinzer, 1923, p. 29).

Specific yield. As applied to a rock or soil it is the ratio of (1) the volume of water which, after being saturated, it will yield by gravity to (2) its own volume. This ratio is stated as a percentage (Meinzer, 1923, p. 28).

Transmissibility coefficient. The rate of flow of water, in gallons per day, at the prevailing temperature, through each vertical strip of the aquifer 1 foot wide having a height equal to the thickness of the aquifer and under a unit hydraulic gradient (American Geological Institute, 1957, p. 303).

Uniformity coefficient. The quotient of (1) the diameter of a grain that is just too large to pass through a sieve that allows 60 percent of the material, by weight, to pass through, divided by (2) the diameter of a grain that is just too large to pass through a sieve that allows 10 percent of the material, by weight, to pass through (Meinzer, 1923, p. 45).

Wilting coefficient. The ratio of (1) the weight of water in the soil at the moment when (with gradual reduction in the supply of soil water) the leaves of the plants growing in the soil first undergo a permanent reduction in their water content as the result of a deficiency in the supply of soil water to (2) the weight of the soil when dry. The ratio is expressed as a percentage (Meinzer, 1923, p. 24).

BIBLIOGRAPHY

Alway, F. J., and McDole, G. R., 1917, Relation of the water-bearing capacity of a soil to its hygroscopic coefficient: Jour. Agr. Research, v. 9, no. 2, p. 27-71.

Thirteen soils, representing some of the most important soil types in Nebraska, were used in a study of the relation of the water-retaining capacity of a soil to tis hygroscopic coefficient. The soils ranged in texture from a coarse sand to a silt loam. In one experiment, cylinders 3 feet in length and 6 inches in diameter were filled with uniform air-dry loam soils. The soils in the cylinders were placed in capillary connection with the loess floor of a greenhouse. A 2-inch layer of gravel was placed on top of the soil and 15 pounds of water was allowed to infiltrate from the top of the column. Eight inches of moist soil was then applied at the top of the soil column and the cylinder was sealed. One-inch samples were taken throughout the total length of the soil columns at the end of 31, 44, 54, 96, 104, 109, and 126 days. Soil-moisture equilibrium was not established in the columns until after 96 days. The average moisture content of samples taken from 96 to 126 days was 13.8 percent. This value was very close to the moisture equivalent which was 13.5.

Another experiment was performed similar to the one described above except that a 6-inch layer of sand and gravel was placed between the loam soil and the greenhouse floor. After 26 days the loam soil had a moisture

Pressure Conversion Factors

Taken or calculated (*) from Zimmerman, O. T., and Lavine, Irvin, 1955, Conversion factors and tables: Industrial Research Service, Inc., Dover, N. H.

	atm	cm mercury	mm mercury	in. mercury	cm water	mm water	in. water	ft water	psi	bars
atm	1	76 (0°C)	760 (0°C)	29.921 (32°F)	1033.2487* (39.2°F)	10332.4867* (39.2°F)	406.79 (39.2°F)	33.899 (39.2°F)	14.6960	1.013250
cm mercury (0°C)	0.0131579	1	10* (0°C)	0.3937 (0°C)	13.5956* (39.2°F)	135.9563* (39.2°F)	5.3526* (39.2°F)	0.44605 (39.2°F)	0.193368	0.01333221
mm mercury (0°C)	0.001315789	0.1* (0°C)	1	0.03937 (0°C)	1.3596* (39.2°F)	13.5956* (39.2°F)	0.5353* (39.2°F)	0.044604 (39.2°F)	0.0193368	0.001333223
in. mercury (32°F)	0.0334211	2.54000508*	25.4000508 (0°C)	1	34.5376* (39.2°F)	345.3765* (39.2°F)	13.596* (39.2°F)	1.1330 (39.2°F)	0.491157	0.0338640
cm water (39.2°F)	0.00096779*	0.07355* (32°F)	0.73553* (32°F)	0.028958* (32°F)	1	10* (39.2°F)	0.3937* (39.2°F)	0.03281* (39.2°F)	0.014223*	0.000980616*
mm water (39.2°F)	0.00009678*	0.007355* (32°F)	0.07355* (32°F)	0.0028963* (32°F)	0.1* (39.2°F)	1	0.03937* (39.2°F)	0.003281* (39.2°F)	0.001422*	0.000098062*
in. water (39.2°F)	0.0024582	0.18683* (32°F)	1.86828* (32°F)	0.073554 (32°F)	2.54000508* (39.2°F)	25.4000508* (39.2°F)	1	0.0833333* (39.2°F)	0.0361265	0.00249077*
ft water (39.2°F)	0.0294990	2.24193 (0°C)	22.4193* (0°C)	0.882647 (32°F)	30.48006096* (39.2°F)	304.8006096* (39.2°F)	12* (39.2°F)	1	0.433518	0.0298899*
psi	0.0680457	5.17148 (0°C)	51.7148	2.03601 (32°F)	70.3099* (39.2°F)	703.0988* (39.2°F)	27.681 (39.2°F)	2.3067 (39.2°F)	1	0.0689473
bars	0.9869233	75.0062 (0°C)	750.062 (0°C)	29.530 (32°F)	1020.6858* (39.2°F)	10206.858* (39.2°F)	401.844* (39.2°F)	33.487 (60°F)	14.50385	1

Multiply unit at left by number in column to get unit at top of column.

content that averaged 6 percent more than the continuous loam columns of the above experiment. The soil immediately above the sand and gravel averaged 6 percent higher in moisture content than the soil near the surface.

Another experiment was carried out in which 2-inch layers of 6 different soils were differently arranged in 7 cylinders, 18 inches long and 6 inches in diameter. The soil at the bottom of the cylinder was placed in capillary contact with the loess floor of the greenhouse. Seventeen pounds of water was added to the surface and moisture samples were taken after a period of from 6 to 9 days. No difference in moisture content was observed in the respective layers regardless of the sequence in which they occurred, except when dune sand was used to separate the layers. The dune sand caused higher moisture contents in the overlying layers. When the sequence of layers was not interrupted by a sand layer, the moisture content of the respective layers was very close to the moisture equivalent.

Archie, G. E., 1942, The electrical-resistivity log as an aid in determining some reservoir characteristics: Am. Inst. Mining Metall. Engineers Petroleum Technology Jour., no. 1, Tech. Pub. 1422.

Determination of the significance of the resistivity of a producing formation as recorded by the electrical log seemed to depend principally on the application of empirical relationships established in the laboratory between certain physical properties of a reservoir rock and the formation factor. This study presented some of the laboratory data and indicated its application to quantitative studies of the electric log. These included the relation of porosity and permeability to resistivity, and the determination of connate water content, porosity, and salinity of formation water from the electrical log.

The usefulness of the electrical-resistivity log in determining reservoir characteristics was controlled mainly by the accuracy with which the true resistivity of the formation could be determined, by the adequacy of detailed data relating resistivity measurements to formation characteristics, by the amount of available information pertaining to the conductivity of connate or formation waters, and by the abundance of geologic knowledge regarding probable changes in facies, both vertically and horizontally (as it affects the electrical properties of the reservoir).

Although the use of the electrical-resistivity method to determine the physical nature of reservoir-material under favorable conditions was recommended, the author suggested that considerable care by exercised in its application.

Beckett, S. H., Blaney, H. F., and Taylor, C. A., 1930, Irrigation water requirement studies of citrus and avocado trees in San Diego County, California, 1926 and 1927: California Univ. Agr. Expt. Sta. Bull. 489.

A part of this study of irrigation-water requirements of citrus and avocado trees was concerned with sampling of experimental plots to determine soil moisture. Samples were taken in 1-foot sections to a depth of 6 feet, unless sampling was prevented by the absence of soil or other unfavorable conditions. During the irrigation season, samples were taken before and after each irrigation and at intervals of 2 to 3 weeks between irrigations. During the winter periods samples were taken after each of the major rains. The average moisture equivalents were determined for most of the experimental plots. The moisture equivalent ranged from 7.3 to 16.9 percent. In soils in the lower part of the range, the field capacity generally exceeded the moisture equivalent, but in soils in the higher part of the range the field capacity generally was comparable to the moisture equivalent.

Bethel, F. T., and Calhoun, J. C., 1953, Capillary desaturation in unconsolidated beads: Jour. Petroleum Technology, v. 5, no. 8, p. 197-202.

Capillary-desaturation curves reflect in general the character and arrangement of the pores within porous media and the distribution of fluids within the pores. This study attempted to evaluate the qualitative and quantitative significance of two variables, interfacial tension and contact angle, as they are ordinarily used to characterize surface forces.

The experiments determined the capillary desaturation of n-octane by water on a series of packs of glass beads which had been rendered oil-wet to different degrees by treatment with silicones. Capillary desaturation was accomplished by the restored-state technique (in which the core sample, saturated with a wetting fluid, is placed on a porous diaphragm permeable to the wetting phase) using a Buchner funnel containing a fritted glass disk. All parts of the cell were rendered oil-wet. The cell was packed by adding glass beads while jarring and rotating the cell filled with 10 cubic centimeters of n-octane. By using this method, packs were obtained which were within 1 percent of being 100 percent saturated. The degree of saturation of the packed beads during desaturation was computed from the volumetric displacement of n-octane, measured in a pipette, as the displacing pressures were increased.

The desaturation curves indicated a regularity of increasing displacement pressures as the oil-wetness increases. This was used to compute apparent contact angles between n-octane and water on glass. The family of desaturation curves also revealed a regular crossover, so that the residual wetting-liquid-saturation values were in reverse order to the displacement pressures. A semilogarithmic variation was shown between the apparent contact angle and the residual saturation. The desaturation data were interpreted to mean that the wetting liquid ceases to be a continuous phase within individual pores before the capillary pressure can be raised sufficiently to force out the wetting phase completely. The saturation of wetting liquid at which this break in liquid continuity occurs was higher as the oil-wetness decreased. These comments had a bearing on the true significance of capillary-desaturation curves. Unless the porous system was sufficiently wet by the displaced liquid to maintain a continuity of liquid to sustain flow from all pores within the system, the desaturation curve lost its physical significance as a measure of pore character. A change in saturation on a desaturation curve which represents true capillary equilibrium or moisture equilibrium at all interfaces within the porous medium must consist of two parts - that due to entry of new pores and that due to change of pendular rings of liquid at the liquid-solid contacts. This suggested that such techniques as the mercury injection or desaturation of a wetting liquid with a gas would be the most reliable type of desaturation curve for computing pore radius.

Any other gross measurements such as electrical resistivity and relative permeability which depend on fluid distribution with a pore matrix would be sensitive to the variations shown in the capillary desaturation curves.

Blake, G. R. and Corey, A. T., 1951, Low pressure control for moisture release studies: Soil Sci., v. 72, p. 327-331.

This paper discussed a simplified, inexpensive, commercially available apparatus that has been found to be entirely suitable for accurate pressure control in the 0 to 2 atmosphere range. The pressure-control unit consisted of two regulators in series, the first giving a reasonably constant pressure output for the second which was a precision regulator. The success of the control system was dependent primarily on this precision regulator which operated from relatively constant supply pressures up to 100 pounds per square inch, but was recommended to be used at 60 pounds per square inch. A desirable arrangement was to use a pressure source greater than 60 pounds per square inch and to maintain the pressure through the first regulator at 60 pounds. This pressure then became a constant source for the precision regulator. Experience with this system of two regulators indicated that pressure settings could be made within 1 millimeter of water and if direct jarring of the regulator itself was avoided, the pressure settings would remain constant indefinitely. Although the precision regulator gave excellent results, an asbestos tension table was found to be more convenient and more expedient to determine moisture characteristics in the tension range between 0 and about 150 centimeters of water.

Bodman, G. B. and Colman, E. A., 1944, Moisture and energy conditions during downward entry of water into soils: Soil Sci. Soc. America Proc. (1943), v. 8, p. 116-122.

In this study particular attention was given to the distribution of moisture content and moisture potential within soil columns during downward penetration of water.

Two soils, the Yolo silt loam and the Yolo sandy loam, were packed to uniform apparent densities in 2-inch-diameter brass tubes. Stacks of cylinders, each 5 millimeters in height, were built up and bolted together to form tubes approximately 40 centimeters in height.

The rates of water entry and water penetration into the soil column were measured until water had reached a predetermined depth, at which time the water supply was removed, the surplus water pipetted from the soil surface, and the soil column rapidly sliced into moisture samples. Each infiltration test was terminated before water had penetrated to the bottom of the soil column.

The permeability of the soils in the water-saturated state was determined directly in a constant-head permeameter and in the water-unsaturated state was determined indirectly from water flow and moisture-potential data obtained during infiltration. Pressure potentials were not measured directly in the infiltration experiment because of the time lag involved in the use of porous-clay tensiometers. Instead, soil samples in porous-clay plates were wetted from water reservoirs until equilibrium was reached with the known tensions at which each reservoir was held. Corresponding values of moisture content and pressure potential were thus obtained for the soils studied.

The wetted zones of the two soil columns were divided into the four distinct parts listed below.

1. The surface 1-centimeter layer reached a moisture content approximating saturation at the time the water had penetrated to a depth of about 10 centimeters.
2. Below the saturated surface layer the soil-moisture content decreased rapidly with depth until at about 6 centimeters it reached a value somewhat higher than half way between the moisture equivalent and saturation.
3. Below the 6-centimeter depth described above, the moisture content decreased with depth until the dry soil was reached. Within this zone, downward progress of the wetted front was accompanied by an increasing moisture content of the soil. This interval represented the wetting portion of the infiltration zone as contrasted to the transmitting zone between it and the soil surface.
4. The wetting zone terminated abruptly at the wet front. Water moved into the dry soil as a distinctly wet wave and no visible evidence of any diffusion of moisture ahead of the obviously wetted soil was observed. Data suggested that the wet-front moisture content was characteristic of the soil but was probably

independent of the depth of the wetted soil.

The moisture equivalent did not seem to represent the minimum moisture at which ready capillary movement could take place. However, the wet-front moisture contents may be significant for the soils tested, possibly representing the lower limit of ready capillary permeability under the condition of soil wetting.

The moisture potential-depth curves of the transmitting zone maintained a constant type of configuration, although the thickness of the zone increased as infiltration proceeded. The average potential gradient and permeability in this zone are of dominant importance in the rate of delivery of water to the zone being wetted below. The infiltration rate decreased with time in both soils. This decrease was not believed due to permeability changes in the upper soil layer, but rather to a decrease in moisture-potential gradient resulting primarily from an increasing depth of penetration.

Similar values of pressure potential at the base of the transmitting zone indicated a possible significance of the determination of permeability at a pressure potential of about -2.8×10^4 ergs per gram.

The results of this study indicated that the soil moisture must be raised to the minimum values found at the wet front before water can move across this front to wet the dry soil below. Under conditions imposed by drainage it became apparent that the wet-front moisture represented the lower limit of capillary permeability.

Bodman, G. B., and Day, P. R., 1943, Freezing point of a group of California soils and their extracted clays: Soil Sci., v. 55, no. 3, p. 225-246.

This study was undertaken to investigate by the cryoscopic or freezing-point method the magnitudes of the free energy of soil water at different water contents, and some of the factors affecting these magnitudes. This method determined the difference between the free energy of water in bulk and that of water in moist soil.

The term "moisture potential" (μ) was used to define the free engergy of escape of soil water, when consideration was given to the influence of pressure, temperature, and solutes, as follows:

$$\mu = \frac{L_f \cdot T}{T_o}$$

in which

μ = moisture potential (in text),

L_f = latent heat of fusion of water (= $-3.336 X 10^9$ ergs gm. -1)

(The quantity of heat necessary to change one gram of solid to a liquid with no temperature change),

T = freezing-point depression of water (^oC),

T_o = freezing point of pure water,

and the moisture potential is obtained in ergs per gram of water.

The soil sample, weighing approximately 5 grams, was enclosed in a small Dewar flask equipped with a thermojunction and the assemble was immersed in a kerosene bath at a suitable temperature. The reference temperature was produced and maintained by means of a second Dewar flask filled with a mixture of ice and water and containing the other junction of the thermocouple. The electromotive force produced by the thermocouple was measured by a sensitive galvanometer and potentiometer, and the difference in temperature between the two junctions was computed from calibration tables.

The freezing-point depression at the moisture equivalent was studied for 14 nonsaline soils. The entire range in moisture potentials was from -0.24×10^6 to -2.3×10^6, but the values of 12 soils were within the narrower range of -0.73×10^6 to -1.6×10^6 ergs per gram. The mean value for all was -1.12×10^6 ergs per gram which was lower than that obtained by the porous-cup-tensiometer technique.

The authors believed that an important factor, not measured by the tensiometer, was depressing the moisture potentials as calculated from the freezing points. Consequently, the electrical conductivities were measured for the soils to determine the abundance of dissolved solutes. From these data the observed variations in moisture potential at the moisture equivalent were found to be due largely to variations in solute concentrations.

Soils were puddled in two different ways and freezing measurements were made subsequently. Drastic changes in mechanical methods had only a slight influence on the energy curves. This was attributed to a high degree of stability of the microstructure.

The moisture contents of soils at moisture potentials between -1×10^6 and -30×10^6 ergs per gram were obtained from the freezing-point depression curves and examined in relation to the corresponding contents of clay finer than 2 microns. The results indicated a distinct correlation between clay content and moisture content at a constant moisture potential. The "clay coefficient" was considerable less at lower energy values. This suggested that factors other than the total amount of clay played an important role as the energy levels of the soil moisture diminished.

Consideration was given to the possible effect of differences in the mineralogical character of the colloidal clays upon the freezing-point curves of soils. To test this, clay samples were taken from seven soils. Freezing points at different moisture contents were obtained for these soil clays and for montmorillonite and kaolinite clay minerals. X-ray diffraction patterns of six soils were examined. The different soil clays displayed widely differing moisture-potential curves. Kaolinitic types seemed to give results similar to coarser grained material, though there was no precise relationship between the position of the energy curve and the type of predominating clay mineral present. Soils of related genetic origin had similar energy curves. Unless this was due to a balancing of factors, the surface properties, extent of surface, and structural configuration were very much the same for colloids from soils of related origin.

Bouyoucos, G. J., 1922, Relation between heat of wetting, moisture equivalent, and unfree water: Soil Sci., v. 14, p. 431-440.

The results of an investigation to ascertain the relationships that exist between heat of wetting, unfree water, and moisture equivalent of soils were reported. A close and consistent relationship between the heat of wetting and the content of unfree water was indicated, but no apparent relationship between the heat of wetting and moisture equivalent or between the content of unfree water and moisture equivalent was observed. The moisture-equivalent method did not give an absolutely equivalent moisture for all of the soils. Some of the fine-textured and colloidal soils contained considerably more moisture than their moisture equivalent.

Bouyoucos, G. J., 1929, A new, simple, and rapid method for determining the moisture equivalent of soils, and the role of soil colloids on this moisture equivalent: Soil Sci., v. 27, p. 233-241.

A new method was presented for obtaining a soil-moisture content comparable to moisture equivalent. The principle of the method was based on pulling water from the soil by suction or vacuum forces. A Buchner funnel was connected with a suction flask. A filter paper was placed in the funnel and the funnel was three-fourths filled with soil in the natural condition, and water was introduced. After the soil became saturated, a suction of about 20 millimeters was applied and was continued for a period of 10 minutes after all the standing water had disappeared from the soil. The funnel was covered with a moist thin cloth during the suction process.

The author stated that the method seemed reliable for determining the comparative water-holding powers of soils and gave a realistic comparison to the moisture equivalents of the soils. If directions were followed explicitly, results could be duplicated within about 1 percent. The results obtained by this method indicated a remarkably close relationship between the water-holding power and the colloidal content of soils as determined by the hydrometer method. Because the relationship was so close, if one result was known, the other could be calculated. No relationship was noted between the amount of coarse silt and sand, and the moisture equivalent.

Bouyoucos, G. J., 1935, A comparison between the suction method and the centrifuge method for determining the moisture equivalent of soils: Soil Sci., v. 40, p. 165-171.

The suction method for determining the comparative water-holding power of soils, or the moisture equivalent, was further studied and improved. The method involved pulling water out of the soils by suction.

Two different systems were used to produce the vacuum: (a) a water-jet or filter pump, and (b) a displacement-type vacuum pump. The method as finally standardized gave very close agreement in replicate tests. The water-jet pump gave almost exactly the same results as the vacuum pump. The suction method was compared with the standard centrifuge method for determining moisture equivalent, and the results indicated that the two methods agree closely in the majority of soils investigated, but, in general, the suction method gave somewhat higher values than the centrifuge method.

Bouyoucos, G. J., 1947, Capillary rise of moisture in soil under field conditions as studied by the electrical resistance of plaster of Paris blocks: Soil Sci., v. 64, p. 71-81.

This study investigated the capillary rise of moisture in soils of uniform texture under field conditions.

Open-end tanks 3 feet high and 3 feet in diameter were filled with soil and imbedded into field soils in which the position of the water table ranged from the land surface to deeper than 6 feet. Plants were grown in the tanks to reduce the moisture content of the soils to different values and to different depths. The plants were removed, the tanks were covered to prevent evaporation, and the moisture contents of the tanks were determined by the electrical resistance of plaster of Paris blocks as equilibrium was established under the new moisture gradient.

Although the amount of capillary movement was somewhat greater in sandy loam than in clay loam, the results of the tests over a 2-year period (a longer period was studied but not presented in this paper) indicated that unless the water table was high enough to wet the soil in the tank, that capillary rise added very little to the moisture content in a tank that was controlled below field capacity. The rate of capillary rise was extremely slow and the height of movement was very small.

Bouyoucos, G. J., 1954, New type electrode for plaster of Paris moisture blocks: Soil Sci., v. 78, p. 339-342.

A new type of electrode utilized in plaster of Paris blocks for soil moisture determinations was evaluated. The electrodes were made from 20-mesh stainless steel screen and were 15/16 inch long and 4/16 inch wide. Two electrodes were placed 3/16 inch apart in a plaster of Paris block, the outside dimensions of which were 1 11/16 by 11/16 by 1 4/16 inch. The blocks were treated with a nylon resin for longer life.

This screen-type electrode constituted a very important improvement in the design of plaster of Paris blocks. The following improvements were noted: The plaster became enmeshed in the screen producing a reinforced multiple contact between the electrode and the casting, thus ensuring stable performance. The blocks had little or no capacitance, which was important because the instrument had no provision for counterbalancing capacitance. The blocks were affected only slightly by "stray currents." The new blocks were sensitive to changes in soil moisture at tensions ranging from 260 to 330 centimeters of water.

Boyd, J. R., 1925, Procedure for testing subgrade soils: Public Roads, v. 6, no. 2, p. 34-41.

Changes in the procedures for testing subgrade material, including the moisture-equivalent test, were given. In the modified moisture-equivalent test a saturated 5-gram soil sample was placed in a damp closet and was allowed to drain overnight. The sample was then placed in a Babcock cup and centrifuged at a force of 1000 times gravity for 1 hour. The Babcock cup was provided with a brass cap to prevent evaporation. The moisture-equivalent tests were run in duplicate. The author suggested that the variation between the two results should not exceed 1 percent for moisture equivalents of less than 15 percent and 2 percent for moisture equivalents of 15 percent or greater.

The moisture equivalent provided a means of comparing directly the relative ease with which subgrades of different physical characteristics could be drained. Soils having high moisture equivalents were relatively more difficult to drain than those with low moisture equivalents. Soils having a moisture equivalent greater than 20 percent were considered to be doubtful for use as subgrade materials.

Briggs, L. J., and McLane, J. W., 1910, Moisture-equivalent determinations and their application: Am. Soc. Agronomy Proc., v. 2, p. 138-147.

The term "moisture equivalent" was used to designate the maximum percentage of moisture a soil could retain in opposition to a known centrifugal force. As a standard basis of comparison, a centrifugal force equal to 1,000 times the force of gravity was adopted. This procedure for determining the moisture equivalent of a soil was given as follows: Each soil was put through a 2-millimeter sieve; a sheet of filter paper was placed in the bottom of a perforated cup; and amount of soil sufficient to give a packed layer 1 centimeter thick was used; the soil was moistened and was allowed to stand for about 24 hours, protected from evaporation; a small amount of water was then added to the soil and the sample was centrifuged for 40 minutes at a velocity of 2,440 revolutions per minute at a temperature of 20°C and the moisture content determined.

The moisture equivalent values of the soils studied ranged from 2 percent in coarse sands to 50 percent in the heavier clays. The authors pointed out that this method of moisture-equivalent determination provided a single-value expression of the moisture retentiveness of a soil measured in a definite way.

Broadfoot, W. M., 1954, Core vs. bulk samples in soil-moisture tension analysis: U.S. Forest Service, 5th Forest Expt. Sta. Occassional Paper 135, p. 22-25.

A comparison was made of the moisture contents of core- and disturbed-soil samples at various tensions. Two procedures designated as A and B were used.

In procedure A, 5 pairs of core and bulk samples were obtained from each of 3 soils: Commerce silty clay, Briensburg silt loam, and

Basket very fine sandy loam. The cores were obtained in stainless steel rings 2 3/4 inches in diameter and 3/4 inches in depth. Each bulk sample was taken close to or adjacent to its corresponding paired core. In the laboratory, the bulk samples were prepared by hand crushing and by sieving through a U.S. Standard, 9-mesh sieve. The soil from the bulk samples was placed in rings that were covered at one end with a filter paper and cheesecloth. The soil cores were also covered with cheesecloth and filter paper. The paired samples were tested simultaneously on a tension table (supported semi-permeable membrane attached to hanging-water column) at moisture tensions of 5 and 60 centimeters of water; and in capillary pressure cells (an enclosed chamber containing a semi-permeable membrane at one end and a pressure tap at the other) at 1/10-, 1/3-, and 1-atmosphere pressure, and, using Visking-sausage casing as a semi-permeable membrane, at 3- and 15-atmospheres pressure. From 6 to 24 hours were allowed for moisture to reach equilibrium for samples on the tension table and from 24 to 48 hours for samples in the pressure cells. Soil-moisture values obtained from procedure A, though useful for comparative purposes, were somewhat higher at high tensions than data previously obtained on two of the soils. The thickness of the filter paper-cheesecloth combination apparently caused the water columns to break tension when the moisture content was reduced.

In procedure B, 16 paired samples of a silt-loam loess were taken at different depths. The cores were taken in brass rings, 2 inches in diameter and ½ inch deep. Except in the 3- and 15-atmosphere determinations where cheesecloth was not used, this procedure differed from procedure A in that a single thickness of cheesecloth was used to hold the sample in the ring.

At tensions up to 1 atmosphere, the bulk samples retained more water than the core samples. The difference was less in sandy soils than in finer textured soils. At higher tensions no consistent difference between core and disturbed samples was observed. Because the core samples represented the "undisturbed" or field condition of the soil, the author suggested that cores should be used in determinations of moisture tension in the range of 0 to 1 atmosphere.

Browning, G. M., 1941, Relation of field capacity to moisture equivalent in soils of West Virginia: Soil Sci., v. 52, no. 6, p. 445-450.

Some important West Virginia soils were selected to study the relationship between the moisture equivalent and field capacity of soils in a humid region. Moisture-equivalent values were determined by the method proposed by Goldbeck and Jackson. For determining field capacity, cylinders of galvanized iron, 6 inches in diameter and 14 inches long, were driven into the ground to a depth of 2 or 3 inches. The cylinders were filled with water, and 48 hours after the moisture had disappeared from the surface, samples for field-capacity determinations were taken from layers 6 inches and 6 to 12 inches below the surface soil.

The results of this study indicated that the ratio of the field capacity to the moisture equivalent increased as the moisture equivalent decreased. The average ratio of field capacity to moisture equivalent for the surface soil was unity at a moisture equivalent of 23, for the layer 6 inches below the surface soil was unity at 21.5, and for the layer 6 to 12 inches below the surface soil was unity at 20.5 percent. The rate of change of increase was small for soils with moisture equivalents in the range of about 12 to 35 percent. For moisture equivalents of less than 12, the change was more rapid. This general relationship was the same for the different sampling depths.

The excess moisture drained from well drained soils within a relatively short time. Changes of moisture content after 1 to 2 hours of drainage were small and an approximately constant value was reached in 24 to 48 hours. In the impermeable soils, the excess water applied to the surface did not pass through the soil for 2 to 3 days. Equilibruim was not established even after a period of 48 hours from the time the excess water disappeared from the surface. The large percentage of the excess water disappeared from the surface. The large percentage of the excess water that drained from well drained soils in a relatively short period of time after a rain or after an application of water indicated that this water movement was through the larger sized pores. The author stated that the moisture in soils below pF 1.6 (logarithm of capillary potential) moved rapidly and was influenced principally by forces of gravity. The tension at the moisture equivalent was about pF 2.7. Moisture at tensions from pF 1.6 to 2.7 moved more slowly and was affected definitely by both capillary and gravitational forces. The author considered a tension of pF 1.6 to be a more logical indication of the amount of noncapillary porosity than field capacity because this tension has been shown to include the pores through which water moves more rapidly as a result largely of the force of gravity. As a measure of the amount of water a soil will hold after a rain or after an application of water, a field capacity value somewhat higher than pF 1.6 was suggested because appreciable movement of gravitational water in many soils occurred until this point was approached.

Browning, G. M., and Milan, F. M., 1941, A comparison of the Briggs-McLane and Goldbeck-Jackson centrifuge methods for determining the moisture equivalent of soils: Soil Sci., v. 51, no. 4, p. 273-278.

Comparison was made of the moisture equivalent values of a number of important soil types of the United States as determined by the Goldbeck-Jackson method using a regular centrifuge equipped with trunnion cups and Gooch crucibles as containers for the soil, and as determined by the Briggs-McLane method.

The moisture equivalents obtained by the Briggs-McLane procedure ranged from 3.9 to 40.9 percent. The range of differences between the two methods was -0.9 to +1.9 percent, the Goldbeck-Jackson procedure averaging 0.76 percent less than the determinations made by the Briggs-McLane procedure. This study indicated that there was a significant difference in results obtained by the two methods unless all values obtained by the Goldbeck-Jackson method were corrected by use of the regression equation.

Buckingham, Edgar, 1907, Studies on the movement of soil moisture: U.S. Dept. Agri. Bull. 38, p. 29-61.

The author discussed in detail the movement of water in soil and defined capillary potential and capillary conductivity.

The attraction of soil for water depends on the water content. This capillary potential for a given state of packing and temperature decreases as moisture content increases. When the soil is completely saturated with water, it takes only an infinitesimal amount of work to remove a finite mass of water, or the capillary potential $\psi = 0$. The capillary potential for a given water content differs from soil to soil; the retentiveness of different soils, or even of the same soil in different states of structure is different. If different soils are subjected to the same force, gravitational or other, which tends to pull water away from them, this force drains some soils more than others. However, the final value of the capillary potential must be the same in all, because it just balances the same outside pull. Hence, in some soils the retention is lower than in others to raise the capillary potential to a given value. To determine patterns of water distribution after imbibitional intervals from 53 days to 10½ months, the author constructed six soil columns, each 48-inches long and 2½ inches in diameter. Evaporation was controlled, water was introduced at the bottom, and the soil and water was allowed to come to equilibrium. The results indicated that where the duration of the test was not more than 2 months, the final state of equilibrium had not been reached.

Capillary conductivity was then considered. If the water content of the soil was gradually reduced, the water retreated more and more into the very fine pore spaces, which remained full. Hence, if the total number of such capillary channels were not diminished the capillary conductivity would not change much. However, as the water content was reduced, the number of continuous paths diminished and communication was maintained only through water films which stretched from one drop to another over the immediate surfaces of the solid soil grains. Further reduction of water content caused the films to become thinner or to lose their conducting properties. Thus, large conductivity occurred at first, decreasing as the number of continuous paths through capillary water decreased. When most of these paths had been broken and the film paths were becoming of importance, though still short, a rapid decrease in conductivity occurred. As the film paths got longer, conductivity decreased rapidly. When the soil approached complete dryness, the films began to break or to lose the properties of liquid water and the conductivity fell rapidly towards zero.

The author then considered mathematically the capillary water held in prismatic wedges, the electrical conductivity of soils, and the conductivity from wet soils to dryer ones when placed in contact.

Buehrer, T. F. and Rosenblum, M. S., 1939, A new dilatometer for determining bound water in soils and other colloidally dispersed materials: Jour. Phys. Chemistry, v. 43, p. 941-951.

The dilatometric method of determining bound water in soils, plant materials, and other colloidally dispersed systems has been employed by a number of investigators. This paper described a new apparatus and technique for its use.

The principal features of the dilatometer used are an elongated freezing tube, a specially ground joint, and a three-way stopcock. A condenser jacket enclosed a capillary tube and scale installed above the dilatometer.

Two calibrations of the apparatus were necessary: (1) determination of the capillary scale reading in terms of true volumes; (2) determination of the amount of expansion obtainable with known amounts of water. A correction also was needed to compensate for the unavoidable expansion of the toluene which is forced up into the capillary as a result of the freezing of the water.

The procedure followed was: The soil sample and water were weighed and allowed to stand in the sample tube to attain uniform moisture distribution. The sample tube was then placed in the freezing tube, and anhydrous toluene was added to cover the sample. The apparatus was then lowered into the freezing bath and was allowed to come to a temperature of -3°C (jacket temperature maintained at 30°C). The air was removed by evacuation. The sample was then frozen, first by an ice-salt mixture at -10°C, and then by return to the -3°C bath. The toluene level in the capillary tube rose when freezing began and continued to rise until it became constant for 10 to 15 minutes. This usually took from 30 minutes to 1 hour. From the reading of the toluene column before and after freezing, the expansion was calculated by using the calibration equation. The true volume expansion was then calculated by applying the expansion correction. Finaly, the amount of water frozen was calculated for any one dilatometer by the simplified equation (if above temperatures maintained):

$$\frac{0.1593(S_f - S_i) \times 100}{W} = \text{percent of water frozen}$$

where

S_f = final scale reading
S_i = initial scale reading
W = weight of water taken initially

The author proved that when conditions were properly controlled and the proper corrections applied, results obtained by any one observer were reproducible within 1 to 2 percent. However, the interpretation of any results obtained should be made only in the light of the conditions of the determination.

Burrows, W. C. and Kirkham, Don, 1958, Measurement of field capacity with a neutron meter: Soil Sci. Soc. America Proc., v. 22, no. 2, p. 103-105.

An experiment to determine the moisture content versus time curve, and hence the field capacity, was carried out at 6-inch increments to a depth of 5 feet in 4 soils. Plots of soil were soaked with water and a soil-moisture meter employing the neutron-scattering principle was used to determine the moisture content of the soil profile at different times following water entry into the soil. The data were plotted as curves of soil-moisture content (based on volume) versus time in hours following wetting. The general shape of the curves was normal for 2 silt loam soils and 1 sandy loam soil, but certain layers in these 3 soils deviated from the normal pattern. The deviations were explained by extrinsic physical conditions at some distance from these soil layers. Examples of such extrinsic conditions are: (a) a water table, (b) an impermeable layer, (c) layers of different capillary conditions, and (d) layers of different antecedent soil moisture. Extreme variation of moisture content with time was indicated for a clay loam soil for which the field capacity could not be determined. The neutron meter used was found to be an excellent device for study of field capacity. (Modified from author's abstract.)

Campbell, R. B., 1952, Freezing point of water in puddled and unpuddled soils at different soil moisture tension values: Soil Sci., v. 73, p. 221-229.

The purpose of this study was to indicate the effect of textural and structural variations on the freezing point at several soil-moisture tension levels. Several freezing treatments were used and each soil sample was frozen, thawed, and refrozen five times to investigate the decrease in the freezing point observed between successive freezings.

Three soil types representing the textural range for mineral soils were selected. Each soil was air-dried, passed through a 2-millimeter roundhole sieve, and subdivided into samples. Half of the soil samples were brought approximately to field capacity by the addition of water.

To obtain a high degree of puddling, these soils were stirred thoroughly with a rod and saturated with water, and then transferred to porous membranes. The remaining subsamples of unpuddled soil were poured on porous membranes and were saturated by applying water to the upper surface of the soil. Pressures of 1/3, 1, 5, and 15 atmospheres were applied to separate membranes. The samples were subjected then to a normal freeze, a deep freeze, and an adiabatic freeze. In the adiabatic freeze, the temperature difference between the sample and its surroundings was kept small to minimize heat loss from the sample during freezing.

Osmotic pressures were calculated from the electrical conductivity of the soil. The conversion of freezing-point depression to soil-moisture tension was made by subtracting the values of osmotic pressure.

Puddled soils were found to have smaller values of freezing-point depression than unpuddled soils at the same soil-moisture tension. For the same soil-moisture tension, the percent water retained in the freezing-point procedure was in general greater than in the porous-membrane procedure. Average observed values of freezing-point depression for a soil frozen adiabatically were not significantly different from values obtained in soil frozen in the normal manner.

Larger decreases in the freezing-point depression were observed between the first and second successive freezings than between any pair of successive freezing values. Between the first and second freezings, soils that were given the deep-freeze treatment showed larger decreases in the freezing-point depression than did soils frozen in the normal procedure. Half of the successive freezing curves showed a gradual decrease in the freezing-point depression after the initial freeze, whereas the rest showed no change or a slight increase in freezing-point depression as a result of additional freezings. The decrease in the freezing-point depressions after the initial freezing were greater at high soil-moisture tension levels than at low tension levels. The decrease in the freezing-point depression was attributed to a decrease in the mechanical resistance of the soil to ice crystal formation.

Smaller values of freezing-point depression were observed in clay-loam soil than in loam or fine sandy loam soils at the same moisture tension.

Cannell, G. H., 1955, Freezing-point depressions in sands, soils, and synthetic soil materials: Washington State Coll. Ph.D. dissert.

The freezing-point method, employing a small thermister as the sensing element, was used to study certain factors believed to affect the freezing-point depression of soil moisture. The factors investigated were particle size, structure, and undercooling temperature. The materials used were homogeneous and consisted of quartz sand, synthetic soil, and IBMA (the colpolymer of isobulylene and the half ammonium salt-half amide of malic acid) treated Palouse silt loam.

Freezing-point depression measurements for Palouse silt loam aggregates indicated that the freezing-point depression increased with decreasing aggregate size. Alternate freezing and thawing on the aggregates of various size indicated that the freezing-point depression decreased in the second and third freezings for all sizes of aggregate.

The synthetic soil used was made from quartz sand and illite clay. Some of the synthetic soil was formed into aggregates of 1 to 2 millimeters and treated with various rates of IBMA. Samples containing the highest rate of IBMA were stable in water, whereas aggregates with smaller rates disintegrated into primary particles when placed in water. The freezing-point depression of the IBMA-treated aggregates decreased with decreasing rates. Freezing-point depressions for different sand groups decreased as the particle size increased, but the freezing-point depression did not change under alternate freezing and thawing.

To determine the effect of undercooling temperature upon the freezing-point depression, the bath temperatures were changed from approximately 0.30 to -2.60°C. Different undercooling temperatures (the absolute difference between the supercooling temperature, normally the bath temperature, and the freezing-point depression) had little effect upon the freezing-point depression values for sands on Palouse silt loam aggregates. However, the freezing-point depression for unaggregated synthetic soils was increased as undercooling temperatures increased.

For freezing-point data to have practical meaning, the author suggested that correction of measurements to some standard was required. Pressure-membrane data were used as a standard for comparing freezing-point data for synthetic unaggregated soil. A correction was made to reconcile freezing-point depression measurements on synthetic soil with pressure-membrane measurements. The author concluded that use of the Schofield equation alone was not adequate to convert freezing-point depression values for salt-free systems to moisture tension as obtained by the pressure-membrane or porous-plate technique (see index). For synthetic soil, two types of corrections were involved. The first was for ice formed in the measuring process and depended upon the undercooling temperature. The second was of unknown cause, but seemed to be needed with soil materials with the synthetic soil. This correction was a linear function of the freezing-point depression. (Modified from author's abstract.)

Cardwell, W. T., and Parsons, R. L., 1949, Gravity drainage theory: Am. Inst. Mining, Metall. Petroleum Engineers Trans., v. 179, p. 199-215.

This paper presented a theory for estimating mathematically the rate of gravity drainage of a liquid from a sand column. The typical equilibrium drainage curve was discussed, by relating the curve to the lower end of the column (100 percent moisture content), to the middle part of the column (transitional region of gradually decreasing moisture content), and the upper part of column (practically constant moisture content). The author considered any other saturation distribution in the column and derived equations employing empirical relationships for permeability, saturation, and height of the top of the saturated region when equilibrium is reached. By these equations, the demarcator height (the height of the boundary between the upper unsaturated region and lower completely saturated region) for different drainage times was calculated. A plot of saturation in a column at different times as related to height of saturation (neglecting capillary-pressure gradients arising from gradients in the saturation) was also shown. The percentage of original liquid recovered by drainage from the column was then calculated and was found to compare closely with previously published experimental data on drainage.

Carlton, P. F., Belcher, D. J., Cuykendall, T. R., and Sack, H. S., 1953, Modifications and tests of radioactive probes for measuring soil moisture and density: Civil Aeronautics Adm. Tech. Devel. Rept., no. 194, p. 1-12.

The purpose of this report was to describe the probe-type nuclear meter for measuring soil moisture and density.

The soil-moisture probe consisted of a thin cylindrical brass casing 1 inch in diameter and slightly more than 6 inches long. Inside the casing was a radium D-beryllium neutron source and a slow-neutron detector, a thin-walled Geiger-Mueller counter tube surrounded by silver foil 5 to 10 millimeters thick. The counter tube was partially shielded against direct gamma radiation from the source by means of a 1-inch lead plug. To check radioactive decay of the source, a 10-gallon container of water was used as a reference standard. The time interval selected for counting was 3 minutes. A moisture-calibration curve (the ratio of count in soil to the count in the standard plotted against moisture content) was made.

The density probe employed in this study was similar to the moisture probe, having the same diameter and being only slightly longer. The increased length was necessary in order to obtain the required distance between the source and the gamma detector. The entire probe was encased in a thin aluminum shell, and a radioactive source consisting of cobalt-60 was used. A 10-gallon container of concrete was used as a density standard. As in the moisture-content measurements, counting periods of 3 minutes were used in both the density standard and in the soil. The reading in soil divided by the readings in the standard gave ratios which were plotted as a calibration curve. This was used to obtain the density of the soil. Sometimes minute gaps between the wall of the access tube and the test hole were found to cause erratic test results.

A summary of test results indicated:

(1) The average precision of the moisture and density meters was ± 0.8 pound of water and ± 3.0 pounds per cubic foot of soil respectively.

(2) The difficulties experienced in driving the access tubes in dense soils could be overcome partially by placing the 1-inch (wall thickness 1/32-inch) stainless steel access tubes in a bore hole filled with a soil slurry.

(3) Additional development should include improvement of accuracy of density probe, design of meter applicable to measurement of thin layers, and improvement of the counter tube.

Carman, P. C., 1953, Properties of capillary-held liquids: Jour. Phys. Chemistry, v. 57, p. 56-64.

Capillary condensation arises from the action of surface tension at a curved meniscus. If it occurs in microporous adsorbents, no sharp distinction between capillary condensate and capillary-held liquid in a microporous system exists (moisture in a damp bed of sand). The tendency is to treat these separately, partly because their study required different techniques, but they should be treated as a whole. Macroporous systems afford a clearer view of the mechanism by which capillary liquid is held in a pore-space; microporous systems indicate more clearly how properties alter from those of bulk liquid as a result of increasing curvature of the meniscus (decreasing pore radius). This paper reviewed the brief data on the physical properties of capillary condensates. The author's comments follow in brief:

Capillary condensation takes place because, at a curved meniscus, surface tension reduces the equilibrium vapor pressure p below that of the vapor pressure of the bulk liquid p_o in accord with the Kelvin equation:

$$\ln \frac{p_o}{p} = \frac{2\sigma M}{\rho a R T}$$

where

σ = surface tension,

ρ = density,

M = molecular weight,

T = absolute temperature,

R = gas constant,

and a is the harmonic mean of the two principal radii of curvature, a_1 and a_2, i.e.,

$$\frac{2}{a} = \frac{1}{a_1} = \frac{1}{a_2}$$

This applies in general for any curved liquid-vapor interface, with the signs of a_1 and a_2 positive if the surface is convex to the liquid and negative if concave to the liquid. In a circular capillary, $a = a_1 = a_2 = r\cos\theta$ (where θ is the contact angle), thereby relating the relative pressure p/p_o, to capillary radius. Experimental results seem to indicate that film formation and capillary condensation take place simultaneously in micropore systems, and to some extent, influence one another. The quantity of adsorbate per unit weight of adsorbent is measured. By assuming for convenience that the adsorbate has the same density as bulk liquid and by expressing it as a volume, v, per unit weight of adsorbent, the amount of filled total pore volume can easily be assessed. The author has indicated this in figure form.

Soil physicists have done considerable work on capillary-held water in macropores. This work is based upon the fundamental fact that, owing to the action of surface tension, a curved surface causes a change in the state of tension or of compression of the liquid below that surface. Thus, in a capillary wetted by the liquid, a curved meniscus which is concave on the vapor side is formed. This produces a state of tension in the capillary-held liquid, compared to bulk liquid. Because a tensile stress has the dimensions of a pressure, this is commonly called a pressure deficiency or suction pressure, P, and its relationship to the curvature of the surface is given by the thermodynamic relationship $P = \frac{2\sigma}{a}$ where a is the harmonic mean radius of the curvature of the surface. If a liquid is placed under tension or compression its vapor pressure is changed according to

$$\frac{po}{p} = \frac{MP}{\rho RT}.$$

From the Kelvin equation above, it is evident that for capillary-held liquids, lowering of vapor pressure may be regarded as a secondary property resulting from the state of tension induced by a curved meniscus.

It is more natural then to plot the proportion of capillary-held liquid in a porous solid against P rather than against p/po. With macropore systems, this has the further advantage than P and not p/po is a quantity that can be measured experimentally. The most convenient way to express concentration of liquid is degree of saturation, S (filled pore volume per unit of total proe volume) assuming capillary-held liquid has the density of normal liquid.

Because pore space is completely interconnected, in a bundle of partly filled interconnected capillaries, the capillary-held fluid will redistribute itself by distillation through the empty part of pore space and by flow through the filled part. Whether liquid is removed by drainage or evaporation, the remaining part readjusts itself by either or both mechanisms until a new, uniform curvature of meniscus is attained (at all points accompanied by a corresponding uniformity of P and p/po). The author then discussed and illustrated by sketch the evolution of the growth of annular rings retained at points of contact.

The influence of capillary radius on density, surface tension, freezing point, and heat of condensation of capillary condensates was considered. Recent work on the flow of capillary condensates was summarized and hysteresis (difference on data obtained by drainage and imbibition) was discussed in detail.

Chatenever, Alfred, 1952, Visual examinations of fluid behavior in porous media - Pt. 1: Jour. Petroleum Technology, v. 4, no. 6, p. 149-156.

This exploratory study was made to determine the possibility of a visual approach into microscopic mechanisms of fluid behavior in porous media. Appropriate apparatus and techniques were developed so that microscopic phenomena could be recorded on color movie film and also be observed visually. The observation flow cells in which the fluid behavior studies were made were essentially single-layered matrices of spheres between lucite or glass plates. The fluids used were water and a filtered crude oil.

Two flow regimes were observed during the flow of the immiscible liquids, channel flow and slug flow. In channel flow, flow was through stable networks of interconnecting channels; in slug flow, part of the flow took place in the form of slugs. Under certain conditions, flood-front patterns were found to be different depending upon which liquids were the displacing and displaced phases and not depending upon whether the matrix was water-wet glass or oil-wet lucite. (Modified from author's abstract.)

Childs, E. C., 1945, The water table, equipotentials, and streamlines in drained land: Soil Sci., v. 59, no. 4, p. 313-327.

As a part of this report the author discussed and defined the capillary fringe.

Use is made of the soil-moisture characteristic, graph of moisture content plotted against hydrostatic pressure, to obtain pore-size distribution. A finite range of pressures less than 0 for which the soil is practically saturated exists, but with lower pressures, the pores are appreciably emptied. The lower limiting pressure, p_c, for saturation, is more or less sharply defined, depending on the soil. In a vertical soil column in which the hydrostatic pressure decreases as height increases, starting with some positive value at the bottom, the height at which the pressure is 0 will be the water table. The soil will be essentially saturated to some greater height, at which the upper limiting pressure is p_c, and thereafter the moisture content will decrease rapidly. The zone between the water table and the upper limit of saturation is known as the capillary fringe.

Childs, E. C. and George, N. C., 1948, Soil geometry and soil-water equilibria: Faraday Soc. Discussions, no. 3, p. 78-85.

The authors discussed pore-size distribution and its determination in detail. The following condensed comments by the author are pertinent to specific yield.

The total force acting on a volume element of water in the void space in soil may be made up of four components: gravitational attraction, hydrostatic-pressure difference, osmotic-pressure difference, and adhesion of water to the solid surfaces. The potential at any point will then be the scalar sum of these four components. The water is in hydrostatic equilibrium when the total force is everywhere zero (where the total potential is everywhere the same, notwithstanding that single components may vary widely from point to point).

Gravity is the work done in lifting a unit volume of water (density ρ) to height (h) above an arbitrary datum level; the potential then is given by $g\rho h$.

Hydrostatic pressure is the unequal pressure across an air-water interface. If S is the surface tension and the air-water interface in which it acts has radii of curvature r_1 and r_2 the pressure on the water side is less than that in air by the amount

$$\Delta p = S \left(\frac{1}{r_1} + \frac{1}{r_2} \right)$$

(where r_1 and r_2 are regarded as positive for curvatures which are concave to the air). It is convenient to take the atmospheric pressure as datum: then the pressure P just inside an interface between soil water and air which is continuous with the outside air is given by

$$P = S \left(\frac{1}{r_1} + \frac{1}{r_2} \right)$$

which is called pressure deficiency or suction pressure.

The osmotic pressure p at any point corresponds to a hydraulic potential contribution -p, since it must be considered as a hydrostatic-pressure deficiency.

Little is known of the adhesion potential which is due to attraction between solid and water dipolar molecules. Because the force is an attraction, the potential contribution will be negative and will be called $-\pi$.

The total potential ϕ is then the sum

$$\phi = g\rho h + (P - p - \pi)$$

The total potential may be measured by allowing the soil system to come to equilibrium with a system of known potential components. Such a system may commonly be a manometer, the only directly measurable potential components being gravitational potential and hydrostatic pressure. Equilibrium may be attained either by adjusting the manometer column or by altering the soil air-pressure, as in the pressure-plate method. The total potential being known, the sum $(P-p-\pi)$ is clearly known and is called ψ. The above equation may be rewritten:

$$P = g\rho h + \psi$$

where ψ is referred to as capillary potential or pressure deficiency. The capillary potential components, and therefore ψ, will determine the moisture content of a given soil. The void space may be considered as a series of cells connected to others. For a given pressure deficiency a cell will be empty of water if its largest interconnecting channel is large enough to accomodate an air-water interface of sufficiently sharp curvature to maintain the pressure deficiency. As the suction is increased, cells will empty in succession, giving a suction-moisture content curve characteristic of the soil; such curves are called moisture characteristics. With nonshrinking, light, sandy soils, at all but the lowest moisture contents, the moisture characteristic may indicate the pore-size distribution in the soil.

The soil property which is most obviously governed by pore-size distribution is the permeability to fluids. However, the experimental determination of permeability at different moisture contents is beset with difficulties. The hydraulic potential gradient also affects the moisture content. Determinations of both potential gradient and moisture content are not straightforward. However, if a sufficiently long column is used the moisture content and pressure deficiency are sensibly constant over a considerable length. The potential gradient in a vertical column is therefore the gravitational gradient only and is known without measurement.

The computation of permeability of a soil at any desired moisture content, at a desired effective porosity, was made possible by the author's theoretical approach. In this total, permeability is given by

$$K \propto \sum_{\sigma = o}^{R} \sum_{\rho = o}^{R} f(\sigma)\ dr\ (\rho)\ dr \cdot \sigma^{2}$$

where two similar soil columns of unit cross section are brought together to constitute a continuous column. The total area devoted to the sequence of a pore size σ followed by one of size ρ will be proportional to $f(\sigma)dr \cdot f(\rho)dr$. The final factor in the above equation changes to ρ^2 when $\rho < \sigma$. The summation is carried out numerically from the moisture characteristic. By stopping at a chosen value R one can compute the permeability at any desired moisture content.

The authors then showed by graph form the comparison between computed and observed permeability as they increase with increasing moisture content. Some evidence suggested that at a given moisture content the permeability is greater when the content is decreasing than when it is increasing.

The definition of another moisture constant was clarified by this discussion. The observed facts indicate that water moves down the soil profile relatively rapidly during irrigation and for a few days thereafter, the added water redistributing itself to wet a certain thickness of materials to a certain moisture content. Thereafter, the water moves relatively slowly. No suggestion of the attainment of equilibrium is reached but only a rapid decrease in movement is noted after the initial free-moving stage. The moisture content at the assessed "end" of this initial stage is the field capacity. Redistributions of this kind are reminiscent of diffusion. The authors showed after experimentation with dry sands that diffusion was slow up to moisture contents of about 25 percent. This justifies the neglect of the gravity term except where the diffusion coefficient is high.

Cole, J. S., and Mathews, O. R., 1954, Soil-moisture studies of some Great Plains soils, Pt. 1., Field capacity and "minimum point" as related to the moisture equivalent: Soil Sci. Soc. America Proc., v. 18, no. 3, p. 247-252.

Field determinations of soil moisture were made to depths of 6 feet or less over a 30-year period at 31 stations in the Great Plains area. The field capacity was higher than the moisture equivalent in the sandy soils and lower than the moisture equivalent in the finer-textured soils. In the uppermost foot of the medium-textured soils, the moisture equivalent and field capacity generally were about equal. A slight but persistent reduction in the difference between field capacity and moisture equivalent occurred with depth. This was attributed to the depth and the manner of wetting.

Coley, F. H., Marsden, S. S., and Calhoun, J. C., 1955, A study of the effect of wettability on the behavior of fluids in synthetic porous media: Pennsylvania State Univ. Mining Industry Expt. Sta. Bull., 68, p. 35-46.

A group of porous pyrex glass cylinders made of fritted glass of relatively uniform pore characteristics and approximately 3.7 centimeters in length and 2.1 centimeters in diameter were divided into sub-groups which were treated with chemical solutions in order to produce a scale of oil wettability. To characterize the performance of these cores, the following experimental data was obtained: (1) porosity, (2) capillary-pressure measurement of drainage, reverse drainage, and imbibition processes, (3) relative-permeability determinations, and (4) water flooding.

The results of the experimental work indicated that wettability was both one of the primary factors to consider in achieving consistent correlations between experimental procedures, and also one which had to be certainly known before proper utilization of the data was made.

Different variations were proposed as possible scales of wettability. Among these were: (1) the ratio of drainage- and imbibition-displacement pressures, (2) the ratio of residual saturation to wetting and non-wetting fluids, and (3) the spread between the saturation end-points on dynamic relative-permeability tests. (Modified from author's conclusions.)

Collis, G. N., 1952, A note on the pressure-plate membrane apparatus: Soil Sci., v. 74, p. 315-322.

Attention was drawn to the requirements for obtaining the true equilibrium-moisture content of a soil under experimental conditions and for distinguishing this equilibrium-moisture content from the equilibrium-moisture content as determined when water flow from a pressure plate-membrane has apparently ceased for any applied pressure.

Results from this study indicated that the moisture content of a soil in a pressure-plate apparatus decreased indefinitely when a dry gas was used and gaseous diffusion through the apparatus was continuous. The author pointed out that for any soil sample there was only one water-vapor pressure of the applied gas which could be in thermodynamic equilibrium with the sample. This vapor pressure depended on the osmotic pressure of the soil solution and the pressure of the gas applied.

Generally the moisture content of a soil sample is determined when, at a given pressure, the outflow of liquid water through the membrane has become negligible. The author suggested that this condition be called a "time equilibrium", which will change with the length of time of the pressure application. He further suggested that the condition under which the the humidity of the gas leaving the apparatus is the same as that entering be called "hydraulic equilibrium". Except for multisoil apparatus, this equilibrium will be thermodynamically correct and the relative humidity involved will be in equilibrium with both applied pressure and osmotic pressure. True hydrostatic equilibrium can be reached only by use of an ideal membrane that is permeable to water but completely impermeable to gases.

Colman, E. A., 1946, A laboratory study of lysimeter drainage under controlled soil-moisture tension: Soil Sci., v. 62, p. 365.

The purpose of this paper was to present the results of a laboratory study in which long soil columns were drained under several moisture tensions and in which the drained soil-moisture condition was related to the saturation and field-capacity values of the soil.

The soil used in the column-drainage study was a clay loam derived from metamorphosed diorite. Its moisture equivalent and field capacity were 19 percent. The soil was packed air-dry in a brass tube that was 7 feet long and had an inside diameter of 6 inches. The tube was filled by slowly adding the soil while a long, sharpened, steel rod was struck repeatedly into the soil in the tube. The agitation and tamping served to settle the soil and to minimize textural and structural layering. Three-fourth-inch holes were drilled through the tube at 6-inch vertical intervals to provide access to the soil for the placement of tensiometers and to obtain moisture samples.

The soil column was filled twice. After the first filling, the soil was irrigated and drained under moisture tensions of 0 and 50 centimeters of water. After being emptied, refilled and again irrigated, it was drained under tensions of 55 and 160 centimeters of water. Zero tension, corresponding to the usual gravity drainage of enclosed lysimeters, was effected by resting the soil on a perforated plate and collecting the drainage water that dripped from it. Drainage tensions of 50, 55, and 160 centimeters of water were maintained on the lower surface of a porous fired-clay plate held in contact with the base of the column. A "hanging water column" for the 50- and 55-centimeter tensions and a vacuum pump for the 160-centimeter tension were used. Moisture- and pressure-potential distributions were determined throughout the soil columns before, during, and after drainage. Measurements were made of rates and duration of drainage as well as of water outflow under each drainage tension.

The selection of drainage tensions of 0,50, 55 and 160 centimeters of water was based on observed lysimeter soil characteristics. Zero tension corresponded to the minimum-pressure potential observed during seepage at the bottom of large confined lysimeters in which a 3-year record of soil-moisture conditions were available.

The 50- and 55-centimeter drainage tensions selected were considered to be in the range of tensions observed in the field at the 6-foot depth in an unconfined lysimeter when the soil at that depth was approximately at its field capacity. The 160-centimeter drainage tension was selected in the laboratory because it was this tension that was required to drain a 100-gram block of soil to a constant moisture of 19 percent on a porous fired-clay plate. The difference of 110 centimeters of tension between the field and laboratory determinations may possibly be attributed to a structure in the unconfined, lysimeter soil which was not present in the laboratory-packed samples.

The rate of seepage was increased by the application of tension at the base of the column. The authors thought that the permeability of the clay plates probably exerted a strong influence upon the rate and duration of drainage. They suggested that this problem might be solved by using plates that had a large absorbing surface or a high porosity.

The proper drainage tension for the soil studied seemed to be about 125 centimeters of water, which is equivalent to 19 percent moisture content, the field capacity of the soil. The pressure potential at the field capacity or moisture equivalent depended upon the moisture content of these values. The pressure potential decreased as the field capacity or moisture equivalent increased. A sandy soil would therefore require a lower drainage tension than a finer-textured soil.

Colman, E. A., 1947, A laboratory procedure for determining the field capacity of soils: Soil Sci., v. 63, p. 277-283.

This paper described a laboratory procedure for the determination of field capacity of drainage of a porous ceramic cell under a constant moisture tension, and related the field capacity values of a number of soils to the equilibrium-moisture contents obtained by this method. Field-capacity values were determined by two methods. In the first method, repeated soil-moisture samples were obtained on small field plots during 5 rainy seasons and in the second method, field plots were irrigated and the moisture determinations were obtained after drainage of the irrigation water had ceased.

The method for determining the 1/3-atmosphere moisture percentage was as follows. Sieved soils were poured into 1-inch lengths of seamless brass tubing, 1 inch in diameter, resting on a ceramic tension cell, and then tap the soil samples lightly. The soils were saturated from below and then were drained at 1/3-atmosphere tension. Three hours drainage was found to be adequate to reach moisture equilibrium for all of the soils studied. If considerable coarse material was sieved from the soil sample, the laboratory determinations of field capacity was corrected for the weight of the coarse material. Approximately 120 samples were analyzed in the laboratory in this manner.

A number of soils were saturated and then were drained under moisture tensions ranging from 2 to 50 centimeters of mercury. The moisture tension corresponding to field capacity was not constant but was increased as the field capacity increased. Whether this relationship would also be found in soils in place, or whether it is a result of the granulating treatment the soils received in sampling and in the laboratory was not determined. However, the results indicated that no single moisture tension applied will bring all soils to their field capacity under laboratory conditions.

A consistent relationship was found between the field capacity and the moisture retained against 25 centimeters of mercury tension. A curve based on the experimental data showed field capacity and moisture content to be at 1/3-atmosphere tension equal at about 25 percent. At lower values the field capacity was greater than the moisture content at 1/3-atmosphere tension, and at higher values it was less. This relationship was determined using young or immaturely developed, free-draining soils in which the water table was so far below the layer under consideration that it exerted no influence upon the rate of completeness of drainage of that layer. The curve would not be expected to hold for soils in which the water table was only a few feet below the surface, or those in which drainage was impeded by a layer of low permeability.

Corey, A. T., 1957, Measurement of water and air permeability in unsaturated soil: Soil Sci. Soc. America Proc., v. 21, no. 1, p. 7-10.

This paper explained a controlled-pressure, steady-state method developed for making measurements on oil-producing rocks, which may also be used for determining water and air permeabilities in unsaturated soils. The method employed the simultaneous flow of air and water under the same pressure gradient to maintain a uniform tension and a uniform saturation within the sample during permeability measurements. The saturation was reduced in increments by reducing the water pressure with respect to the air pressure. The techniques described probably are not adequate for the small permeabilities that exist at saturations smaller than field capacity. Neither are they likely to be successful on very fine soils that have high air-entry values or high bubble pressures. The chief limitation here is the comparatively small air-entry values of the capillary barriers or semi-permeable membranes. Barriers that have higher air-entry pressures are available but their use makes the process much more difficult.

Results of permeability measurements under unsaturated conditions on two relatively undistrubed samples of a sandy soil were presented. The interrelationship between permeabilities to gas and liquid as a function of saturation was discussed, and a method of calculating the water permeabilities under unsaturated conditions from the more easily measured air permeabilities was suggested. The magnitude of gas slippage (or slip flow where no wetting occurs) in a dry soil sample was determined by measuring the air permeability at several mean pressures and extrapolating the air permeability-mean pressure function to an infinite pressure. The permeability to water of the unsaturated soil was only about half the permeability to air of the dry soil even when the air permeability was corrected for slippage. (Modified from author's abstract)

Crank, J. and Henry, M. E., 1949, Effect of a variable diffusion coefficient on the concentration - distance relationship in the nonsteady state: Faraday Soc. Trans., v. 45, p. 1119-1130.

The authors calculated concentration-distance curves for diffusion coefficients which depend on concentration in three different ways, and which may have a range of as much as 200-fold. The results were presented graphically in a form convenient for interpolation.

A method was suggested by which an approximate concentration-distance curve could be obtained, with relatively little labor, for any diffusion coefficient which depends on concentration.

The importance of the results in relation to the proposed method of measuring variable diffusion coefficients was discussed. Some insight into the nature of the sharp advancing boundary observed in many systems was provided, and the significance of the mean diffusion coefficient, deduced from the rate of advance of such a boundary was considered.

Crank, J., and Henry, M. E., 1949, The effect of a variable diffusion coefficient on the rates of absorption and desorption: Faraday Soc. Trans., v. 45, p. 636-650.

In order to determine how the form of the diffusion coefficient for liquid influences the relative behaviour of the absorption and desorption-time curves, numerical solutions of the diffusion equation in one dimension were obtained for a number of diffusion coefficients varying in different ways with the concentration of the diffusing substance. The results indicated that, when the diffusion coefficient increased uniformly with increasing concentration, absorption was quicker than desorption throughout, but that, when the diffusion coefficient decreased with increasing concentration, the reverse was true. The conditions for which the diffusion coefficient first increases, passes through a maximum value, and finally decreases as the concentration increases, and when the absorption and desorption-time curves may cross each other were examined.

Da Costa Botelho, J. V., and Alves, J. A., 1942, The determination of the moisture equivalent by the silt-suction method and the shifting pF curve of stored soil samples: Jour. Agr. Sci., v. 32, p. 294-297.

The reliability of the silt-suction method for estimating the moisture equivalent of soils was discussed. In this procedure a fine layer of silt, 1 millimeter thick, was poured on the filter paper and a soil sample about 6 millimeters thick was placed on the silt. The sample was soaked with water for 24 hours and a suction equivalent to a pF (logarithm of capillary potential) of 2.9 was applied for 2 hours. The moisture contents at pF 2.9 were found to be considerably lower than the moisture contents obtained by the moisture-equivalent method.

Davis, R. O. E., and Adams, J. R., 1927, Methods for physical examination of soils: Internat. Cong. Soil Sci. Proc., v. 1, p. 434-442.

A discussion of the methods used by the U.S. Bureau of Soils in determining certain physical characteristics of soils was given. A number of moisture-equivalent determinations were made to determine if the moisture equivalent of soils in the undistrubed condition was different from that of samples passed through a 2 millimeter sieve. In most instances the soils in the field condition had slightly lower moisture-equivalent values than the disturbed samples. The differences were so slight, however, that changes in the ordinary method of determining the moisture equivalent on the crushed sample of soil were not considered necessary.

Day, P. R., and Luthin, J. N., 1956, A numerical solution of the differential equation of flow for a vertical drainage problem: Soil Sci. Soc. America Proc., v. 20, no. 4, p. 443-447.

The experiments described in this article were undertaken for the purpose of testing the application of Richards equation, $\frac{dc}{dt} = \Delta \cdot K\Delta\phi$ to a drainage problem. In this equation c = volume of water per unit volume of soil, t = time, K = capillary conductivity, and ϕ = hydraulic head. The problem was to predict from theory the water content and pressure head at different depths and times during drainage, to calculate the rate of discharge at selected times, and to test these predictions by experimentation. In order to do this, the water content of a soil as a function of soil-moisture tension under equilibrium conditions and the capillary conductivity of a soil as a function of soil-moisture tension under steady conditions of flow needed to be known before the theoretical predictions could be made.

A column of Oso Flaco fine sand, 87.1 centimeters high, contained in a 3-inch-diameter brass tube was used in the experimental work. Tensiometers were located at heights of 5.0, 19.9, 35.5, 50.5, 66.2, and 82.5 centimeters from the bottom of the soil column. Capillary conductivity values at different tensions were determined by supplying water dropwise to the soil surface at various sustained rates, starting with the maximum rate and diminishing in successive steps. Flow was continued at each rate until constant readings were obtained on the tensiometers. Capillary-conductivity values were calculated by dividing the rate of entry in cubic centimeters per square centimeter per minute by the hydraulic gradient in different parts of the column. Each value obtained in this manner was recorded as a function of the moisture tension at the corresponding level in the column. A desorption curve for the soil column was determined by sampling for water content at several depths. In the drainage experiment, measurements of pressure head and outflow volume were made at selected times.

Because the problem was a special case of unidirectional flow parallel to the Z-axis, the vertical axis, the Richard's equation was rewritten in the following simpler form:

$$\frac{dc}{dt} = \frac{d}{dz}\left(K\frac{d\phi}{dz}\right)$$

where z is the vertical elevation of the tensiometer tip above the outflow plane.

Special conditions of the problem consist of streamlines that were vertical at all times during drainage and an outflow plane that constituted a fixed piezometric surface at atmospheric pressure.

The method of solution was an approximate numerical method, employing the determination of pressure as a function of depth for a sequence of time intervals. The distribution of pressure with depth at any time could be determined approximately from the above equation by substituting finite differences for the differential quantities and by applying the conditions expressed by this equation simultaneously for all levels in the column. The procedure required an initial assumption of the pressure distribution, followed by a progressive improvement of the assumption by repeated calculation. The time intervals could not be chosen arbitrarily because their determination emerged as part of the solution.

The results of this study indicated that the general trend of the drainage process can be predicted from a knowledge of the desorption and capillary-conductivity characteristics of the soil. Accurate predictions can be expected only when these characteristics can be controlled and measured precisely. The greatest handicap was the complicated procedure for solving the differential equation of flow. The authors considered the equation a fundamental relationship in drainage theory and stated that it might become exceedingly useful in drainage investigations when improved experimental techniques and better methods of calculation become available.

De Golyer, E. L., ed.,, 1940, Elements of the petroleum industry:
New York, Am. Inst. Mining Metall. Engineers, p. 191.

An apparatus involving essentially the same technique as the Washburn-Bunting porosimeter was that developed by the U. S. Bureau of Mines. This consisted of a pressure bomb, in which the test specimen was placed, spring gauge and dead-weight gauge for measuring pressure, and a burette into which the gas was expanded and measured at atmospheric pressure.

The volume of the bomb was determined by introducing compressed air, by observing the pressure, and then by releasing the air into the burette and measuring its volume at atmospheric pressure. The procedure was repeated with the core sample in the bomb. The compressed air filled not only the part of the bomb not occupied by the specimen but also the voids in the specimen.

The difference between the two volumes was the volume of the sand grains. The bulk volume of the sample was then determined by a phcnometer using mercury. Porosity was then determined by the following equation,

$$P = \frac{V_B - V_G}{V_B}$$

where

P = porosity,

V_B = bulk volume of sample,

V_G = volume of the sand grains.

Deryagin, B. V., and Zorin, Z. M., 1954, Optical investigation of adsorption and surface condensation of vapors at near the saturation pressure: Doklady Akademiia Nauk, SSSR, v. 98, p. 93-96.

The multimol adsorption of vapor on a smooth surface was investigated to clarify the nature of phase changes between a gas and the liquid film adsorbed. The thickness of the layers of several n-alcohols, water, CCl_4, C_6H_6, and some other polar and nonpolar substances adsorbed upon an optically polished glass surface was measured by the elliptic polarization of light, reflected under an oblique angle, and analyzed with a polorization microscope. The accuracy of the process is not claimed to exceed ± 5A., but the method permits the observation of the nature of condensation and of the adsorbed film, and thus controls its uniformity. A multimol adsorption was observed at relative vapor concentration of 95-97 percent; at lower concentration, the adsorption was uni-and bimol. The adsorbed layer must be considered as a special "boundary phase". Details of apparatus construction and of the glass surface preparation are given. (Abstracted by W. M. Steinberg for Chem. Abs., v. 49, 8656 g, 1955))

Deryagin. B. V. and Zorin, Z. M., 1955, Study of the surface condensation and adsorption of vapors near saturation by the optical micropolarization method: Zhur. Fizicheskoĭ Khimii, v. 29, p. 1755-1770.

The thickness h of the adsorbed film on polished crown glass and flint glass at relative vapor pressures p/ps above 0.93 was determined optically. Presence of H (hydrogen) in the vapor did not affect h. For H_2O, alcohols, etc., h gradually increased with p/ps; e.g., for EtOH, h was 10, 30, and 62 A. at p/ps of 0.955, 0.985, and 1.00 respectively, and h at p/ps 1 was 75, 50, 60, 100, 110, 35, 40, 30, 20 and 40 A. for H_2O, PrOH, BuOH, n-hexanol, n-heptanol, n-octanol, n-nonanol, n-deconal, $PhNO_2$, and hexanoic acid, respectively. For nonpolar substances, h was often small up to, e.g., p/ps = 0.97 and rose of further increase of p/ps so rapidly that the value of h at p/ps 1 could not be determined; This was particularly true for C_6H_6 and CCl_4, less so for n-C_5H_{12} and n-C_7H_{16}. Since the adsorption isotherm of polar substances intersects the ordinate at p/ps = 1, the transition between the adsorption film and bulk liquid phase must be discontinuous. The polar bulk phase starts as discrete droplets coexisting with the adsorption layer; apparently adsorption film is a separate phase having a characteristic structure. On the contrary, the adsorption film of nonpolar substances does not differ from their bulk phase. The conclusion by Bowden and Throssel (Chem. Abs., v. 47, 9713c) that multimol adsorption does not exist is incorrect. (Abstracted by J. J. Bikerman for Chem. Abs., v. 50, 6131i, 1956.).

Drombrowski, H. S., and Brownell, L. E., 1954, Residual equilibrium saturation of porous media: Indus. and Eng. Chemistry, v. 46, p. 1207-1219.

This article described the different variables and their influence on residual equilibrium saturation of porous media.

Porous media having a wide range in porosity were selected, including glass spheres, quartz sand, glass helices, alluminum cylinders, and nickle saddles. Several liquids of different viscosities, densities, and surface tensions were used and the desaturating driving forces of gravity, centrifugal force, and the pressure gradient of air as a displacing fluid were investigated.

An x-ray technique was developed primarily for analyzing changes in liquid saturation in porous media during progressive desaturation by increased pressure gradients in air. This technique used film to record the transmission of x-rays through the porous beds. Microphotometer recordings of photographic film density together with suitable calibration charts established point saturations.

A method was developed to predict the residual liquid content of unconsolidated porous beds in terms of the permeability and length of bed, the density, surface tension, and contact angle of the liquid, and the forces of gravity, centrifugal force, and pressure gradient. Liquid contents were discussed as the volume percent of the total volume of pore space filled with the liquid.

Eaton, E. M., and Horton, C. R., 1940, Effect of exchange sodium on the moisture equivalent and the wilting coefficient of soils: Jour. Agr. Research, v. 61, no. 6, p. 401-425.

The effects of adsorbed sodium, calcium, potassium, and magnesium on the moisture equivalent values of soils were reported. Two sets of soils were chemically treated for this investigation. The first set, designated as washed calcium and sodium soils, provided information for the comparisons of the effects of calcium and sodium on the moisture equivalent. The second set, designated as calcium, magnesium, potassium, and sodium soils, was used only for supplementary moisture-equivalent measurements where the effect of potassium was of principal interest. The moisture equivalents were determined on 30-gram samples of soil which were centrifuged for 30 minutes at 2,440 rpm.

The time required for the centrifuge to reach full speed was found to affect the moisture-equivalent values. Higher moisture-equivalent values were obtained when the full speed was reached in 15 seconds than when full speed was reached in 4½ minutes. The difference was 1 percent or less for the calcium-treated soils but from 1.9 to 10.6 percent for the sodium-treated soils.

Measurements were made of the distribution of water in samples of the treated calcium and sodium soils after centrifuging. The percentage of water in the calcium soils increased from the inner surface to the outer; in the sodium soils the percentage of water increased from the outer surface to the inner. Sodium soils contained much water and, as the soil was dispersed, particle segregation occured readily in the plastic masses of the fine-textured soils. During centrifuging the large particles tend to migrate towards the periphery, displacing small particles towards the axis of rotation.

The amount of the effect of sodium on the moisture equivalent was closely related to the percentage of clay, exchange capacity, and quantity of adsorbed sodium in the soil. The moisture equivalents for the sodium-treated soils were in many instances two or more times as high as those for the calcium-treated soils. The effect of adsorbed sodium on the moisture equivalent, consequently, was considered as being due to three coincident factors: hydration, dispersion, and the segregation of a relatively impervious layer of clay on the surface of some of the soils. Hydration contributed to the results both because of the tightly-held water and because of the effect hydration may have on impermeability. Dispersion increased the free surfaces and thereby the water retentiveness of the soils.

S-shaped moisture-equivalent curves resulted when calcium and sodium soils were mixed in successive proportions indicating that the relationship between adsorbed sodium and either or both hydration and dispersion is not linear. The moisture equivalents of soils did not seem to be measurably affected by less than 2 millequivalents of adsorbed sodium. The maximum effect on the moisture equivalent was approached when 8.17, 12.26, and 12.74 millequivalents of adsorbed sodium was used for 3 soils. The maximum effect was not attained with 4.9 millequivalents of adsorbed sodium for a fourth soil.

Twelve sodium-treated soils were wetted with a normal solution of calcium chloride, and several were wetted with a normal solution of sodium chloride. The moisture equivalents of the sodium soils wetted with these solutions tended to be equal to, or slightly lower than, the moisture equivalents of calcium soils wetted with distilled water. The adsorbed sodium apparently did not affect the moisture equivalent when samples were wetted with a strong salt solution rather than distilled water.

The moisture equivalents of calcium and magnesium soils were found to be similar to and nearly always slightly less than those of the potassium soils.

Eckis, Rollin, and Gross, P. L. K., 1934, South Coastal Basin investigation - geology and ground-water storage capacity of valley fill: California Div. Water Resources Bull. 45, 273 p.

Specific-yield and storage-capacity studies were made in 35 ground-water basins of the South Coastal Basin. The storage-capacity estimates were for zones 50 feet above and below the water table as of January 1933, respectively.

Part of the study included a detailed investigation of specific yield, porosity, and specific retention. Several hundred samples were collected from surface exposures and shallow test holes, and several thousand samples were obtained from wells. Porosity and water-yielding capacity were determined, and mechanical analyses were made primarily for sand and gravel samples. Mechanical analyses were made to determine particle-size distribution, and the statistical measures, mean grain size, surface factor, uniformity coefficient and effective size; and the standard deviation, ratio deviation, and skewness of the distribution were computed.

Because most of the experimental work involved the determination of specific retention (which must be subtracted from porosity to obtain specific yield), the study of methods for determination of porosity were considered important. The five porosity-analysis methods involved: (1) repacking of unconsolidated sediments in the laboratory by tapping a sample cylinder until the volume reached a minimum, (2) driving a sharpened-edge steel cylinder to obtain a known volume of the sediments, (3) digging a pit, weighing the excavated material, and measuring the volume of the excavation with a uniform sand, (4) determining the volume of a weighed sample of consolidated sediments by displacement in mercury, and (5) estimating porosity from data obtained from particle-size analyses. The method of measuring the volume of an excavation with a uniform sand was concluded to be more accurate possibly than the method of driving a cylinder. The results in graph form indicated also that sediments of similar particle-size distribution had similar porosities. Porosity was related also to ratio deviation as follows:

Porosity = 45.5 percent - 3.35 ratio deviation

Samples analyzed for their water-yielding capacity were primarily sands and gravels, accordingly, the specific retention values were relatively small and varied only a few percent.

Five methods for determination of specific retention were studied. Three cylinders, 4 inches in diameter and 40 inches in length, were packed with carefully sorted sand and saturated with a measured quantity of water. Water was then drained from the sand columns by means of a well outside the cylinder attached through a U-tube to the bottom of the cylinder. Entrapped air was removed by tapping the cylinder and allowing the materials to settle over prolonged periods of time. The water table was lowered slowly over a period of one month so that no capillary columns would be isolated and so that drainage to specific retention values could be accomplished. The water table was allowed to stabilize for 30 to 60 days. A measured amount of water was then removed slowly and a period of 30 to 60 days was allowed for the new water table to reach equilibrium. The difference between the water levels was then used to calculate the specific yield. The columns were allowed to drain for an additional 18 months, the cylinders were split, the the moisture contents were determined for approximately each 1-inch length of the column. The moisture content was plotted for each inch above the water table. Although all tests were carried out in a constant temperature room at 25°C, evaporation effects at depths as great as 27 inches below the top of the column were noted. Less than 10 percent of the water retained in the sample at the end of the 30- to 60-day period drained out in the 18-month period.

Moisture content was determined for samples collected from pits dug to the water table and from auger borings. Samples were obtained, at depths as great as 500 feet, from materials that were drained as the water levels fell in the summer. The moisture retention of the pit samples was considerably greater than that of the materials drained in the laboratory. The moisture retention of the auger samples was considerably less than that of the pit samples but only slightly less than that of the laboratory drainage samples. The authors concluded that some evaporation may have taken place in the relatively shallow auger samples but that probably very little had taken place in the deeper pit samples.

Steel cylinders, 18 inches long and 4 or 8 inches in diameter, were driven into the sediments, removed, and the moisture content then determined. Two sets of samples were collected, the first about 36 to 48 hours after a rain, the the second after an additional 36- to 48-hour period had elapsed.

Cylinders 6 to 8 inches long were driven in duplicate to enclose a volume of natural sediments. They were then saturated by addition of excess water, covered, and allowed to drain in place for a period of 11 or 22 days. Because the retention after 11 days was about the same as it was after 36 hours (for similar materials by gravity drainage of columns when an excess of water was not used), the drainage of excess water was considered to be a very slow process.

An indirect method for estimation of specific retention was developed. The surface factor or index of fineness was determined from mechanical analysis and plotted against moisture retention. A straight-line relationship thus obtained was useful in estimating specific retention as follows:

$$W = 3s$$

$$\text{and specific retention} = (W)(d)(100 - p)$$

where

W = moisture content, percent by dry weight,

s = surface factor,

d = specific gravity,

p = porosity.

The surface factor (s) is proportional to the surface (s) per unit volume (V) of solid material or $s = k\frac{S}{V}$, where k is a proportionally constant. For spherical particles the formula becomes $s = \frac{k4\pi r^2}{4/3\pi r^3} = \frac{k^3}{r}$. The surface factor is then calculated from $s = 0.1\sum\frac{3p}{r}$ if $p_{\overline{r_1 \text{ to } r_2}}$ is the percentage of material in the class interval between the limits r_1 and r_2 and r is the average radius in the interval.

The investigation failed to indicate any detectable dependence of specific retention on porosity.

The authors concluded that an increase in temperature would speed up the process of drainage and that specific retention was thus dependent on temperature. This conclusion was not based on experimental evidence, however, but upon the reasoning that surface tension decreased with an increase in temperature.

Since both specific retention and porosity were shown to vary with particle size, both values were averaged for each maximum 10-percent grade size and plotted on a graph against the 10-percent grade size. Values on the specific retention curve were subtracted from the corresponding values on the porosity curve to obtain the points necessary for drawing a specific-yield curve.

The authors emphasize that weathering of solid rock may increase the specific yield of a formation whereas the weathering of alluvial gravels may cause a decrease in specific yield owing to the disintegration of the large particles into a clayey matrix. Compaction of sediments was noted to reduce specific yield owing to a decrease in the size of the pores. Cementation had essentially the same effect on specific yield as compaction.

Edlefsen, N. E., and Bodman, G. B., 1941, Field measurements of water movement through a silt loam soil: Am. Soc. Agronomy Jour., v. 33, p. 713-731.

The objective of this study was to measure over a period of 842 days the magnitude of the vertical flow across different horizontal planes in undistrubed soils.

The study was made on a plot 16 by 16 feet on silt loam soil. Irrigation water was applied in an amount more than sufficient to saturate the entire soil mass to the water table, which at all times was deeper than 22 feet below the soil surface. The time at which the free-water surface coincided with the surface of the ground was chosen as the zero time for calculations of moisture content or velocity. At zero time the plot was covered with 2 layers of heavy roofing paper. A 6-inch layer of dry soil was applied over the roofing paper and a sheet-iron roof was built over the entire area to keep out rain. An area, 8 by 8 feet, in the middle of the plot was chosen for moisture sampling work. Samples were taken with a standard soil tube at 6-inch intervals to a maximum depth of 9 feet. Mechanical analysis, and determinations of apparent density and moisture equivalent were made on each 6-inch layer.

The results indicated that the downward flow of water under the influence of gravity decreased with time and that a general decrease in the relative wetness at all depths occurred with time for the entire 28 months. The relative wetness of the soil was found to increase, at a given time, with depth from the soil surface.

Ellis, A. J., and Lee, C. H., 1919, Geology and ground waters of the the western part of San Diego County, California: U.S. Geol. Survey Water-Supply Paper 446, p. 121-123.

As a part of the investigation, the authors presented a method for determining the water-retaining capacity of valley fill. Pits were dug to the ground-water level at points selected so as to give differing distances to the water table and differing types of material. Samples of the material were taken during the dry season at intervals of 1 foot, from the surface down to the water table, by driving metal cylinders, 9 inches long and 5-5/8 inches in diameter into the material. The moisture content of each sample was determined and expressed as a percentage of initial volume of the sample.

The percentage of water retained differed throughout the interval sampled and was maximum at the water table and minimum at a distance of about 1 foot below land surface.

The volume of water represented by the annual rise and fall of the water table was computed. For an average annual fluctuation of approximately 3.5 feet, the effective porosity or specific yield (difference between total porosity and water-retaining capacity) ranged from an average of 41 percent for sand to 16 percent for fine sandy loams.

Elrick, D. E., and Tanner, C. B., 1955, Influence of sample pretreatment on soil-moisture retention: Soil Sci. Soc. America Proc., v. 19, no. 3, p. 279-282.

The moisture-release curves (or moisture-tension) of eight soils were computed for samples passed through a 2-millimeter sieve, and for undisturbed soil cores. Eleven different moisture tensions in the range of 0.01 to 15 atmospheres were used in determining the extent to which sieve soils could be used to approximate reliably the moisture retention of undisturbed soils. The release curves were also computed for puddled soils to evaluate the effect of puddling.

Undistrubed cores, retained in brass cylinders 5 centimeters in diameter and 1 centimeter in height, and a bulk sample of soil were obtained at each soil sampling site. The cores were wax-coated in the field. Prior to determining the moisture-release curves in the samples, the wax coats were removed and the soil samples were trimmed to obtain the best possible membrane contact. To prevent the loss of soil from the cores, a single thickness of facial tissue was placed at the bottom of the soil core and a nylon stocking mesh was placed over the tissue and glued to the bottom rim of the brass sample cylinder. Seven cores of each soil were soaked for at least 48 hours and were tested at 11 tensions without rewetting. The air-dried bulk sample was passed through a 2-millimeter round-holed sieve and then quartered to sample size. Triplicate 20-gram sieved samples of dry soil were poured into rubber rings 5 centimeters in diameter on the porous membranes and then wetted by capillarity for 16 to 24 hours before tension was applied. Triplicate 20-gram puddled samples were obtained by triturating sieved soil which had been wetted to about the saturation percentage. The puddled soils were allowed to stand for 16 to 24 hours, after which more water was added and the soils again triturated. The thin paste was then poured into rubber rings on the porous membrane. Soil-moisture tensions of 10, 20, 40, and 80 centimeters of water were obtained on alundum tension plates. Tensions of 160, 325, and 650 centimeters of water were obtained on porous ceramic plates in the pressure-cooker apparatus and tensions of 1.5, 3, 8, and 15 atmospheres were obtained in the pressure-membrane apparatus.

The moisture retention of sieved samples of the medium-textured soils studied was greater than that of the core samples at tensions of less than 0.4 atmospheres. No general relationship existed between values obtained by the two methods in the tension range of 0.4 to 1.0 atmosphere. Sieving decreased moisture retention about 10 percent at tensions greater than 1.0 atmosphere. The authors stated that core samples should be used rather than sieved samples for tests at tensions of less than 1.0 atmosphere to avoid errors greater than 10 percent. Puddled samples retained considerably more water than core samples in the tension range of 0.1 to 2.0 atmospheres. Little difference in water retained was observed at tensions less than 0.1 atmosphere and greater than 5.0 atmospheres. Puddled soils gave a better approximation of undisrupted soils than sieved soils at tensions between 5 and 15 atmospheres.

Gardner, W. H., and Gardner, Willard, 1951, Flow of soil moisture in
the unsaturated state: Soil Sci. Soc. America Proc. (1950),
v. 15, p. 42-50.

When soil is saturated with water the flow can be described by means of the Darcy law where

$$V = -K\Delta\Phi$$

and

V = velocity of flow,

K = Proportionality constant having dimensions of time and representing the permeability of the soil,

$\Delta\Phi$ = Symbol representing the force per unit mass that drives water through the soil.

The success of the Darcy generalization in saturated flow has made it desirable to adopt it to the solution of problems of unsaturated flow. Gardner in 1936 suggested that for unsaturated flow an undetermined function of moisture content, f, be inserted in the Darcy equation; thus $V = -Kf\Delta\Phi$. It was the purpose of this research to discover the nature of this function f by studying the horizontal radial flow of water through soil contained in a semi-cylindrical lamina 1 centimeter deep from a source where water could be furnished under any desired pressure.

Contrary to the simple assumption that a function of moisture content existed that could be multiplied by the permeability constant K of the Darcy equation to make it applicable to movement of water in unsaturated soil, the authors found that experimental data indicated in addition to a pressure gradient due to variation of soil moisture from point to point that there must be a gradient due to the dynamic process itself.

The data did indicate that the permeability decreased rapidly with the moisture content.

Gatewood, J. S., Robinson, T. W., Colby, B. R., Hem, J. D., Halpenny, L. C., 1950, Use of water by bottom-land vegetation in lower Safford Valley, Arizona: U. S. Geol. Survey Water-Supply Paper 1103, p. 81-92.

The authors studied evapotranspiration of phreatophytic vegetation and therefore were interested in specific yield of sediments for a short period of time, usually 12 hours. The terms coefficient of drainage, where the water level was declining, and coefficient of saturation, where the water level was rising, were coined to designate the specific yeild with time as an element. Four methods of determining the coefficient of drainage were discussed: (1) the cylinder method, (2) the tank method, (3) the laboratory method, and (4) miniature pumping tests.

Three large cylinder samples, 42 inches in length and 14 inches in diameter, were obtained from undistrubed materials close to the transpiration wells. A pit was excavated and as the cylinder was driven downward into the bottom of the pit, the pit was deepened outside the cylinder to reduce friction. The material in the bottom 4 inches of the undisturbed cylinder sample was removed. Perforated pipe was inserted through the side of the cylinder in this interval and the space was filled with a fine gravel. The cylinder was made watertight by welding a steel plate to the bottom, and a pipe tee and a stop cock were attached to the outer edge of the cylinder. Glass tubing for observation of water level was then attached vertically to the tee and the cylinders were placed vertically on a recessed shelf in a pit 5 feet deep. The tops of the cylinders were insulated with rock wool and the sides and tops of the cylinders were covered with a layer of soil, leaving visible only the gage glass and stop cock. The pit was covered with a roof to protect it from rain and from rapid or large changes in temperature and relative himidity. The effects of changes in barometric pressure were minimized by using only the readings of water level that were made within a narrow range of barometric pressure. Holes were bored to the top surface in each of the samples and water was added until the water level stood at heights of about 1, 1-1/2, and 1-3/4 feet, respectively, above the bottoms of the 3 samples. Water levels in the gage glasses were observed for a 15-day period and a measured quantity of water was then withdrawn through the stop cock from each of the samples. The resulting decline of water level was observed for about 25 days. Water level in the 3 samples had ceased to decline at the end of 9, 10, and 15 days, respectively. The volume of water withdrawn from each sample was divided by the volume of material unwatered and multiplied by 100 to obtain in percent the coefficient of drainage for the period.

In the tank method, 5 tanks, 5¼ feet in depth and 4 feet in diameter, were installed in excavations provided for them. Two wells were installed in each tank so that the water level could be observed and so that a uniform water level could be maintained across the tank. Successive layers of fine sand, gravel, coarse sand, and finally the excavated material, were placed in the tanks. Each tank was covered first with a layer of tar paper and then a layer of rock wool to prevent evaporation from the soil surface. A canvas cover over a center pole produced a tent-like shelter against wind and rain. A continuous water-stage recorder was installed over the recharge well of each tank. The coefficient of saturation was determined by adding measured amounts of water to the tanks and evaluating the resulting rises in water level. The coefficient of drainage was determined by withdrawing measured amounts of water from the tanks and evaluating the resulting declines in water level. The tanks were allowed to stand undisturbed for 8 days or more after each addition or withdrawal of water. The coefficient of saturation was usually greater than the coefficient of drainage.

In the laboratory method, a drainage cylinder 7 inches long and approximately 5 inches in diameter was used. A drain cock and screen were installed at the bottom and de-aired water was added until the cylinder was approximately one-third full. Material was then poured uniformly into the cylinder, adding water as needed to keep the water surface above the top of the sample material at all times. The cylinder was filled in 3 parts with each part tamped vertically 25 times with a wire rod about 0.1 inch in diameter. After the last tamping, material was coned above the top of the cylinder and the side was tamped 25 times with a hard rubber tube. The material was then leveled and water was drained out of the sample. As soon as the water level had declined to the level of the screen, a 3/16-inch hole at the base of the percolation cylinder was opened to allow air to enter below the screen. The water removed during the preliminary drainage period of 5 to 15 minutes was measured and recorded. The cylinder was then removed from the apparatus and a paraffin paper cap was placed over the top to prevent evaporation. A wick of turkish toweling about 18 inches long was forced through the bottom of the permeameter, through the 1/4-inch hole in the screen and up into the sample about 5 inches. This provided a rapid means of removing much of the remaining water from the sample and disposing of it by dripping and evaporation. The percolation cylinder was then weighed and placed on a drain rack with the free part of the wick exposed to the air. During the early development of the method the samples were left to drain on the rack for 192 hours and were weighed twice a day. A graph showing rate of drainage was prepared by blotting the gross weight of the cylinder and sample, in grams, versus the elapsed time. The percentage of loss by draining was the volume of total loss (included both periods of drainage) divided by the volume of the sample, multiplied by 100. A 48-hour drainage period was finally adopted because it was the shortest period that gave a consistent relation between the percentage of loss by draining and the coefficient of drainage. A curve for the drainage coefficient determined from the long cylinders and tanks versus the loss by draining determined by the laboratory method was plotted for sand and gravel samples.

In the miniature pumping-test method, an attempt was made to determine the water-bearing characteristics of only a thin zone by means of a pumping-test that would affect only a few feet in the upper part of the saturated zone. An 8-inch bored well that penetrated only 2 feet of saturated material was used. One-inch sand point observation wells were driven to about 3 feet below the water table. Four of the wells were in a line with the pumped well, spaced about 3 feet apart, with two on each side of the pumped well. Two more observation wells were installed about 100 feet from the pumped well, one on each side. The test covered a period of 7 hours and 45 minutes; the average discharge of the pump was 0.783 gallons a minute, and the maximum drawdown in the pumped well was 1.55 feet. Transmissibility of the material for a zone of unknown thickness was computed to be about 5,000 gallons a day per foot. The authors concluded that the coefficients of transmissibility and storage were erroneous because existing formulae assume flow through the entire thickness of the aquifer. The authors believed that method should be developed further before it could be successfully used in field determinations of the coefficient of drainage and saturation in a thin zone.

Goldbeck, A. G., and Jackson, F. H., 1921, Tests for subgrade soils: Public Roads, v. 4, no. 3, p. 15-20.

A series of laboratory tests used for evaluating soil types as they affect the stability of roads were discussed. The moisture equivalent and water-holding capacity represented two of these tests.

In the moisture-equivalent procedure, a 5-gram sample was placed in a Gooch crucible and allowed to take up water by vertical capillarity until saturated. The sample was then placed in the Babcock cup of a centrifuge and centrifuged at 750 times the force of gravity for 1 hour. Moisture equivalent was obtained by weighing the soil after centrifuging, drying to constant weight in an oven at 100 degrees centigrade, and weighing again. The moisture equivalent, which represented the percentage of water retained by the soil after centrifuging, increased with the percentage of very fine silt and clay in the soil.

The water-holding capacity of a soil was considered the maximum percentage of water that could be retained by a soil under any circumstances. In the determination of the water-holding capacity, a soil was placed in a brass cylindrical box, 1 centimeter in height and 6 centimeters in diameter, the bottom of which was perforated with holes. The box was weighed, submerged in water for 1 hour, removed from the water, and again weighed immediately. The water-holding capacity was determined by comparing this weight with the weight of the dry soil.

Gurr, C. G., Marshall, T. J., and Hutton, J. T., 1952, Movement of water in soil due to a temperature gradient: Soil Sci., v. 74, p. 335-345.

This paper reported an attempt to assess the contribution of liquid and vapor flow to the temperature effect by measuring changes in the distribution of a small amount of soluble salt in the soil. The movements of soluble salts were assumed to be due to the transport in the liquid phase only.

A "Perspex" cylinder, 10 centimeters in length and 14 centimeters in internal diameter, held the soil. On one end was bolted a heavy brass plate to which an electric heater was attached. On the other end was bolted another brass plate. Cold water could be pumped then through the cylindrical water chamber thus formed. The temperature of the plates or faces was measured by thermistor and the temperature gradient in the column also was measured by 4 thermistors installed in the soil column at 2 centimeter intervals. The soil, which had been wetted previously and which had been allowed to stand in a sealed can for at least a week, was compacted firmly in the cylinder. Samples were taken to determine the initial water and salt content. Then the apparatus was placed horizontally in a constant temperature cabinet at 20 degrees centigrade. A temperature of approximately 10 degrees centigrade was established at the cold face and a temperature of approximately 25 degrees centigrade was established at the hot face. At the end of the run, the final temperature was determined. The temperature gradient approximated 1.6 degrees centigrade per centimeter. After the apparatus was dismantled, the soil column was sampled for moisture content. These samples were then transferred to 75 milliliters of distilled water and shaken for 1 hour. An index of the content of total soluble salts was obtained by measuring the electrical conductivity of the suspension. The chloride content was determined by an electronic titration. Two soils, a loam and a fine sand, were used in the tests. The fine sand was washed and sodium chloride was added subsequently. The loam soil was used without addition of salt. Pressure-membrane, freezing point, and tensiometer data were obtained during drying and vapor pressure was obtained during wetting. Different amounts of water were added to the loam soil to provide a wide range of initial moisture contents.

In all but the wettest and driest columns of loam soil, a transfer of water towards the colder end of the columns and of chloride toward the hotter end occurred. This was attributed to a set transfer of water from hot to cold, in which water evaporating from the hotter soil moves as a vapor into colder soil, where it condenses and returns as a liquid when a favorable gradient of pressure potential has been established. Theoretical considerations indicated that when the soil is sufficiently wet to permit liquid flow, a state of equilibrium cannot be reached, and a continuous circulation of air takes place. The maximum net transfer of water from the hot to the cold end occurred in columns where the initial water contents were approximately one-third of the moisture equivalent. Movement of water in the liquid phase was known to have occurred at a low water content, which corresponded to a pressure potential of -0.8×10^7 ergs/gm.

Hanks, R. J., Holmes, W. E., and Tanner, C. B., 1954, Field-capacity approximation based on the moisture-transmitting properties of the soil: Soil Sci. Soc. America Proc., v. 18, no. 3, p. 252-254.

The object of this investigation was to determine the relationship of the field-capacity approximation to the field capacity, and to compare the approximation with commonly used methods of estimating field capacity. The field capacity, field-capacity approximation, moisture equivalent, 1/3-atmosphere percentage, Buckner-funnel percentage (Bouyoucos moisture equivalent), and modified Wilcox estimate of field capacity were determined for twenty-three soils. In the field-capacity approximation procedure nondisturbed soil cores 3 inches in height and in diameter were used for most of the tests. The samples were wetted from below. About twenty-four hours after the soil became wet at the top, the samples were placed on a large porous plate in a pressure chamber and subjected to a pressure of 0.20 atmosphere. The moisture extracted from the soil samples was measured at selected intervals, and when the moisture extraction rate became negligible, the moisture content of the upper part of the samples was determined. Loose soil at the top and the bottom 1 inch of the 3-inch cores was discarded before making the moisture determinations.

Because it was impossible to obtain nondisturbed cores of the more cohesionless soils that were sampled, bulk samples were collected and packed by hand in cylinders about 8 inches in height and 3 inches in diameter. The bottom 3 to 4 inches of the 8-inch samples were discarded before making the moisture determinations.

The field capacity was determined on soils which had deep-water tables and no impervious layers. An area of 1 square rod was cleared of vegetation and covered with tar paper for a period of 2 weeks before wetting. Water was added to a centrally located 6-foot-square area, or to infiltration rings, until the soil was wet to a depth of 30 inches. Soil-moisture samples were taken from the wetted areas at about 2-day intervals for 10 to 13 days. Plots were kept covered except when sampling. To minimize evaporation effects the top 2 inches of the subsoil samples were discarded. Field capacity designaged as the moisture content at which drainage was negligible was determined from the time-moisture content curve.

The Wilcox method was modified considerably for these tests. Undisturbed soil cores, 3 inches in height and in diameter, were taken in pairs. One core from each pair was air-dried and the other core was saturated with water. The saturated core was placed on its air-dry "twin."

A layer of cotton gauze was fastened on the bottom of the wet core and a "facial tissue" was placed between the cores. The top wet core was weighed at about 2-hour intervals. After the weight remained the same for two consecutive weighings, the moisture content was determined. Most soils required about 8 hours drainage. No serious error was found due to leaving the cores in contact a few hours after drainage became negligible. The procedure was not applicable to the soils that had the coarser textures.

Results from this study indicated that the correlation between field capacity and the field-capacity approximation was significantly higher than between field capacity and any of the other estimates except the modified Wilcox method. There was no significant difference between the correlation coefficient of the field-capacity approximation and the modified Wilcox method with field capacity for the nine soils tested. The modified Wilcox method, however, was much more time consuming and inconvenient that the field-capacity approximation. In addition, the variability of the individual sample results obtained by the modified Wilcox method was greater than for any other method.

The addition of a wetting agent (Ultrawet) at a concentration sufficient to lower the surface tension of the water by a factor of about 3, did not affect the field capacity or the field-capacity approximation, but decreased the moisture equivalent and 1/3-atmosphere percentage. This further demonstrates that field capacity and the field-capacity approximation are not equilibrium measurements as are the moisture equivalent and 1/3-atmosphere percentage. This also further strengthens the close relationship of the field-capacity approximation to field capacity.

The data of this investigation supported the view that field capacity of a thoroughly saturated soil is the moisture content which occurs after drainage has decreased the soil-moisture content to such an extent that the consequent large decrease in capillary conductivity impedes further drainage of water. The decrease in capillary conductivity is not considered to be the entire cause of the field-capacity phenomomem. A decreasing hydraulic gradient is also important.

Hazen, Allen, 1891, Experiments upon the purification of sewage and water at the Lawrence Experimental Station, No. 1, 1889, To Dec. 31, 1891: Massachusetts State Board Health Pub.; Dec., no. 34, p. 425-633.

The performance of different filtering sands was studied in relation to their particle size. The amount of water retained after the sand had drained depended on the size of the particles, the finer particles retaining more water that the coarser particles. The percentage of water retained in the coarser sands was about the same throughout the sand bed. The percentage of water retained by the finer materials was greater in the lower part of the bed than in the upper. The author explained this by stating that the height to which capillary attraction was effective was in inverse proportion to the square of the particle size.

Hilgeman, R. H., 1948, Changes in soil moisture in the top eight feet of bare soil during twenty-two months after wetting: Am. Soc. Agronomy Jour., v. 40, no. 10, p. 919-925.

In irrigation experiments with guayule in Arizona, a high consumptive use of water was observed and high values for field capacity were obtained from soil samples taken 3 days after irrigation. These facts suggested the possibility that considerable quantities of water were being lost by downward drainage. This study was undertaken to evaluate this possibility.

The soil on which these studies were made was Cajon silt loam, an unweathered alluvial soil, which was laid down in layers. This soil contained varying proportions of very fine sand, silt, and clay in the top 10 feet. The water table ranged from 65 to 70 feet below the surface during the period of study.

The first experiment was made on three bare plots, each 32 x 110 feet in size. The amount of soil moisture in each foot to a depth of 8 feet was determined by collecting moisture samples at selected time intervals from two areas in each plot after irrigation.

The second experiment was made in a 60- by 100-foot area. This land had been in guayule and had not been irrigated for 18 months prior to its sampling. All available soil moisture had been removed to a depth of over 8 feet. Duplicate sets of tensiometers were placed at depths of 24, 36, 48 and 60 inches in a circle in the central part of the area. The movement of soil moisture was determined by taking moisture samples from 4 stations around these tensiometers. Irrigation water was applied for a 72-hour period. Moisture samples were taken at 5, 10, 14, 19, 26, 33, 42, and 60 days after irrigation. Tensiometers were read 4 hours after irrigation and at frequent intervals thereafter.

Results from the first experiment indicated that over a period of 22 months considerable water was lost from the upper 8 feet by drainage and evaporation. This loss represented 47 percent of the available water. Approximately 75 percent of the available water in the third and fourth foot was lost. This large loss suggested that the movement of water was both upward and downward from this zone. The rate of loss of water seemed to be related to the temperature and the moisture content of the soil. The most rapid losses occurred in the summer. More water was lost during the first summer than during the second.

In the second experiment the soil was wet to a depth of 4½ feet. Changes in moisture and tension were studied in the upper 90 inches of soil over a 60-day period. Soil moisture moved downward throughout the entire 60-day period. A rather marked decrease in the rate of movement through the upper 30 inches of soil occurred during the first 19 days. Losses from the 30- to 54-inch zone were gradual and became less with time. Approximately 10 percent of the total available water in the soil 5 days after irrigation was not present 55 days later. This loss, which evidently extended to a depth of 54 inches, was attributed to evaporation. Results from this experiment indicated that field capacity was attained 10, 14, and 19 days after wetting in the 0- to 12-, 12- to 18-, and 18- to 30-inch zones, respectively. Field capacity in the 30- to 54-inch zone was attained apparently prior to 5 days after wetting because no break in the drainage curve was recorded. Moisture tension found in the different zones at the time when field capacity was attained, ranged from 190 to about 455 centimeters of water. The determination of field capacity for the soil studied was difficult to obtain and was subject to considerable error. The high rate of drainage traced in the subsoil suggested that considerable quantities of water frequently attributed to loss by transpiration of plants in water studies may actually be lost through drainage.

Israelsen, O. W., 1918, Studies on capacities of soils for irrigation water and a new method of determining volume weight: Jour. Agr. Research, v. 13, p. 1-35.

This paper reported upon observations of the capacities of soils of the Sacramento Valley, Calif., to retain water under different field conditions. Plots of land were selected where the water table and capillary fringe were at a considerable depth below the land surface. Then the soils at the surface were wetted by enough water to add a given percentage to a unit weight of soil. They were then allowed to drain. Over a period of several years (and many irrigations), samples for determination of moisture content and porosity were taken before and after irrigation. These results were presented in tabular and graphical form.

A new method of determining volume weight in place was described. This required essentially the calculation of volume by filling a hole at any depth less than 6 feet with a known volume of water. This was accomplished by inserting a thin-walled, elastic rubber tube into the hole and filling it with water to a point flush with the surface. By correcting for the volume of the rubber tube, and determining oven-dried weight of the soil as usual, the volume weight of the undisturbed soil was obtained.

Jamison, V. C., and Reed, I. F., 1949, Durable asbestos tension tables: Soil Sci., v. 67, p. 311-318.

A modification of the soil-moisture tension table of Leamer and Shaw was described. Water-stable asbestos board clamped over a copper screen in a tray constructed from angle and sheet brass was used as a membrane. Also described was a method of arranging table units in a convenient cabinet form to conserve space. The operation of the device was explained and the results were given for a test with soils, including samples of a sand, a loam, and two clays. The results indicated that a 1-day equilibration time is more than adequate to take most soils from saturation to any tension up to 120 centimeters of water.

Joseph, A. F., 1927, The moisture equivalent of heavy soils, pt. 2: Jour. Agr. Sci., v. 17, pt. 1, p. 11-20.

This study evaluated the extent to which the moisture equivalent of a soil is an additive property dependent on the nature of the particle-size fractions. The moisture equivalent for six soils differed on the average by 10.7 from estimated values determined by use of Middleton's formula. In this formula the moisture equivalent equals 0.0426 (percent clay) plus 0.291 (percent silt) plus 0.063 (percent sand.) Positive evidence that the moisture equivalent was sometimes an additive property was presented. Moisture equivalent determinations were made on several mixtures of pure clay and ignited sands.

For these mixtures, the formula 0.595 clay - 0.032 sand gave a reasonable approximation of the moisture equivalent. The author emphasized that true additive relationships can, in general, only be obtained from series of soils of the same nature and if taken at the same depth. The properties of clays differ greatly in their chemical properties and moisture equivalents. The silt fraction of soils may also exhibit considerable differences in physical and chemical properties. The results of studies of aeolian and alluvial soils indicated more differences both in the chemical composition and in the moisture equivalent in the silt fraction than in the clay fraction.

Joseph, A. F., and Martin, F. J., 1923, The moisture equivalent of heavy soils: Jour. Agr. Sci., v. 13, p. 49-59.

Relationships between the moisture equivalent and other soil properties such as salinity and alkalinity were reported. Fine-textured soils from northern Sudan were used.. Except in some of the thicker samples where water accumulated on top, the moisture equivalent decreased as the quantity of sample centrifuged was increased. This decrease in moisture equivalent was believed to be due to an increase in the density of the soil. Moistening the soil with a 0.1 percent solution of calcium sulfate, ammonium nitrate, or ammonium sulfate reduced the moisture equivalent, whereas moistening the soil with a 0.1 percent solution of sodium carbonate increased the moisture equivalent.

Kellogg, F. H., 1950, Rate of depletion of water-bearing sands: Mississippi Geol. Survey Bull. 70, p. 1-15.

Because unsteady-state flow methods of estimating rates of depletion of aquifers were considered to be poorly founded, an alternate method was suggested.

Methods based on this premise use a basic equation as follows:

$$K_w\Delta^2 hdt = G_s dh$$

where K_w = constant for ground-water flow known as the permeability coefficient,

Δ = temperature increment,

h = hydraulic head,

dt = increment of time,

G_s = total quantity of water that can be drained from a unit volume of sand (specific yield),

dh = increment of head change.

This is represented graphically by the ordinate of the horizontal part of a yield-pressure curve. The right-hand side of the equation must represent the net flux (net difference between inflow and outflow) due to a head change dh or $dqw/dh = G_s$. Hence G_s must represent the slope of the linear yield-pressure diagram. The equations above require a sloping straight line where the pressure head required to remove all drainable water be G_s divided into the volume of drainable water. No actual relationship of this kind existed. Therefore, the assumption that specific yield in ground-water flow is analogous to the product of specific heat times density of conducting medium in heat flow seemed untenable.

Comparison of field observations with computations based upon the above equation have shown close agreement provided the line of observations passes through the center of the well. This is the condition expected analytically because any differential equation must be solved so as to conform to certain boundary conditions. However, the farther away from the boundary (the face of the well where head equals zero) the point occurs, the greater will be the discrepancies due to the erroneous differential equation.

The author's conclusions were:

(1) The assumptions of analogy between unsteady-state flows of heat and of ground water is approximately correct only when drawdown does not exceed mean height of capillary rise.

(2) The conception that specific yield is analogous to specific heat multiplied by density has no analytical foundation.

(3) Drainage of aquifers seems to be a problem of flow in the steady state.

The author discussed the last conclusion in detail. Drainage was considered in a steady-state system where water flowing out was replaced by an equal amount of air flowing in. Thus, atmospheric pressure existed in the pores of the sand immediately above the capillary water. However, in the unsteady-state system, the assumption was made that the whole body was full of capillary water which was drained only when drawdown exceeded height of capillary rise for a channel ending in a given pore space at the top of the sand. Under such conditions, pressures less than atmospheric would exist at all times in numerous pore spaces throughout the body. If the sand were sufficiently pervious, a constant stream of air entering all pore spaces at the top of the sand could be imagined as drainage proceeded.

A steady-state flow in two phases would be the result and the rate of advancement of the air-water interface would become a strong interest.

The final conclusion was that estimates of the rate of depletion of aquifers be based on the analysis of two-phase steady flow involving air and water.

Kemper, W. D., and Lutz, J. F., 1955, Hydraulic conductivity in large channels as determined by an electric analog: Soil Sci., v. 81, p. 283-286.

An electric analog was set up to measure the contribution of highly conductive channels such as cracks and root holes to the total hydraulic conductivity of soils. The conductivity of channels approximately 0.5 centimeter in diameter was very large in relation to that of the surrounding soil. The electric analog was constructed so that this same relationship held for the conductivity of different lengths and amounts of polished wire suspended in an agar gel. The percentage increase in conductivity seemed to be a linear function of the total length of wire. An equation was then developed, based on the work with wire, which gave an approximate relationship of the number of conductive channels per unit volume, the average length of these channels, and their contribution to the overall hydraulic conductivity of the soil. This equation was

$$C = 44.57\ mnL^3$$

where

C = percentage increase in conductivity of system containing no wire, and nL = length of wire added per unit volume.

The equation was thought to be most accurate when the channels were large in diameter, equal in length, homogeneous in distribution, random in orientation, and relatively few in number.

King, F. H., 1898, Principles and conditions of the movement of ground water: U.S. Geol. Survey 19th Ann. Rept., pt 2, p. 85-91.

As a part of an extensive investigation of certain theoretical phases of the movement of ground water, the author discussed some experimental work on the amount of water retained by capillarity.

To determine the water-holding power of long columns of material, five galvanized-iron cylinders, 8 feet long and 5 inches in diameter, were constructed. A plate was placed on the bottom and top of the cylinders and drainage or recharge was permitted by a stopcock through the lower plate. Evaporation was minimized by inserting in a hole in the upper plate, a cork containing a glass tube drawn to a fine capillary. The cylinders were then filled with five sorted and chemically inert sands ranging from No. 20 sand (effective diameter = 0.4745 millimeters; porosity = 38.86 percent) to No. 100 sand (effective diameter = 0.0827 millimeters; porosity = 39.77 percent). After water had been slowly introduced from the bottom and the cylinders were full, the water in the columns was allowed to drain into receiving flasks. The rate of drainage was measured by readings at frequent intervals during nine consecutive days and less frequently then over the following 2½ years. These results were shown in tabular form. The moisture retained at the end of this period was noted, determined gravimetrically by cutting the tubes into 3-inch sections and drying the sand to constant weight at 110 to 120°C. The total quantity of water retained and the distribution of the retained water was shown in tabular form. This data was then related to ground-water recharge in humid climates.

Klute, Arnold, 1952, Some theoretical aspects of the flow of water in unsaturated soils: Soil Sci. Soc. America Proc., v. 16, no. 2, p. 144-418.

Application of Darcy's law to flow in unsaturated media, its validity, and the assumptions made in its use were discussed. The total soil-moisture potential was defined and its components discussed. Also briefly considered was the relationship of the permeability coefficient in Darcy's law to various soil parameters. By use of the equation of continuity and Darcy's law, an equation of flow was derived, and the general problem of obtaining solutions of this equation was outlined. The special mathematical difficulties encountered when the permeability was considered as a function of the moisture content or the capillary potential were also indicated. Using a functional relationship between the permeability and the moisture content, a numerical solution of the flow equation for a semi-infinite system was given and the phenomenon of a wetting front was shown to be indicated.

Kocatas, B. M., and Cornell, David, 1954, Holdup and residual saturation of hexane in gravity-drained soybean flake beds: Indus. and Eng. Chemistry, v. 46, no. 6, p. 1219-1224.

The effect of variations of average diameter, flake thickness, and bulk density on the holdup of hexane in soybean-flake beds has been determined for three samples of soybean flakes with a range of properties covering the commercial limits. Changes in holdup (or pounds of total solution remaining in the bed per pound of oil-free and moisture-free flakes) due to changes in these properties vary inversely with the resultant changes in permeability.

Depending on the temperature and bed properties, the approach to equilibrium becomes very slow after a rapid initial draining period (which reduces the holdup to about 0.5 pound of solution per pound of oil- and moisture-free flakes). Draining was believed to take place entirely by viscous flow. The authors expected that the ultimate equilibrium holdup would be about 0.3 to 0.4 pound of solution per pound of oil- and moisture-free flakes.

At the end of the draining period the column was disassembled. The saturation distribution curves determined by analysis of flake samples taken from the bed at this time showed a wide range in saturation from the top to the bottom of the column. Although the saturation gradient was partially due to the capillary end effect, the main reason for the large saturation gradient in high beds was the slow approach to equilibrium. Short beds had a very high initial draining rate but a higher ultimate holdup than high beds because of the importance of the end effect. The choice of bed height for minimum holdup depended on the permeability of the bed and the length of draining time involved.

Lambe, T. W., 1948, The measurement of pore-water pressures in cohesionless soils: Internat. Conf. Soil Mechanics and Foundational Eng., 2d, Proc., v. 7, p. 38-41.

The value of pore-pressure measurements in cohesive soils has received greater recognition in the engineering field. Research into capillary and other flow problems can often be aided by knowledge of the heads which are causing the flow. This paper described an apparatus and technique for measuring pore-water pressures in cohesionless soils.

The test apparatus included a lucite tube of 2½-inch diameter containing piezometers spaced along the length of the tube, and water flowing vertically. The head was measured by a 12½-foot water manometer or by a differential mercury manometer. A balancing tube compensated for capillary effects within the apparatus. Two types of piezometer pilots were used, one a 200-mesh screen over a one-eigth-inch opening in a brass compression fitting, and the other, a hypodermic needle. Either one screwed into a threaded hole in the side of the lucite tube. Transparent plastic tubing was used to connect pilots with valves. The pilot was flush with the inside of the tube. Experimentation indicated that pressure readings on pilots mounted at equal elevations on both sides of the tube were the same, suggesting equal pressures throughout the cross-section of sample at any one elevation. The problem of air entrapment in the pilot was encountered and corrected only by flushing. The test was run on partially saturated fine sand.

The author made measurements during a falling-head permeability test on a partially saturated soil using the above apparatus.

Lambe, T. W., 1950, Capillary phenomena in cohesionless soils: Am. Soc. Civil Engineers Trans.: paper 2435, p. 401-430.

In the design of roads and airport runways, earth dams, earth slopes, and walks, one of the important considerations is the proper control of subsurface water. The movement and retention of this soil water can be dependent largely on the phenomena of surface tension as exhibited by capillaries.

The purpose of this study was to obtain numerical measures of capillarity in many different soils and to evaluate and explain some of the current theories of capillarity.

All experimental work reported upon was done, under similar test conditions, on a fine, uniform, natural sand which had a 60-percent size of 0.19 millimeter and 10-percent size of 0.08 millimeter.

The author first discussed the mathematics of horizontal flow in unconnected and interconnected systems of capillary tubes of varying diameter.

The flow in two interconnected horizontal tubes had a number of characteristics which had counterparts in capillary flow through soils:

(1) The meniscus in the smaller tube preceded that in the larger tube.

(2) The only tube diameter that was effective at any instant and at any point in developing capillary head was that diameter where a meniscus was formed.

(3) The gradient in tubes back of the air-water interface in the large tube was essentially constant, because lateral flow between the tubes equalized heads.

(4) The average gradient for the system depended on the larger tube, since a large part of the greater capillary head of the smaller tube was lost in the distance between the two menisci.

(5) In this zone between the two menisci, k' (which corresponds to permeability in soils) and the degree of saturation for the system were lower than those back of the lagging meniscus.

Some of the more important inherent differences between soils and a system of capillary tubes, which make their flow processes unlike, were:

(1) The ideal situation of 100-percent saturation usually was not attained in soils because of air-entrapment phenomena.

(2) In soils, the equivalent-capillary tubes changed diameters from point to point. Capillary flow was consequently very jerky.

(3) There was an infinite number of effective tube sizes in soils.

In studying horizontal capillarity the author used a segmented lucite tube mounted horizontally. Heads were maintained at 35 and 181.7 centimeters and pressures were determined from piezometer outlets at increasing greater distances from the end of the tube.

The degree of saturation was determined by running two similar tests and plotting the percent saturation versus distance behind the wetter surface. This plot showed less than 80-percent saturation.

The author then developed theoretical methods of analysis of horizontal capillary flow that were different from standard permeability tests. He emphasized the importance of air entrapment in comparison of test results. In summary he stated that an equation using data from his horizontal capillarity test gave a good measure of the permeability of the soil that has been wet under the influence of capillarity.

The ideas developed in horizontal capillary flow were applicable, in general, to vertical flow, except that in vertical flow elevation head was more important. In capillary rise, the effect of elevation head was to decrease total head, thus reducing the rate of flow, and to make the range of low degree of saturation much larger and more important than in horizontal flow.

In order to obtain curves of the degree of saturation and the total head versus the distance, a vertical segmented lucite tube with piezometer outlets was used.

During these tests, water from a constant-head reservoir was permitted to rise by capillarity into the soil sample; the distance above the bottom of the soil, the water pressures at the piezometer outlets, and the elapsed time were recorded. Plots were then made of height of rise in centimeters (from bottom of soil) against degree of saturation.

Theoretical methods of analysis of capillary rise were developed and, as before, the author went from a discussion of simple systems of tubes to the discussion of soils. The following conclusions were drawn regarding drainage of soils:

(1) Drainage proceeded at an increasingly slower rate owing to decrease in gradient with time.

(2) In a system of two tubes with different diameters, the larger tube had the greater influence on the rate and the quantity of drainage.

(3) The water held in the tubes above the elevation of the maximum capillary head was not continuous, but was water entrapped in the cross connections. Therefore, the quantity retained was not a function of the elevation.

In addition to the differences in tubes and soils listed in the discussion of horizontal capillarity, another difference was important in drainage. This was that natural soils could be partly saturated from the beginning. Entrapped air caused time lags for changes in water to occur; if the entrapped air was not distributed uniformly the gradient was affected.

The drainage data on soils were plotted as head and quantity against time.

Two interesting conclusions were drawn:

(1) A considerable amount of drainage occurred after the visual line of saturation appeared to have reached its ultimate position.

(2) The visual line of saturation was not the true line of saturation.

The drainage of a partly saturated soil was of much practical interest because a soil saturated by capillary action may have a degree of saturation considerably less than 100 percent. A test was run on a soil which had an average degree of saturation of 79 percent. This test revealed two main differences between the drainage of a saturated and partly saturated soil mass. The first was the lower total ultimate quantity drained from a partly saturated soil. The second difference was the slower rate of flow and the slower rate of head dissipation in the partly saturated soil. The rate of flow of water was lower because the permeability was decreased by entrapped air.

Another time lag occurred in the drainage of partly saturated soils which was due to the time required for adjustment to changes of water pressure. This was due to the compressibility of air (therefore, in order that air may adjust itself to a difference in pressure, it must compress or expand). This volume change in the air necessitated a flow of water which in turn required time.

A graph comparing drainage of a partly saturated soil versus a completely saturated soil was shown. The curves agreed in early readings but diverged during later readings.

Finally this paper emphasized that the capillary characteristics of a given soil cannot be completely represented by one or two head values. Several capillary heads as follows were required for adequate representation:

(1) Maximum capillary rise - Point at which curve ceases to be vertical.

(2) Saturation capillary head - Highest point at which complete saturation exists.

(3) Capillary rise - Highest point to which capillary water rose.

(4) Maximum capillary head - Point at which the maximum capillary degree of saturation exists.

The author suggested that the term "capillary head" be used in place of "capillary rise."

Leamer, R. W., and Shaw, B. T., A simple apparatus for measuring noncapillary porosity on an extensive scale: Am. Soc. Agronomy Jour., v. 33, p. 10003-10008.

A simple inexpensive apparatus for measuring soil moisture content at low tensions was described. The apparatus consisted basically of an ink blotter which acted as a porous material that would support an extended column of water. By the use of a double layer of blotters, tensions as high as 100 centimeters could be obtained satisfactorily.

Lebedeff, A. F., 1927, The movement of ground and soil waters: Internat. Cong. Soil Sci. Proc., v. 1, p. 459-494.

In order to simplify the study of the complex process of the movement of ground and soil waters, the author considered it expedient to investigate and discuss two phases of this process, namely, the movement of water in the form of vapor and the movement of water in the liquid phase.

For the study of the vapor phase, a hair hygrograph was used. The hair was placed in the soil kept at a constant temperature in boxes. These investigations have indicated first that if the soil contained moisture in an amount greater than the maximum hygroscopicity, the relative humidity of the soil air equaled 100, and second, if the soil moisture was lower than its hygroscopicity, the soil air was not saturated and its relative humidity was less than 100 percent. The drier the soil was, the less was the relative humidity of its air. With the moisture content of the soil remaining constant, the relative humidity of the air increased with an increase in the temperature.

Observations on the relative humidity of soil air in nature have indicated that soil air under natural conditions was always saturated with water vapor below the depth of 5 to 10 centimeters. Therefore, the vapor pressure of the soil could be correlated with its temperature. In the upper layer the soil air did not reach saturation and the vapor pressure was less--the drier the soil, the lower its temperature.

Under natural conditions, the enrichment of the soil with condensation water must be greater than what has been found because during the spring, summer, and part of the fall, the moisture content of the upper layer of the soil dropped very frequently to a point lower than its maximum hygroscopicity. Under such conditions the vapor pressure in the soil was lower than the maximum vapor pressure at a given temperature. Therefore, the difference between the absolute humidity of the air and the vapor pressure in the soil (condensation potential) was greater, whereby the intensity of condensation increased. A dry-surface layer of the soil played a very important hydrological role. In the process of water evaporation, the surface layer not only lost but also gained some water during the dry periods. In the upper-most layers a relation of vapor pressure existed whereby this layer underwent a double evaporation into the atmosphere and into the deeper layers of the soil. The water being distilled into the lower layer was condensed and increased the resources of liquid water. In the formation of ground water by condensation, the place of formation and location of the water table may or may not coincide, depending on the physical structure of the corresponding horizons of the earth's crust. With an increase in depth, a constantly increasing temperature caused the water vapor to move from the lower to the upper layers. In places where this movement was the more intensive, an accumulation of liquid water could be observed. In those layers where the thermal gradient was lower, the intensity of the movement of the vapor, all other conditions being equal, must be greater than in the layers with a higher thermal gradient. The consequence of this was the accumulation of rings of liquid water which could not only be seen with the naked eye but could be determined gravimetrically.

With soils having a moisture content greater than their maximum hygroscopicity, experiments have indicated that when two layers of different moisture contents came in contact, the water moved as a liquid from the more moist layer of the soil into the layer of less moisture. This was not true where the soils have moisture contents smaller than their maximum hygroscopicity. Here, movement occurred as vapor only. Thus, the maximum hygroscopicity of a soil was that lower limit of moisture at which there was no possibility for liquid water to move under the influence of molecular forces. To approach this limit the author drained tubes of sandy soils 4 to 5 centimeters in diameter and 1 to 3 meters in length. A gauze was used in the bottom of the tube and a head of water of 2 centimeters was used to saturate the soils. After drainage the soil moisture in 10-centimeter segments was determined gravimetrically. The experimentation indicated that a uniform moisture content (1.81 percent) prevailed in the upper part of the column (60-150 centimeters). This moisture content was considered by the author to be the "molecular moisture-holding capacity" of the soils.

To speed up the drainage of excessive moisture from the soil and to obtain molecular moisture-holding capacity the author used a centrifuge which developed a force of 400 times gravity. Moist sand was placed in tubes in layers 3 centimeters thick and centrifuged 1 hour. After centrifuging, the amount of moisture left in the sand was determined (eliminating 2 to 3 millimeters of the dried outer layer of the sand). The moisture content after centrifuging was 1.91 percent.

The maximum molecular moisture-holding capacity compared very favorably with the value obtained by the method of long columns. If the sand from the upper part of the long columns after drainage was given the centrifuge treatment, no water was forced out, indicating a similar effect at 1 and 400 times gravity. The author then determined that the intensity of movement of film water (within the limits of maximum hygroscopicity to maximum molecular moisture-holding capacity) in moving from one layer into another was not dependent on the forces of gravity. In proving this, however, compaction of a sample by pouring it into a tube became a factor.

To determine the characteristics of gravitational water the author saturated (under 2 centimeters head for 3 hours) six pairs of sand tubes 10, 20, 30, 40, 50, and 100 centimeters high. One tube of each pair was immediately sampled for moisture content. The other tube was allowed to drain until drainage ceased and then sampled for moisture content throughout the tube. The original moisture content in each height of tube was equal to 21 to 23 percent. But after drainage the moisture content at the upper part of the tubes was about 22, 22, 9, 4, 2, and 2 percent in moving from the 10- to the 100-centimeter-height columns. The longest tube drained for 3½ days during which 300 cubic centimeters of water was collected. In this column the distribution of moisture content became constant at about 40 centimeters. Changing the height of the column did not change this distribution; it only gave a longer section of column with constant moisture distribution.

The author reported in tabular form the results of drainage of layered materials and loess. The experiment suggested at the moment of equilibrium that the moisture in heterogeneous materials distributed itself differently than in soils with homogeneous structure; if the large-grained mass was below the finer-grained mass then more water was retained in the fine-grained mass than if it were underlain by material of the same type. Data were shown that emphasizes the importance of particle size on the distribution of water in columns after drainage. In a column of four layers starting with the finer at the bottom and the coarser at the top, the moisture content from top to bottom was about 2, 2.5, 29, and 44 percent, respectively.

The author also noted that in a uniform soil the movement of film water from moist to dry zones took place slowly. This picture changed when two layers of different particle size were placed in contact with one another; under such conditions the water moved rapidly from the layer with a relatively lower inner surface of soil particles towards the finely-grained layer. These observations explained the variations in the moisture content in layered materials in drainage.

The author discussed in some detail the accumulation of ground water. The soil water in the upper horizon of temporary moistening was believed to percolate through the underlying dry horizon in the same way that lithium chloride solution percolated through sand columns during this experimentation. In other words, water percolated through dry layers without saturating them. The gravitational waters combined with the condensation waters and together formed ground water. Thus, in the complex process of the formation of ground waters the movement of water in the form of vapor from the surface of the soil downward, and from the deeper layers of the ground upward as well as the movement of liquid water under the influence of gravity, were considered to be only separate items. The author emphasized that in the temperate latitudes the condensation process and infiltration combined and either one predominated depending on climatic and geological conditions of the region.

Leverett, M. G., 1941, Capillary behavior in pore solids: Am. Inst. Mining Metall, Petroleum Engineers Trans., v. 142, p. 152-168.

The application of established thermodynamic and physical principles to problems of petroleum production was the object of this paper.

Although dynamic problems involving analysis of the motion of mixtures of immiscible liquids in porous media were discussed, the study was concerned primarily with problems involving the static balance between capillary forces and those due to the difference in densities of the fluids; that is, gravitational forces.

The abandonment of the "capillary tube" concept of sand structure and its replacement by the concept of a characteristic distribution of interfacial two-fluid curvatures with water saturation was urged.

The curvature-saturation relation for unconsolidated sands was determined experimentally and the results obtained were correlated so as to apply to all clean unconsolidated sands. To do this, the author selected the height-saturation method. Water and air were allowed to reach capillary equilibrium both by drainage and imbibition in a vertical sand column (3/4 inch in diameter, 10 feet in length) and the resultant water saturations measured at numerous heights after at least 2 weeks time. Saturation changes were followed by a conductometric technique and moisture contents of 10-centimeter sections of tube were determined gravimetrically. By plotting saturation against height for four clean sands, data fell satisfactorily near two curves, one for imbibition (lower) and the other for drainage. Data for clayey sands did not fit this relationship because water was more tightly held owing to the absorption effect of the hydratable material.

The author showed, in addition, that the two-fluid interfacial surface area was determinable from thermodynamic consideration of the curvature-saturation relation.

Low, P. F., and Deming, J. M., 1953, Movement of water in heterogeneous systems with special reference to soils: Soil Sci., v. 75, p. 187-202.

The total potential concept was applied to the movement and equilibrium of water in heterogeneous systems such as the soil. A general equation was developed, whereby the change in the total potential of a constituent during any process was considered to be due to changes in the positional potential energy of the constituent, to changes in concentration, and to changes in pressure and in temperature occurring in that process. Gravitational, electrostatic, and Van der Waals force fields in the system were considered to cause the positional potential energy of the water in the soil. The hypothesis was formulated that the rate of movement of the water in the soil was directly proportional to the driving force, which was believed equal to the negative gradient of the total potential of the water. Based on this hypothesis and the general equation for the total potential, an equation for the velocity of water in a heterogeneous system was developed. As a special case, the latter equation yielded Darcy's law.

The integration of the general equation produced a second general equation expressing the equilibrium-pressure difference between any two phases as a function of the concentration ratio and positional potential energy difference for any constituent distributed between the two phases. The capillary rise and Van't Hoff's law emerged as special cases. The general equation was applied to the water in the interfacial region of a clay particle, with the result that the hydrostatic pressure was shown to be greater in the micellar solution than in the intermicellar solution. The magnitude of the difference depended on the salt concentration in the intermicellar solution, the electrostatic potential in the micellar solution, and the distance from the particle. Simplification of the equation resulted in an equation for equilibrium osmotic-pressure differences originally attributed to Langmuir.

Lull, H. W., and Reinhart, K. G., 1955, Soil-moisture measurement: U.S. Dept. Agriculture, Southern Forest Expt. Sta. Occasional Paper 140, 56 p.

The authors reviewed most of the methods used to estimate soil moisture. These included consistency tests, the gravimetric method, soil points, water-absorbing liquids, heat of solution, heat diffusion, calcium-carbide method, constant-volume methods, resistance to penetration, electrical resistance method, tensiometers, air-picnometer, and the nuclear method. Only four methods, the gravimetric, electrical resistance, tensiometer, and nuclear methods were discussed in detail. A shortened author's evaluation of these methods follows:

Electrical resistance units were the most responsive to moisture-content changes from below wilting point to field capacity. At moisture contents between field capacity and saturation, the resistance change per unit change of moisture for the fiberglas and nylon units was less than at lower moisture levels. Also, field calibration of these units at moisture contents above field capacity was inaccurate because gravitation water was sometimes lost or free water was included.

Tensiometers were considered the best method for determination of soil moisture from field capacity to saturation. However, continuous records of moisture content could not be obtained using them because they could not be used at low moisture contents.

The accuracy of the various methods was difficult to evaluate. Bouyoucos and Mick believed that the relative experimental error of laboratory-calibrated plaster of paris units was from ± 0.1 to ± 1.0 percent. In a recent comparison at a Vicksburg site, nuclear equipment and fiberglas units gave 1.4 and 1.0 percent deviation by weight from the calibration curves, respectively. In more than 1,100 observations at three Vicksburg sites, the average deviation of the mean of duplicate gravimetric samples from fiberglas-unit calibration curves was 1.6 percent by weight.

Plaster of paris blocks, fiberglas, and nylon units responded with about equal rapidity to changes of soil moisture but tensiometers were from 40 minutes to 6 hours slower. Plaster of paris blocks were subject to error because of hysteresis and because of the rather high proportion of electrical conduction which followed paths partially outside the block. Soil-moisture records secured with tensiometers also exhibited considerable hysteresis.

The salt concentration of the soil solution did not affect the gravimetric, tensiometer, or nuclear methods but for the electrical units it was the deciding factor as to whether certain types were used. For instance, Ewart and Baver found that increasing the salinity from 0 to 0.2 percent had little or no effect on moisture readings from Bouyoucos blocks. When the concentration was raised to 0.5 percent, drops in resistance were noticeable. Fiberglas units were significantly affected by an increase from 0 to 0.1 percent.

Temperature must be considered in the tensiometric and resistance methods. To secure accurate records with electrical units, resistance must be adjusted to a common temperature. Temperature gradients between the porous cup of the tensiometer and the surrounding soil may cause a variation in tension reading.

In general, then, the authors suggested that the gravimetric method be used unless measurement by one of the indirect methods was absolutely necessary to satisfy study requirements. Electrical resistance instruments were the most useful for securing daily records of considerable duration. If the entire moisture range was to be covered, the fiberglas or nylon-gypsum units were recommended. If temperature corrections were considered necessary, the fiberglas unit should be used. If the soil had been fertilized, increase in salt concentration of the soil solution might prevent satisfactory measurements with electrical units. Tensiometers were more useful in irrigation studies. Elsewhere the soil frequently was too dry for them to be of value. Nuclear instruments were considered to hold promise although sufficient testing had not been accomplished to demonstrate their suitability for routine field work.

Luthin, J. N., and Miller, R. D., 1953, Pressure distribution in soil columns draining into the atmosphere: Soil Sci. Soc. America Proc., v. 17, no. 4, p. 329-333.

The experiments reported in this study were designed to investigate some of the physical phenomena associated with the drainage of water from columns of soil. Measurements were made of the pressure distribution in the soil column during drainage and of the flow of drainage water out of the column. At the conclusion of the experiment the soil columns were sectioned and the moisture content was determined.

Columns 122 centimeters long were assembled by fastening brass cylindrical sections together. Each section was 5.08 centimeters long with an inside diameter of 7.32 centimeters. Special plastic sections were made to accomodate tensiometers. The columns were packed in place by allowing the soil to fall from a funnel into the columns. The columns were tapped to settle the soil. All of the experiments were conducted in a constant temperature room at a temperature of 25°C. Water was ponded on the soil surface to a depth of about 3 millimeters and water was allowed to percolate through the soil column for several hours. The ponded water was then siphoned off the soil surface and the drainage started. Measurements were made of the subsidence of the soil surface and frequent tensiometer readings were taken. The tensiometer readings were made with the aid of a null-point type of tensiometer. When the drainage rate became very slow the soil was removed from the cylinders and the moisture content was determined. The degree of pore saturation was calculated from the data.

Drainage proceeded rapidly in the coarse-textured soils. Water drained out of the upper soil portions until the capillary forces resisting the downward movement of water were sufficient to neutralize the downward forces. The hydraulic gradient approached zero in the lower portions of the soil column and the hydraulic conductivity remained high. The upper portion of the soil column became unsaturated as drainage proceeded. Water continued to move out of the upper zone under the driving action of a high hydraulic gradient, but movement was slow due to a reduced hydraulic conductivity. Static equilibrium was reached when the hydraulic gradient was zero throughout the column.

With some minor exceptions, desorption curves for the soils studied were about the same as the moisture profile. In the drainage experiments the soil moisture was higher at the 10-20 centimeter level than at the bottom of the soil column. The desorption curves indicated a gradual decrease in soil moisture with increasing tension until the air-entry value was reached.

The experimental results emphasized the importance in drainage of the tension required to produce entry of air into the soil surface. The difficulty experienced in the drainage of tight-clay soils might be due more to the high tension required to produce entry of air into the soil than to the low hydraulic conductivity of the soil.

Luthin, J. N. and Worstell, R. V., 1957, The falling water table in tile drainage, a laboratory study: Soil Sci. Soc. America Proc., v. 21, no. 6, p. 580-584.

Several writers have indicated that the equation to be solved in the development of a satisfactory drainage theory is the Richards' equation

$$\frac{dc}{dt} = \Delta \cdot (K\Delta\phi)$$

where

$\frac{dc}{dt}$ is the rate of change of moisture content,

K is the soil hydraulic conductivity,

ϕ is the hydraulic head.

K is a function of the moisture tension in the region above the water table. All the theories of the falling water table which had been developed prior to the writing of this paper had been based on a simplification of the

Richards' equation. In this paper experimental data were presented and used to check the validity of the simplifications.

An experimental tank 6 by 10 by 2 feet was used. Tile drains were placed at a depth of 4½ feet below the top at each end and at the middle of the tank. The tank was filled with dry Oso Flaco fine sand. Several wetting and drainage cycles settled the sand. After settling, tensiometers (Coors' filter candles 762, porosity 5) were inserted in the tank. A null-type tensiometer for making rapid measurement of the hydraulic head was used. The basic component was a fluctuating surgical diaphragm. One side of the diaphragm was in hydraulic contact with the tensiometer and the other side was in contact with a compensating chamber. The air pressure in the compensating chamber. The air pressure in the compensating chamber was controlled by a relay which operated two solenoid valves. The position of the fluctuating diaphragm governed the energizing of the relay. When the relay was not energized a vacuum line was open to the compensating chamber and when the relay was energized a pressure line was open to the compensating chamber. Any change of pressure on the tensiometer side of the diaphragm was automatically compensated for by the same change of air pressure on the other side of the diaphragm. The air pressure required to balance the diaphragm was indicated by a water column connected to the compensating chamber. This device measured the water pressures and tensions in the sand tank to within 1 centimeter of water. Occasional calibration checks were made.

Tensiometric measurements were made during the falling water-table phase. The time at which the ponded water disappeared from the soil surface was called time zero. Periodic readings were made of the rate of outflow from the tile line and of the distribution of hydraulic head. The results obtained from each tensiometer were plotted as a function of time and a curve was fitted visually. This curve was used to determine the hydraulic head at a tensiometer at any time. Although the type of tensiometer used in this experiment was fairly rapid in its operation (15 seconds per reading), it was necessary to repeat the experiment six times to get the desired number of readings. All the measurements were made with only one tile opened at one end of the tank. This arrangement simulated a tile spacing of 20 feet, a tile depth of approximately 4 feet, and an impermeable layer 1½ feet below the tile line.

Although the results of this study were not directly comparable to other studies because of the differences in geometry, the shape of the falling water table, as determined in these experiments, corresponded in a rough way with that determined theoretically by Kirkham, Gaskell, and Childs. The shapes of the experimental water tables were similar to the theoretical shapes, but the rate of fall was not accurately predicted by the theory. The results of this study demonstrated that the assumption used in the theoretical developments of the falling table, that the soil possesses a fixed-drainage pore space led to an incorrect solution. The magnitude of the error would vary from soil to soil depending on the relationship between the soil-moisture tension and the soil moisture. The authors concluded that a complete theory of the falling water table must be based on a solution of Richards' equation. The capillary conductivity is a function of soil-moisture tension, and is continuously changing in the region above the water table as drainage takes place. Because the relationship between the capillary conductivity and the moisture tension varies from soil to soil the shape of the water table is not likely to be independent of the soil hydraulic conductivity. The region above the water table constitutes an important part of the flow region. However, the way in which flow occurs must vary from soil to soil.

Maloney, J. O., 1956, Centrifugation: Indus. and Eng. Chemistry, v. 48, p. 482-485.

The rapid advances in the field of centrifugation were discussed emphasizing the use of the centrifuge in making turbine fuels. The paper contained a large bibliography.

The author discussed several papers submitted at the meeting of the American Institute of Civil Engineers in March 1955. Among these was a paper by H. F. Irving discussing the residual water content obtained by the centrifuge. He noted that the residual moisture in a cake of fixed thickness gave:

$$M = \frac{KA\mu}{(c)^{0.492}(t)^{0.235}}$$

where

M = residual moisture content,

A = crystal surface area per unit mass of crystals,

c = centrifugal force,

K = constant,

μ = viscosity of liquid,

t = drying-spin time,

Marx, J. W., 1956, Determining gravity-drainage characteristics on the centrifuge: Jour. Petroleum Technology, v. 8, no. 4, p. 88-91.

A method was given for predicting the complete gravity-drainage characteristics of long columns from centrifuge drainage measurements on oil-reconstituted core samples. Oil residuals corresponding to hundreds of years of normal gravity depletion could be obtained in a few hours on the centrifuge. The gravity-flow rate at any stage of the depletion process could be determined from the time correlation derived, and could be experimentally checked.

The laboratory prototype column was constructed of a continuous sandstone cylinder, 154 centimeters in length and about 5 centimeters in diameter and its lateral surfaces entirely sealed in lucite tubing. Centrifuge sample cores, cylinders about 3 centimeters in length and 2 centimeters in diameter, were cut from discarded end sections of the parent column. The sandstone was initially oil and moisture free.

The large column was supported vertically by a special pinion rack, which was mounted on a platform balance. When drainage measurements were being made, the receiving vessel was supported by a bridge out of contact with the balance pan. During weighings the liquid level in the receiver was lowered until it broke contact with the bottom of the column.

Temporary caps were sealed over the open ends of the column and the system was evacuated to a constant pressure of about 8 millimeters of mercury for about 75 hours. While maintaining vacuum, about 6 pore volumes of distilled water were drawn through, and the system was allowed to stand for 8 days under excess water. Then an additional pore volume of water was passed through under gravity, aided by 1 or 2 inches of hydrostatic head. The water level was then lowered to the top of the sand and the ratio of the weight increment of the water-saturated system to the dry system was used to calculate the volume of occluded water.

Crude oil was then added to column until no noticeable water was displaced. Reweighing showed that 69 percent of effective pore space was oil.

The oil in the small centrifuge sample cores was then extracted and, after oven-drying, the cores were resaturated under vacuum with distilled water. The cores were then resaturated by centrifuging them at accelerations of about 18,000 gravities while they were immersed in the sand-filtered crude oil. After centrifuging 50 minutes, an average oil saturation of about 70 percent was obtained.

In a graph, the author showed a good correlation of data between the gravity drainage of the prototype column and centrifuge data plotted against a relative time scaled from a newly developed equation.

In principle, the author concluded that the correspondence between centrifuge prediction and actual gravity drainage may be considered to be neither more or less precise than the sampling technique itself. For each small, reconstituted sample, at any given centrifuge acceleration, there will be a corresponding ideal, gravity-drained prototype column, and the centrifuge drainage of the samples can be used to predict gravity drainage of this ideal prototype. The question as to whether this ideal prototype corresponds physically to some actual system of interest to the observer appears to be answered in the affirmative if certain boundary conditions are met. (Modified from author's abstract.)

Meinzer, O. E., 1923, Methods of determining porosity in The occurrence of ground water in the United States: U.S. Geol. Survey Water-Supply Paper 489, p. 11-16.

Several methods have been used to determine the porosity of rocks and soils. These methods differ in the amount of time they require and in the accuracy of the results they produce. Some are adapted only for testing coherent rocks, others only for testing incoherent materials; still others can be used for either coherent or incoherent materials. The different methods are interrelated in various ways but for convenience can be designated as follows:

(1) Measuring the quantity of water required to saturate a known volume of the dry material;

(2) Comparing the volume of a sample with the aggregate volume of its constituent grains;

(3) Comparing the specific gravity of a sample with the weighted average of the known specific gravities of its constituent materials;

(4) Comparing the specific gravity of a dry sample with that of a saturated sample of the same material;

(5) Obtaining the uniformity coefficient and estimating the porosity on the basis of the observed relation between porosity and uniformity coefficient; and

(6) Producing a partial vacuum in a vessel that contains a dry

sample and observing the change in air pressure when this vessel is connected with another that contains air under atmospheric pressure, the volume of each vessel and of the sample being known.

If the specific gravity of water is taken as unity, the following equations express porosity, in percentage, according to the first four methods.

$$P = 100 \left(\frac{W}{V}\right) = 100 \left(\frac{V - v}{V}\right) = 100 \left(\frac{S - a}{S}\right) = 100 \; (b - a)$$

where

P = porosity (by volume),

W = volume of water required to saturate the sample of rock or soil when it is dry,

V = volume of the sample,

v = aggregate volume of the solid particles that constitute the sample,

a = specific gravity of the dry sample,

b = specific gravity of the saturated sample,

The volume (V) of the sample should be determined at the time the sample is taken. It can also be measured in the laboratory directly or with any calibrated vessel. The volume of coherent rock (V) can be obtained by coating the sample with paraffin and then weighing it in air and in water. Its loss of weight in water is the weight of the volume of water it displaced. The aggregate volume of solid particles that comprise the sample (v) can be determined by measuring or weighing the quantity of water that they displace. The weighted average of specific gravities of minerals (S) can be determined by ascertaining the proportions in which the different minerals occur in the sample or by determining the amount of water displaced by a weighed quantity of the solid particles that constitute the sample. The specific gravity of a dry sample of incoherent material (a) can be obtained by weighing a measured volume of the material and dividing the weight by the weight of an equal volume of water. The specific gravity of a saturated sample (b) is equal to the weight of the saturated sample divided by the weight of an equal volume of water.

A rough estimate of the open space in incoherent material can be made from the uniformity coefficient as suggested by Hazen. Sharp-grained materials having uniformity coefficients below 2 have nearly 45-percent open space and sands having coefficients below 3 have 40-percent open space. With more mixed materials the closeness of the packing increases until, with a uniformity coefficient of 6 to 8, only 30 percent of the open space is obtained.

In the sixth method (suggested by Washburn and Bunting and using a glass porosimeter) of determining porosity, air or some other gas is used to fill the interstices and the volume of the interstitial space is determined by observing changes in the gas pressure.

Its great advantage over water saturation methods is the more accurate results it gives, especially for materials that have small and poorly connected interstices. This is due to the perfect expansibility and low viscosity of gases. The essential features of this method are:

An air-tight vessel (B) is connected by a capillary tube to another air-tight vessel (A); a stopcock is installed on each vessel and in the tube that connects the two vessels; a manometer is connected to vessel (A); a known volume of dry sample is placed in the latter vessel, most of the air removed by pumping, and the remaining air observed; the air in the other vessel (B) is at the atmospheric pressure; the stopcock in the connecting tube is then opened and the resulting pressure is observed. After each operation the apparatus must be allowed to come to temperature equilibrium. The porosity can then be determined by the equation,

$$P = 100 \left[\frac{(p_1 - p_3)\, v_1}{(p_3 - p_2)\, v_3} - \frac{v_2 - v_3}{v_3} \right]$$

P = porosity, in percentage by volume,

p_1 = initial pressure in vessel B,

p_1 = atmospheric pressure,

p_2 = initial pressure in vessel A,

p_3 = pressure in both vessels after stopcock between them has been opened,

v_1 = volume of vessel B,

v_2 = volume of vessel A,

v_3 = volume of sample.

Meinzer, O. E., 1932, Outline of methods for estimating ground water supplies: U.S. Geol. Survey Water-Supply Paper 638c.

The author discussed, among other subjects, storage methods and specific yield in relation to storage methods. He indicated seven methods which had been used to determine specific yield:

(1) Saturating samples in the laboratory and allowing them to drain.

(2) Saturating in the field a considerable body of material situated

above the water table and capillary fringe and allowing it to drain naturally.

(3) Collecting samples immediately above the capillary fringe after the water table has declined.

(4) Ascertaining the volume of sediments drained by heavy pumping, a record being kept of the quantity of water that is pumped.

(5) Determining the volume of sediments saturated by a measured amount of seepage from one or more streams.

(6) Making an indirect determination in the laboratory by the application of centrifugal force to small samples.

(7) Making mechanical analyses and determinations of porosity and estimating from them the specific retention and specific yield.

The use of these methods by different workers was discussed briefly.

Messer, E. S., 1951, Interstitial water determination by an evaporation method: Am. Inst. Mining Metall. Petroleum Engineers Trans., v. 192, p. 269-274.

This paper described a method of determining the magnitude of the irreducible interstitial water in a porous medium whereby a saturant in the pore space of a core was allowed to evaporate to an irreducible volume.

The experimental procedure consisted first of saturating a core with water or an appropriate organic liquid that did not leave a residue when evaporated. Then the saturated sample was placed on a balance in an air stream and weight readings were taken at equal intervals until the rate of evaporation became approximately constant. By converting to fluid volume retained, irreducible volume was determined either graphically or by a mathamatical comparison of fluid flows.

The author concluded that the evaporation method satisfied the basic requirements of a production procedure by its simplicity in technique and calculations, and by its reproducible results obtained in a short interval of time. The comparison of results or irreducible water obtained by the capillary pressure method and this method for 110 samples showed a difference of ±1 percent. This method has the advantage that the irreducible saturation can also be determined for samples of varying sizes and shapes, even for core chips.

When highly volatile fluids were used as saturants, a volume factor based on the relative volume of the liquids adsorbed on grain surfaces and retained in pores was calculated.

A theoretical discussion of the calculation of capillary pressure from the evaporation curve was made. The author concluded that further experimentation was needed before a method could be developed to allow the determination of a capillary pressure curve from evaporation data.

Middleton, H. E., 1920, The moisture equivalent in relation to the mechanical analysis of soils: Soil Sci., v. 9, p. 156-167.

The object of this investigation was to determine whether or not the moisture equivalent could be used to assist in the classification of soils and to determine if any relationship existed between the moisture equivalent and the mechanical analysis.

A total of 172 determinations of moisture equivalent and mechanical analysis were made on samples from different states. The determinations were divided into classes based on mechanical analysis. The mean values of moisture equivalent obtained for each class, with the notable exception of silt loam, showed a gradual increase from the coarser-textured classes to the finer-textured classes. The minimum and maximum values for each class varied considerably and overlapped with values for other classes. From these results the author considered it impossible to relate with any agreement the moisture equivalent to classes of the currently used soil classification.

Briggs and Shantz had earlier derived the following formula for calculating the moisture equivalent from the mechanical analysis:

Moisture equivalent = 0.02 sand + 0.22 silt + 1.05 clay.

This formula was applied to several of the samples and found to give results which were considerably higher than the observed values. This was probably due to the high coefficient used for clay. An analysis of the data in this study gave the following equation:

Moisture equivalent = 0.063 sand + 0.291 silt + 0.426 clay.

The determinations were then divided into three groups and an equation was derived for each group:

Group A soils containing less than 20 percent silt and clay,

Moisture equivalent = 0.02 sand + 0.40 silt + 0.53 clay;

Group B soils containing 20 - 50 percent silt and clay;

Moisture equivalent = 0.02 sand + 0.35 silt + 0.63 clay;

Group C soils containing more than 50-percent silt and clay,

Moisture equivalent = 0.11 sand + 0.27 silt + 0.40 clay.

Groups B and C were further subdivided into textural classes and an equation derived for each class.

These formulae were tested by taking 34 samples at random and comparing the moisture equivalent with the calculated moisture equivalent. The probable error of a single determination varied from ±5.44 to ±5.65 depending on the formula used.

A study of the effect of organic matter upon the moisture equivalent indicated that the organic matter has practically the same effect as clay.

Miller, R. D., 1951, A technique for measuring soil-moisture tensions in rapidly changing systems: Soil Sci., v. 72, p. 291-301.

The usefulness of the conventional tensiometer in rapidly changing systems is impaired because an appreciable quantity of water must enter or leave the tensiometer as it responds to a change in the external tension. This transfer not only causes lag but influences the soil-moisture content.

A new form of tensiometer, which makes it possible to eliminate lag and to lessen materially the amount of water transferred, has been developed. By using a manually controlled water manometer and a sensitive-null indicator, the operator was able to recognize the correct manometer setting. The accuracy of this null method was considered adequate for most purposes, but further improvement in the quality of tensiometric measurements in rapidly changing systems could be accomplished by a dynamic method of measurement with similar apparatus.

Miller, W. M., 1953, A pressure distribution panel for soil-moisture investigations: Agr. Eng., v. 34, p. 104-106.

A pressure-distribution panel was described whereby a group of pressure-membrane apparatus, or other equipment requiring the maintainance of a supply of air at constant pressure, may be independently connected or disconnected from any 1 of 4 pressure regulators, regardless of the simultaneous use of them by other apparatus and without disconnecting any fittings.

Moore, R. E., 1939, Water conduction of shallow water tables: Hilgardia, v. 12, no. 6, p. 383-426.

The introduction of the potential function gave rise to the dynamic method in the study of unsaturated flow. This paper reported pressure potential and saturated and unsaturated permeability data obtained under laboratory conditions.

Six California soils, ranging in texture from sand to clay, were packed gradually into metal cylinders 8 inches in diameter and 3 to 4 feet in length, by rotating a Tremie tube. The soils were supported in the columns by wire-screen diaphrams soldered 2 inches above the column bases, and a water inlet one-quarter inch in diameter was inserted 1 inch above the bases. Four rows of holes five-eights inch in diameter were punched through the metal cylinder to give one hole for each inch of cylinder height. These were closed with patches of celluloid. Water flowed upward through the columns under positive hydrostatic pressures in the saturated zone and under negative hydrostatic pressures, or by capillarity, from the water table to the surface of the columns where it was removed by evaporation. Tensiometers consisting of porous fired-clay elements connected to vacuum gauges of the manometer type were mounted in the soil columns at intervals of 10 centimeters. When the rate of water uptake and pressure potential throughout the soil column became steady, the soil was sampled for moisture content.

The columns were then drained, maintaining an air-water interface at their bases. During drainage the change in pressure potential with time was recorded.

The relation between pressure potential and moisture content was presented graphically. Hysteresis was found for all soils, according to whether they were in a wetting or drying cycle. The pressure potential also depended upon the range through which the soil had dried. In an unsaturated soil at constant moisture content, the pressure potential was found to increase with increasing temperature and the amount of water held in the soil at a given pressure potential decreased with increasing temperature. The rate of water uptake from a high water table changed markedly during periods of temperature change.

The pressure potential of the water films in the soil was a measure of the curvature of the films and an index of the degree to which the soil was saturated. The relation of soil permeability, K, to pressure potential was studied and curves were developed. Permeability was a maximum at or near saturation and decreased rapidly with decreasing moisture content to approximately the moisture equivalent, at which moisture content the permeability was very low. From this point permeability remained constant or decreased only slightly with further decreases in moisture content. Here $\frac{dK}{d\psi} = 0$ (approximately) (where K = permeability and ψ = pressure potential). The moisture content where $\frac{dK}{d\psi} = 0$ was also approximately that of the wetted front generated as the water advanced upward through the dry soil above the water table. These two criteria, the moisture content at which $\frac{dK}{d\psi} = 0$ and the moisture content of the wetting front, were interpreted as the moisture content at which the moisture films in the soil became discontinuous and at which the capillary permeability of the soil was zero.

Knowledge of the variation of permeability with texture, especially the pressure potential at which capillary permeability was approximately zero, was considered fundamental to the relative rates and maximum height of capillary rise, to field capacity in stratified and nonstratified soils, and to single-valued "constants" such as moisture equivalent, the field capacity, and the normal moisture-holding capacity. These experiments indicated that the pressure potential of soil moisture at these constants changed with the texture of the soil. The constants were at the approximate pressure potential on curves of $K = f(\psi)$ at which the moisture films in the soil became discontinuous and the capillary permeability became zero.

Moore, R. E., 1941, The relation of soil temperature to soil moisture, pressure potential, retention, and infiltration rate: Soil Sci. Soc. America Proc., 1940, v. 5, p. 61-64.

Experiments have demonstrated that temperature has considerable influence on such soil-moisture properties as pressure potential, retention, and infiltration rate. The relation of temperature to the retention and distribution of water in soils after irrigation was investigated by irrigating columns of air-dry soil and by determining their moisture content for each inch of depth. The soils were irrigated in a constant temperature bath with sufficient water being added to wet them to a depth of about 14 inches. In general, the amount of water retained decreased with increasing temperature. There was approximately 18 percent more water retained at 10°C than at 45°C. A second maximum of retention occurred in the temperature range of 20 to 25°C.

Mortland, M. M. 1954, Specific surface and its relationships to some physical and chemical properties of soil: Soil Sci., v. 78, p. 343-347.

The relationships between certain physical and chemical properties and the specific surfaces of soils were studied. The purpose was to determine whether measurements of specific surface would be a useful index to these properties and whether the measurements were more closely related to total, external, or internal specific surface. Specific surface, determined by the ethylene-glycol retention method, was found to be very closely related to cation-exchange capacity, percentage of 2µ clay, and moisture retention. Total specific surface was more closely related to cation-exchange capacity than was 2µ clay content. From 24 soil samples studied, a fairly close estimation of cation-exchange capacity could be made from a total surface area measurement. The reverse was also possible. For these samples the mean area per exchange position was about 95 A^2 (Angstrom units). The separation of internal and external surface provided an insight into the locale of physical and chemical reactions of soil-particle surfaces. The study indicated that soil specific surface as determined by ethylene-glycol retention was a valuable index to some chemical and physical properties of soil.

Philip, J. R., and DeVries, D. A., 1947, Moisture movement in porous materials under temperature gradients: Am. Geophys. Union Trans., v. 38, no. 2, p. 222-232.

A theory of moisture movement in porous materials under temperature gradients was developed which reconciled a growing body of experimental information including the large value of the apparent vapor transfer, the effect of moisture content on net moisture transfer, and the transfer of latent heat by distillation.

The simple theory of water vapor diffusion in porous media under temperature gradients neglected the interaction of vapor, liquid, and solid phases, and the difference between average temperature gradient in the air-filled pores and in the soil. By considering these factors, an approximate analysis was developed which predicted orders of magnitude and general behavior in satisfactory agreement with the experimental facts.

An important implication of this approach was that appropriate experimental methods have not distinguished between liquid and vapor transfer because series-parallel flow through liquid "islands" located in a vapor continuum has been confused with vapor transfer.

Equations describing moisture and heat transfer in porous materials under combined moisture and temperature gradients were developed. Four timely moisture-dependent diffusivities were discussed briefly. (Modified from author's abstract.)

Pinckney, R. M., and Alway, F. J., 1939, Reliability of the proposed suction method of determining the moisture equivalent of soils: Soil. Sci., v. 48, p. 403-411.

To compare moisture equivalent as obtained by the Bouyoucos suction method with the moisture equivalent as determined with the Briggs-McLane centrifuge, 113 Minnesota soils were used. These soils contained from 0.001- to 2.03-percent nitrogen and ranged in pH from 4.7 to 8.1. The centrifuge moisture equivalents ranged from 1.5 to 90.6. Although the relationships found were in general agreement with those reported by Bouyoucos, they did not fully support his conclusions as to the reliability and general desirability of the method. With loams and soils of finer texture, the suction moisture equivalent of the individual soils ranged from practically equal to one-third higher than the centrifuge moisture equivalent and averaged about one-tenth higher. With sands of the coarsest texture the suction moisture equivalent was at least twice as high, and with intermediate soils it varied widely but in general was between the coarse and the fine values. Duplicate determinations by the suction method were found to be much less concordant than those with the centrifuge. The authors concluded that the use of the suction method was only desirable where a moisture-equivalent centrifuge was not available, and that the values so obtained should be referred to by some other designation than moisture equivalent.

Piper, A. M., 1933, Notes on the relation between the moisture equivalent and the specific retention of water-bearing materials: Am. Geophys. Union Trans., v. 14, p. 481-487.

The centrifuge moisture equivalent and specific retention have been widely used as quasiphysical constants to measure the power of a water-bearing material or a soil to retain water against gravitative force. This paper was a preliminary statement of the apparent relation between these two measures of retentive power and was based on a study of alluvial fans materials made in connection with an investigation by the U.S. Geological Survey into the ground-water resources of the Mokelumne area, central California.

The following procedure was used to obtain specific retention for this paper. Each of 16 samples were taken by driving vertically for a distance of 12 inches into the material exposed in a dug pit a steel cylinder 6 inches in diameter by 15 inches in length. The material surrounding the cylinders was saved. Of the 16 pairs of undisturbed samples, 13 were taken from 1 pit, thus giving 2 columns of material from the ground surface to 13 feet; the remaining 3 were taken from separate pits. The samples were then placed in an earth-covered observation pit dug in the sandy subsoil at Lodi, Calif. There on benches in the wall of the pit, bottomless wooden bins 6 inches deep were constructed and filled with additional portions of the 16 samples properly compacted. On each bin so-filled was placed the corresponding pair of cylinders screened at the bottom with 14-mesh copper screencloth. The atmosphere of the pit was kept approximately saturated. To the top of each sample was then added water equivalent to about 25 percent of its volume. Each sample was allowed to drain about 10 days and then wetted again by placing the cylinder in a shallow pan of water and allowing it to saturate from the bottom for 24 to 96 hours. After the sample would no longer absorb water, it was returned on its bin and allowed to drain, the top of the cylinder being sealed with wax paper pierced by only one perforation. The rate at which the sample drained was determined by weighing periodically over a 96-day period. The moisture contents of 3 axial sub-samples (from 1 cylinder of each pair taken from the top, middle, and bottom thirds of a sampling tube 0.9 inch in diameter) were then determined along with the moisture content of the gross sample. Subsequently, the moisture equivalent was obtained on the oven-dried sub-samples and the gross sample. The moisture content of the gross samples of the 16 samples exceeded the moisture content of the corresponding 3 sub-samples taken from the cylinders, possibly due to unequal distribution of the water throughout the undisturbed samples. The remaining set of the initial pairs of samples was allowed to drain for approximately 1 year. Moisture contents and moisture equivalents were determined as before. Incomplete data suggested that, after drainage for 1 year, the retained moisture was much more uniformly distributed but the moisture content of the whole sample differed but little from that of the axial sub-samples taken after 96 days of drainage.

The experimental conditions maintained in the investigation at Lodi have satisfied rather effectively the theoretical requirements for a determination of the specific retention. The specific retention of the alluvial materials studied was measured approximately by the moisture content of the sub-samples from the axis of the cylinders after 96 days of drainage.

Piper, A. M., Gale, H. S., Thomas, H. E., and Robinson, T. W., 1939, Geology and ground-water hydrology of the Mokelumne area, California: U.S. Geol. Survey Water-Supply Paper 780, p. 101-122.

Two methods used to determine the specific yield of materials within the zone of water-table fluctuations were: (1) volumetric method--measuring the volume of material saturated and unwatered when measured volumes of water were added or withdrawn from columns of undisturbed soil; and (2) drainage method--draining undisturbed samples for periods as long as 390 days and determining the difference between the porosity and resultant specific retention.

The volumetric method utilized was first described in detail by Stearns and others in 1930. The authors used the same tests but made a number of modifications. Forty-two-inch or longer cylinders were recommended to accomodate the capillary fringe. The sample cylinders were confined between a top and bottom plate bolted together with tie rods. The cylinders were then inverted, the base plate removed, and the end of the sample allowed to dry. The base plate could thereby be soldered firmly in place and leaks thus prevented. The cylinders were then inverted to their normal position for the rest of the test. Because of the effects of temperature it was necessary to move the cylinders from the sampling sites to a covered and insulated observation pit where the temperature range could be maintained within 2°F each day, or 30°F for the year. Temperatures were measured by thermometers lowered into an observation well in each cylinder or by a water-tank thermometer tapped through the wall of the cylinder. Evaporation from the cylinders was minimized by a top seal of tar paper and tar perforated with seveal small holes so air could move in and out as water was added or withdrawn. On subsequent cylinders only one 1½-inch hole was bored in each cylinder for installation of an observation well. This well was perforated only in the lowermost 6 inches. Depths to water were measured in the observation well by means of a miro-hydro-gage, and electrical device for measuring depths to a thousandth of a foot. Instead of the second observation well used formerly, a plumber's water-gage fitting was installed near the bottom of the cylinders. A layer of screened sand was suggested for the bottom of each cylinder to speed the equilibrium condition of a new water table, especially in fine-grained materials. In these tests, no less than 21 days was allowed for a new water level to attain equilibrium, some tests being continued for as long as 220 days.

In the volumetric method the water levels were influenced by both temperature and barometric pressure. Rising temperature caused the water level to rise, but increasing barometric pressure caused the water level to fall. The observed water levels were adjusted to 60°F by temperature coefficients derived from the plotting of water levels against temperatures for observations made when the barometric pressure was within the range

29.9, ±0.04 inches of mercury. Some temperature coefficients ranged from 0.002 to 0.008 foot for each degree Fahrenheit. After the temperature coefficients were determined, it was then possible to adjust all water levels to the standard of 30 inches of mercury barometric pressure. The longer-term (220 days) equilibrium periods increased the specific yield only by 1 and 3 percent of the results derived from the shorter-term equilibrium periods. For fine materials, specific yield by saturation was about 3 times as large as that obtained by unwatering. If the columns were too short to contain all of the capillary fringe, specific yield tended to be too small.

In another paper, Piper (1933) described the drainage method in detail. Duplicate 6-inch-diameter by 12-inch-long cylinders of undisturbed material were saturated and allowed to drain on benches of material of the same type in an insulated pit. One set of samples was drained for 96 to 111 days and the second set was drained for 322 to 390 days. At the end of the drainage period the specific retention of each sample was obtained by determining the retained moisture content. Moisture equivalent tests were made on the same samples and a graph was prepared relating the moisture equivalent to the specific retention.

The longer-term drainage tests showed that the quantity of retained water decreased steadily with time though at a diminishing rate. The retention at the end of the long period tests ranged between 61 and 97 percent of the respective quantities retained after the shorter drainage period. In general, the greater additional drainage came from the coarser-grained samples. A graph showed the relation between the specific yield (drainage method) and the mean size of sample particles.

The percentage increase in specific yield derived from the longer-term drainage was greater than the increase in long-term tests by the volumetric method. The volumetric method tended to yield results that were too small, whether the test was by unwatering or by saturation. Average values of specific yield for the two test methods were: gravel and coarse sand, 34.8 percent; medium and fine sand, 24.2 percent; and very fine sand, silt, and clay, 4.2 percent. These values were correlated with drillers records of wells and the data were used to determine the specific yield of two zones of water-level fluctuation.

Prescott, J. A., and Poole, H. G., 1934, The relationships between sticky point, moisture equivalent, and mechanical analysis in some Australian soils: Jour. Agr. Sci., v. 24, p. 1-14.

The inter-relationships between sticky point (that moisture content at which a soil no longer sticks to a foreign object), mechanical analysis, and moisture equivalent, were investigated. The technique for the determination of the moisture equivalent included the use of 30 grams of soil centrifuged at 1,000 times the force of gravity for a 30-minute period. Some soil samples did not reach equilibrium at the end of the centrifuge period. This was indicated by either free water on the surface of the sample or by the upper layers of sample being wetter and softer to the touch than the lower layers. For those soils that were of particular interest, a dilution method was employed. By mixing the soil with a specified percentage of sand, results were obtained which could be readily extrapolated to obtain a reasonable approximation of the moisture equivalent.

In general, the values for sticky point were not found to be closely related to moisture equivalent. Although the moisture-equivalent values for normal calcium-hydrogen clays were probably very close to the sticky points, the organic matter appeared to contribute more to the sticky point than to the moisture equivalent. Results of this study indicated that an excellent multiple correlation between silt, clay, and organic matter and the moisture equivalent was likely to be found if the statistical examination was restricted to closely related groups of soils.

Variation in dry content affected moisture-equivalent results more than variation in organic matter. Discrepancies between observed and calculated values generally could be related to the clay fraction. Moisture-equivalent values were high for soils with appreciable quantities of exchangeable sodium. The authors suggested that a standard technique might be developed in which soils high in exchangeable sodium could be treated with calcium-chloride solution and subsequently washed prior to the moisture-equivalent determination.

Purcell, W. R., 1949, Capillary pressures--their measurement using mercury and the calculation of permeability therefrom: Am. Inst. Mining Metall. Petroleum Engineers Trans., v. 186, p. 39-48.

A method and procedure for determining capillary-pressure curves for porous media was discussed in which mercury was forced under pressure into the evacuated pores of the solids.

The essential components of the mercury injection apparatus were a mercury displacement pump, a sample holder, and a manifold system wherein the gas pressure could be controlled from high vacuum to 2,000 pounds per square inch. Saturation equilibriums were established rapidly over this range so that an entire saturation-capillary-pressure curve was determined in 30 to 60 minutes. Temperature fluctuations over this period were considered unimportant.

The pressure-volume relationships were reasonably similar to capillary-pressure curves determined by the generally accepted porous-diaphragm method. However, the advantages of this method were the rapidity with which the experimental data were obtained and their application to analysis of drill cuttings and cores.

The author showed in graph form the relatively close relationship between mercury-capillary-pressure curves and water and air curves. He also derived an equation which indicated theoretically the relationship between the permeability of a porous medium and its capillary-pressure curves and showed experimentally how the equation applied.

Rapoport, L. A., and Leas, W. J., 1951, Relative permeability to liquid in liquid-gas systems: Am. Inst. Mining Metall. Engineers Trans., v. 192, p. 83-98.

The purpose of this paper was to demonstrate the general validity and comparative simplicity of an experimental method to measure relative permeability of cores to liquid under conditions of parallel flow and to relate the method to a theoretical analysis. Capillary equilibrium was maintained between the liquid and gas throughout the entire core specimen. In the method presented, capillary pressure between the two phases was established in a relative-permeability cell by means of semipermeable barriers and was controlled by pressure placed on the gas. The elimination of "end-effects" was obtained by cementing permeable barriers to the end of the core sample.

The procedure consisted first of completely saturating the sample, then completely forcing liquid from the cell through end plates by gas, and finally again flowing liquid through the system with the gas held back by the barriers. When steady-state conditions were reached, the pressure drop in the liquid over the length of core and the rate of flow were determined. The effective, and consequently, the relative permeability were then obtained and plotted against the corresponding saturation.

Close agreement was found between measured and calculated relative-permeability data. In addition, a procedure for calculating limiting relative permeability from capillary-pressure data was established and the possibility of applying the calculated curves to connate-water saturations was indicated. The use of liquid curves of limiting relative permeability to calculated from capillary-pressure data was suggested for general engineering purposes. Their use for reservoir studies was particularly recommended in conjunction with the method for measuring relative permeability to gas which simultaneously yielded the capillary-pressure data necessary for the calculations.

Richards, L. A., 1931, Capillary conduction of liquids through porous mediums: Physics, v. 1, p. 318-333.

This paper derived a general theory of capillary flow in terms of known or measurable functions. A brief discussion of the author's theory is given in the following:

The forces acting in the boundary surfaces of liquids directly cause all capillary phenomena and have their origin in the cohesive and adhesive attractions which are exerted between molecules. Capillary action occurs commonly when one surface of a liquid is in contact with air. It is more convenient to express the physical properties of this air-liquid surface in terms of surface tension. Because of the inequality between cohesive and adhesive forces at a liquid-solid interface, these surfaces also have a free energy which is important in determining the capillary behavior of liquids. Because of their effect on the angle of contact, adhesive forces are directly involved in an initial wetting process such as the spreading of a liquid in a dry porous medium, but after the medium is wetted, adhesive forces are no longer effective in producing a motion of the liquid and influence capillary action only in proportion to the force with which they hold a thin film firmly in contact with the solid surface. The liquid lying outside the adsorbed films is free to move under the action of unbalanced forces. Capillary flow may be distinguished from other kinds of liquid flow only because of the relation of surface tension and curvature to the pressure and to the effective cross-sectional area of the liquid transmitting region. If there is a steady flow of liquid through a porous medium which is only partially saturated, then the larger pore spaces contain air and the effective cross-sectional area of the water conducting region is reduced. If these air spaces could in some way be filled with solid, the condition of the flow would be unchanged and the proportionality between the flow and the water-moving force would still hold because Darcy's law is independent of the size of particles or state of packing. Hence, the essential difference between flow through a porous medium which is saturated and flow through a medium which is unsaturated is that in the latter the pressure is determined by capillary forces and the conductivity depends on the moisture content of the medium. The motion of liquids through unsaturated porous mediums is referred to as capillary flow and K here is called capillary conductivity. The author then developed a flow equation using the following terms:

K = capillary conductivity in cubic centimeters of water crossing 1 square centimeter perpendicular to the flow in 1 second when there is a water-moving force of 1 dyne per gram,

ψ = capillary potential (pressure forces determined by surface tension and curvature of air-liquid interface) expressed in ergs per gram,

ρ_s = weight of dry medium in unit volume,

A = rate of change of the moisture content with respect to the capillary potential (capillary capacity),

g = gravity,

q = volume of water crossing unit area perpendicular to the flow.

A discussion of this equation is taken directly from the author's abstract as follows:

The flow of liquids in unsaturated porous media follows the ordinary laws of hydrodynamics, the motion being produced by gravity and the pressure-gradient force acting in the liquid. By making use of Darcy's law, that flow is proportional to the forces producing flow, the equation

$$K\Delta^2\psi + \Delta K \cdot \Delta\psi + g\frac{dK}{dz} = -\rho_s A\frac{d\psi}{dt}$$

may be derived for the capillary conduction of liquids in porous media. It is possible experimentally to determine the capillary potential $\psi = \int \frac{dp}{\rho}$, the capillary conductivity K, which is defined by the flow equation of $q = K(g - \Delta\psi)$, and the capillary capacity A, which is the rate of change of liquid content of the medium with respect to ψ. These variables are analogous, respectively, to the temperature, thermal conductivity, and thermal capacity in the case of heat flow. Data are presented and application of the equations is made for the capillary conduction of water through soil and clay but the mathematical formulations and the experimental methods developed may be used to express capillary flow for other liquids and media. The possible existance of a hysteresis effect between the capillary potential and moisture content of a porous medium is considered.

Richards, L. A., 1948, Porous plate apparatus for measuring moisture retention and transmission by soil: Soil Sci., v. 66, p. 106-110.

An improved method for mounting and using porous plates for measuring the relation between soil-moisture tension and the moisture content of soils was discussed and two pressure-control systems, a water tower for pressures from 0 to 100 centimeters of water, and a mercury tower for pressures from 100 to 1,000 centimeters of water, were used in connection with the porous plates described.

The porous plates were arranged for installation in a conventional domestic pressure cooker so that various soil-moisture equilibria in the one atmosphere range could be attained by application of air pressure to a chamber containing the plates. The porous-plate mounting was equally suitable for connection to vacuum pressure. The soil samples, usually contained in metal rings, were spaced on the porous plate. Free water was maintained on the plate surface until the soils were thoroughly wetted. The plate assembly with soil was then transferred to the pressure cooker.

Richards, L. A., 1949, Methods of measuring soil-moisture tension: Soil Sci., v. 68, p. 95-109.

The author described the soil-moisture tensiometer, a combination of a porous cup and a vacuum gauge, which can be used for measuring soil-moisture tension values up to about 0.85 atmosphere. Numerous papers dealing with the development and construction of tensiometers and reporting the experience of various workers with tensiometers under field conditions were listed. References were also cited which relate to the use of porous-clay and cellulose-membrane apparatus for measuring the uptake and release of moisture by soils at various moisture-tension values. Typical moisture-retention curves for many soils were given at soil-moisture tensions up to 100 atmospheres. Methods for making moisture-retention measurements at known soil-moisture tension were reviewed, and the relation of such measurements to the field-moisture properties of soils were briefly discussed.

Richards, L. A., 1950, Experimental demonstration of the hydraulic criterion for zero flow of water in unsaturated soil: Internat. Cong. Soil Sci. Trans., v. 1, p. 66-68.

An experimental demonstration of the hydraulic criterion for zero flow in unsaturated soil was reported. A metal can, 31 centimeters in diameter and 100 centimeters in height, was filled with soil and sealed against evaporative loss. The soil air in the sealed can was equalized to atmospheric pressure. Five tensiometer tubes, 21.6 centimeters apart, were soldered to the side of the can and were connected by flexible rubber tubing to water manometers. The can was filled with Pachappa loam, a soil that has a wilting percentage of 6 percent and a field capacity of 18-percent moisture. Sufficient water was applied to the soil to bring the moisture content to 18 percent. The metal can was then set in a vertical position. In the absence of tension gradients, gravity produced a downward movement of water and after 4½ months the overall hydraulic gradient between extreme tensiometer cups was near zero or 0.015. From the Darcy equation V = Ki, the transmission velocity "V" for the static equilibrium of soil moisture under gravity, was equal to zero when the hydraulic gradient "i" was equal to zero. The metal can was next tipped from the vertical to the horizontal position. Immediately upon rotation of the can the hydraulic-head difference between cups was approximately 21.6 centimeters. In 53 days the overall hydraulic-head difference between extreme cups was 2.4 centimeters. By never reversing the can again to the vertical position, a pattern in potential difference was obtained that was just the reverse of that for the horizontal position. In time, the meniscus pattern approached the pattern in the original vertical position. (Modified from author's abstract.)

Richards, L. A., 1954, Multiple tensiometer for determining the vertical component of the hydraulic gradient in soil: Soil Sci. Soc. America Proc., v. 18, p. 7-10.

An instrument for measuring the hydraulic head and hydraulic gradient of water in unsaturated field soil was described. Porous ceramic sections were mounted between plastic spacers to make a rod-shaped instrument for insertion into a hole made by the Veihmeyer soil sampling tube. Mercury manometers mounted at the top of the assembly were connected to each of 5 porous sections spaced 10 centimeters apart vertically.

After 44 centimeters of water had infiltrated a uniform, fine sand loam, the downward hydraulic gradient averaged 1.3 in the depth interval of 10 to 30 centimeters. This corresponded to a downward water-moving force of 1.3 times gravity. Six days later, a short time after a 1-centimeter rain, the average downward water-moving force in the same depth interval was five times gravity. Four days after the rain, a net upward water-moving force of 36 times gravity occurred in the depth interval of 10 to 20 centimeters, owing to the influence of surface evaporation. A much higher hydraulic gradient existed in the 0- to 10-centimeter layer because of the greater moisture gradient near the soil surface.

In describing the soil-water system, the term static zone was used to designate the locus of points above which water movement was upward and below which water movement was downward. A static zone passed downward through a fallow soil after wetting. Over a 4-day period of warm dry weather after a heavy irrigation, the static zone passed downward in a fallow fine sandy loam at an average rate of 6 centimeters per day.

Richards, L. A., and Fireman, Milton, 1943, Pressure-plate apparatus for measuring moisture sorption and transmission by soils: Soil Sci., v. 56, p. 395-404.

This paper described an apparatus for obtaining sorption data in the range of 0 to 2,000 centimeters of water tension, and discussed the effect of the history and treatment of the soil sample on the moisture retention as determined with this apparatus.

The apparatus consisted basically of a porous ceramic plate enclosed in an air-tight chamber. A porous plate was selected which would withstand at least two atmospheres of air pressure before leaking air.

Data obtained by this apparatus agreed with data obtained by the suction method. Packing of the dry sample by ordinary hand methods had little effect on the moisture retention at the one-third atmosphere tension. Moisture-retention values obtained on disturbed and undisturbed samples of Fallbrook loam were compared. Any manipulation of the soil, such as breaking up the cores or sieving the soil, in either the wet or the dry state, increased the moisture retention significantly throughout the 1-atmosphere tension range. Wet cores that were gently broken by hand had moisture contents in the lower tension range intermediate between the undisturbed and the sieved cores, but retained the same amount of water as the dried and sieved soil in the higher tension range. The wetting of air-dried and sieved soil with 100 milliliters of sodium chloride at concentrations of 0, 0.4, 0.8, 1.6, 3.2, and 6.4 percent had little effect on the moisture retention of Las Flores soil. Cajon silty clay loam was tested in a similar manner with solutions of 0, 0.8, 3.2 and 12.8-percent calcium chloride. An increase in concentration of calcium chloride caused a decrease in moisture retention. Calcium saturation had only a slight effect on the moisture retention of either the puddled or the unpuddled soils. Sodium saturation, however, had a marked effect, increasing the moisture retention for both the puddled and the unpuddled soils. Puddling decreased the moisture retention of the unsalted soil at all tensions below 640 centimeter of water, but increased the moisture retention slightly at tensions higher than 640 centimeters of water. Puddling decreased the moisture retention of the sodium-treated soil throughout the 1-atmosphere tension range.

Richards, L. A., and Gardner, Willard, 1936, Tensiometers for measuring the capillary tension of soil water: Jour. Am. Soc. Agronomy Proc., v. 28, p. 352-358.

"Soil pull," the security with which water is held by soil, was considered as capillary potential, capillary pressure, suction pressure, pressure deficiency, tension, or capillary tension (negative pressure). It was measured by the combined use of a porous cell and a vacuum gauge. The range of capillary tension for which tensiometers could be used did not exceed one atmosphere. This was but one-sixteenth of the tension range in going from a saturated soil to a soil at the wilting point.

Richards, L. A., and Moore, D. C., 1952, Influence of capillary conductivity and dept of wetting on moisture retention in soil: Am. Geophys. Union Trans., v. 33, no. 4, p. 531-540.

This paper pertained to those water relations in soil where an approximate continuous phase of gas existed. It contained a summary of capillary-conductivity measurements made by many workers which indicated that as the soil-moisture tension increased to about 25 centimeters of water, the capillary conductivity dropped rapidly. The capillary conductivity for the soils summarized was reduced to a relatively low value before the tension reached 150 centimeters of water. The authors stated that the shape of the capillary-conductivity curve was related directly to the moisture-storage and transmitting properties of soils in the field. The usual definition for the field capacity, which is applied for deep, well-drained soils, might be paraphrased as follows: Field capacity is the moisture content of soil 2 or 3 days after a heavy rain or irrigation when downward drainage has so reduced the moisture content of the soil and the thickness of the moisture films that the capillary conductivity is reduced to a low residual value and additional downward drainage of water is negligible. This view was further supported by data indicating that the tension range (from 30 to 150 centimeters of water) over which the capillary conductivity dropped to a low residual value was also the range over which tensiometers were observed to read the moisture condition at field capacity for the soils summarized.

Capillary-conductivity measurements were made for six soils. The procedure used for these measurements represented a new development. A soil sample approximately 6 centimeters in height and 25 centimeters in diameter was mounted between two porous plates. These plates were backed with a bronze screen and sheet rubber on the side opposite the soil contact. The soil column was supported by a metal cylinder. This assembly was placed in an air-tight pressure chamber. Water was supplied to the soil column from a constant-pressure source through the upper porous plate. Water that passed through the soil column and the lower porous plate dripped into a burette. Small porous tensiometers were installed at each end of the soil column to measure the net hydraulic head-loss as water moved through the soil column. The amount of air pressure applied to the pressure chamber determined the soil-moisture tension and the corresponding moisture content of the soil. Capillary conductivity at any desired moisture content could be determined by applying the necessary air pressure to the pressure chamber and then by measuring the water that flowed through the soil column from the upper porous plate through the lower porous plate.

Capillary conductivity values for the six soils were in harmony with the summarized data. A comparatively rapid reduction to low-conductivity values was observed as the tension was increased. Measurements at all tensions gave capillary-conductivity values measurably greater than zero. The sandy soils had higher conductivity at low tension and lower conductivity at high tension than the medium- or fine-textured soils.

A method was used where a small predetermined volume of water was added to air-dried soil. After a certain length of time the moisture content of the wetted soil was determined. The volume of water applied, or depth of wetting, greatly influenced the rate of change of the moisture content in the wetted part of the soil. The moisture content of a part wetted with about 0.1 milliliter of water dropped below the 15-atmosphere percentage within about 1 to 2 hours, whereas the moisture content of a part wetted with 20 milliliters of water remained at much higher value for a much longer time.

The authors believed that field capacity generally represented a moisture content in the field soil that changed slowly with time because of the dynamic moisture transmitting properties of the soil. If the factors that control the rate of transfer of water through the wetting zone tend to make this rate of transfer somewhat constant, then the thicker the zone of wetted soil the more slowly would the moisture content change with time. The authors stated that field capacity likely corresponds to that tension and moisture content at which conductivity becomes negligible. Measuring techniques were believed to be approaching the stage where moisture retention and conductivity in soils having undisturbed structure could be expressed as functions of tension. Improved measuring techniques would make possible a theoretical solution for the soil-wetting process in the field and extend the application of basic principles in the soil-moisture system. However, one complicating factor in this application would be that the hydraulic gradient during infiltration in the field varies considerably between soils. A substantially undisturbed sample might be brought to the laboratory for measurement and study but the boundary conditions that control the hydraulic gradient in the field might not be conveniently reproduced.

Richards, L. A., and Ogata, Gen, 1956, Materials for retainer plates and their use for retentivity measurements: Soil Sci. Soc. America Proc., v. 20, no. 3, p. 303-306.

Ceramic materials suitable for use as sample retainer plates and methods for measuring their permeability, air-bubbling pressure, and air-entry pressure were discussed. The transfer conductance between retainers and a ceramic suction-control surface was believed to be satisfactory if the contacting surfaces were reasonably flat and clean. Several tests had consistently indicated, however, that better results were obtained if a capillary contact medium was used when ceramic retainer plates were used on cellulose membranes. At high suctions the retentivity for cores might be significantly different than for air-dried and fragmented soil. For example, the authors found that the average 15-bar percentage for 7 cores of sandy loam soil was 5.96, whereas the average 15-bar percentage for the soil in these cores after air drying and fragmenting was 5.48. Tests indicated that the water-release curves for cores of a fine, sandy loam soil that were wetted in the field by flood irrigation were similar to those same cores wetted to zero suction in the laboratory. (Modified from authors' abstract.)

Richards, L. A., and Weaver, L. R., 1944, Moisture retention by some irrigated soils as related to soil-moisture tension: Jour. Agr. Research, v. 69, no. 6, p. 215-235.

The main objective of this paper was to present data on the relation between the equivalent negative pressure or tension in the soil water and the moisture content of 71 samples.

The pressure-membrane apparatus was used to obtain moisture data at tensions above one atmosphere.

Moisture-retention data at tensions between zero and one atmosphere were obtained either with pressure-plate or suction-plate apparatus.

The procedure for obtaining all of the moisture-retention data was as follows: A layer of screened, air-dried soil was placed on the porous moisture-extracting surface, the soil was wet thoroughly with an excess of distilled water, and then the moisture was extracted until the moisture tension in the soil increased to a constant predetermined value and moisture outflow from the sample ceased.

The amount of moisture a soil will retain at a given tension depends somewhat on the time allowed for wetting the air-dried sample. No increase in moisture retention for a wetting time beyond 15 minutes is observed for some sandy soils, but some fine-textured soils require as much as 18 to 24 hours before the moisture retained is independent of the wetting time. An overnight wetting time of 16 to 18 hours was used for the determinations reported in this paper.

For all except very impermeable soils, 4 to 6 hours was ample time for a layer of soil 1 to 2 centimeters thick to come to equilibrium after the pressure differential was applied to the porous ceramic plates. Most of the soil-moisture-tension data were obtained at a temperature of 21°C.

The relation between moisture equivalent and the moisture retained after the soils were wetted and brought to equilibrium on the suction plate at 4 tension values of 250, 345, 440, and 518 centimeters of water was studied. In general, a fairly close relation between moisture equivalent and the moisture retained at the moisture tension of 345 centimeters of water (one-third atmosphere) was found. However, the percentage at one-third atmosphere was slightly lower than the moisture equivalent for the coarse soils and slightly higher than the moisture equivalent for the fine soils. The authors blamed these inconsistencies on possible sampling errors.

Twelve soils were studied at the ½-atmosphere and 15-atmosphere tensions at 5 different temperatures that were approximately 0, 12, 21, 30, and 37°C. A decrease in moisture retention usually occured with increase in temperature. The change in moisture retention per degree of change in temperature increased from coarse- to fine-textured soils but did not seem to be related linearly to the moisture retention of different soils at any given temperature and tension. The variation in moisture retention for the different soils at one-half atmosphere was from less than 0.5 percent to as much as 3 percent.

The soil-moisture-tension values at any distance from the periphery or moisture-outflow surfaces of the centrifuge sample may be calculated from the equation $T = (\omega^2/2g)(r_1^2 - r_2^2)$ where

T = tension, in centimeters of water;

ω = angular velocity of centrifuge, in radians per second;

g = acceleration of gravity, in centimeters per second per second;

r_1 = distance of bottom of sample from center of rotation, in centimeters;

r_2 = distance of specified point in sample from center of rotation, in centimeters.

The authors stated that the moisture equivalent could not be used generally as an index of either the upper or the lower limit of moisture usable by plants in the field. Although moisture equivalent has the advantage of being a definite reproducible quantity easily determined, its continued use seemed unjustified unless something more closely related to the available moisture range cannot be found. The moisture-retention values in the ¼- to ½-atmosphere ranges are too closely related to moisture equivalent to be of appreciably greater use or significance, except that they are less expensive to determine and represent a more definite physical property of the soil. As a substitute for moisture equivalent, the 1/3-atmosphere percentage seemed to merit some consideration, but the authors felt that the expression "moisture equivalent" should be used only in connection with determinations made with Briggs and McLane equipment.

From tensiometer data available for several soils, field capacity may correspond to a tensiometer reading somewhere near one-tenth atmosphere, but no distinctive feature of the tension-time curve following irrigation occured which could be associated with the condition at field capacity. Because field capacity depends on the nature of the whole profile, including the initial moisture distribution, the moisture-transmitting properties of the soil, the moisture-retaining properties of the soil, and the amount of water applied, it becomes more difficult to estimate field capacity on an isolated soil sample.

An estimate of field capacity of an undisturbed field sample would likely be more reliable than on a soil that is dried or screened but the advantage of using a distrubed sample for routine work is obvious. Centrifuge packing may partly overcome the structural disruption caused by screening, but the ratio of field capacity to moisture equivalent is considerably higher for coarse than for fine soils. The possibility that a moisture-retention value at a lower tension than the 1/3-atmosphere percentage may be a better indication of field capacity was suggested because this tension empties a relatively larger fraction of the pore space for the coarse-textured soils than for the fine-textured soils. The authors found that for the soils studied, about half the water retained at 40 centimeters of water tension by a sample that has been air-dried and screened, closely approximated the field capacity determinations.

Based on experimental results obtained with the pressure-membrane apparatus, the forces contributing to the energy of retention of moisture by soil were divided into two classes, those arising from dissolved materials as expressed in terms of osmotic concentration of an extracted sample of the soil solution, and all other forces. Force action of the second class can be measured by the use of membranes permeable to the soil solution. The physical quality that is determined experimentally by such membranes is the negative pressure to which a solution must be subjected to be at equilibrium through the membrane with the same solution in the soil.

Insofar as the presence of soluble material changes such physical properties of the system as surface tension and density of the soil solution, or hydration and flocculation of the soil colloid, soil-moisture-retention data obtained with tensiometer, suction-plate, pressure membrane, or centrifugation apparatus are independent of and do not involve solution concentration effects.

Roe, H. B., and Park, L. K., 1944, A study of the centrifuge moisture equivalent as an index of the hydraulic permeability of saturated soils: Agr. Eng., v. 25, p. 381-385.

The movement of water through both saturated and unsaturated soils is a prime consideration in both irrigation and drainage. The rate of movement is dependent upon the hydraulic gradient, $\frac{h_f}{L}$ (ratio of effective head to length of soil column), and upon the coefficient of hydraulic permeability, k_s, for, by Darcy's law of fluid flow through porous media, $V = \frac{k_s h_f}{L}$. Because vertical downward flow is involved in both irrigation and drainage, the assumption that the hydraulic gradient, $\frac{h_f}{L}$ equals unity introduces no serious percentage of error into practical problems. The determination of the coefficient of hydraulic permeability, however, presents serious practical difficulties. Numerical values of this coefficient are thus far available for only a very small number of soils, equipment and knowledge of procedures for making such determinations exist at relatively few places, and the determinations often take considerable time. The general problem is further complicated because hydraulic permeability for capillary flow in any unsaturated soil is apt to differ greatly in magnitude from that in the same soil when saturated. Thus, this study related moisture equivalent to hydraulic permeability of saturated soils. The moisture equivalent is an important and well known property by means of which the physical character of a soil may be approximately determined.

Because the centrifuge moisture equivalent, M_{eq}, is generally more readily and quickly determined than the hydraulic permeability, this study was undertaken at the Minnesota Agricultural Experiment Station in the hope of finding the usable empirical relation between corresponding k_s and M_{eq} values. Because moisture equivalent is known for so many soils throughout the country, the determination of a definite relationship between it and the hydraulic permeability of soils, if such a relationship exists, would greatly simplify the application to practice of the law of water flow through soil.

In this study, the tests were made on laboratory prepared samples from Utah, Arizona, and Minnesota rather than on undisturbed field samples.

The moisture-equivalent determination was made in the customary manner, 2 to 5 tests being made, until consistency of results was secured. The average M_{eq} values thus determined were tabulated in this article.

To determine the hydraulic permeability, a variable head permeameter was designed. Very little published data were available regarding permeability tests on disturbed samples in the laboratory. Consequently a considerable number of preliminary test determinations were made in triplicate to find the best means of preparing and packing the samples. One problem was that of choosing a filter to be used at the bottom of the layer of soil. A layer of fine sand 2 centimeters thick gave the least difficulty and most consistent results. Also, the pulverized soil always tended to deposit itself in strata according to size of soil particles when poured into the permeameter and in this process of pouring and tamping, vertical veins or strata of the coarser material were sometimes formed through which the water percolated downward much more rapidly than through other parts of the soil column. The combination of these conditions resulted in serious lack of consistency in the triplicate tests. To eliminate this problem the soil column composed of air-dried soil was thoroughly mixed by pouring three layers of thickness approximately one-third the height of the soil column into the cylinders and stirring with a fork, and then tamping according to standard procedure. This procedure greatly improved the consistency of the results but, even then, satisfactory consistency was not obtained in about 50 percent of the tests.

The average values of hydraulic permeability were plotted as ordinates on double logarithmic paper, against the corresponding M_{eq} values as abscissas. Although a trend was noted towards a type of mathematical relationship, the experimental values clearly showed too wide a deviation from any exact empirical rule to be of value in practical determination of permeability (k_s) values of soils on which no laboratory determinations had been made. Deviations of several hundred percent were noticeable in several instances.

However, these data may be of some value in making rough estimates of permeability coefficients where moisture equivalents are known. Although such estimates may be highly inaccurate, they are better than judging alone, as the data gives some basis for judgment. For possible use for such purposes, these data were replotted on semilogarithmic graph paper, the k_s values on the logarithmic scale as ordinates and the M_{eq} values on the rectilinear scale as abscissas. This was done to make the soil-class zones, as identified by the M_{eq} values, more readily comparable.

Rose, A. C., 1924, Practical field tests for sub-grade soils: Public Roads, v. 5, no. 6, p. 10-15.

The author suggested a practical field test by which moisture equivalent could be estimated. This was done by taking a 500-gram sample of air-dried soil, breaking up the lumps, and then slowly adding water until the soil reached the consistency of putty and could be compacted with a spatula without any free water remaining on the surface. Water was then allowed to drop on the smoothed surface as it was absorbed. Before the moisture equivalent was reached the sample absorbed water readily, but after the critical value was passed, the surface retained a wet shiny appearance. The sample was then oven-dried at 105°C and the percentage of water was calculated as a percentage of the dry weight of the soil. A comparison of moisture equivalent of 29 soil types, determined by the U.S. Bureau of Soils Laboratory, the Oregon Agricultural College Laboratory, and by using this field method, was presented in tabular form. Considering the results of the Bureau of Soils as the standard, the deviation by methods of least squares for the Oregon results and the field method results were ±1.9 and ±1.4 percent, respectively.

Russell, J. C., and Burr, W. W., 1925, Studies on the moisture equivalent of soils: Soil Sci., v. 19, p. 251-266.

The effect of speed of centrifuging, period of centrifuging, and thickness of the soil layer on the moisture equivalent was studied for 16 soils having a wide range of properties. A procedure which was modified slightly from that described by Briggs and McLane was used for determining the moisture equivalent. In this modified procedure the pans were fitted with muslin instead of filter paper and an amount of soil that would give a 10-millimeter layer after centrifuging was used. The soil was saturated from below on a blotting paper and allowed to drain for 18 hours. Results obtained from this procedure were the same as those obtained by the Briggs and McLane procedure. An unsuccessful attempt to prevent circulation of the air during centrifuging and thereby decrease evaporation from the soil was made by packing the centrifuge drum with cotton.

The relationship of the moisture equivalent to the centrifugal force employed was represented by the equation $M = \frac{K}{Gn}$, where M is the moisture equivalent, G is the centrifugal force in terms of gravity, and n and K are characteristics that are constant for each soil. The moisture equivalent decreased about 1 percent when the thickness of the soil was increased from 10 to 12 millimeters. The moisture equivalent decreased 0.6 percent for each 20-minute increase in the period of centrifuging after 20 minutes. The authors stated that the definition of moisture equivalent given by Briggs and McLane was reasonably adequate.

Russell, M. B., and Klute, A., 1954, Movement of water in soils: Agr. Eng., v. 35, p. 808-810.

This paper is a review of 30 reports which pertain primarily to flow of water in saturated and unsaturated soils.

Of interest in specific yield was the authors' short discussion regarding the estimation of the permeability of unsaturated soils:

Deductions based on a statistical treatment of the pore-size distribution curve as proposed by Childs and George have advantage over the direct measurement of unsaturated permeability. The methods for the determination of unsaturated permeability are slow and involve considerable difficulty. The pore-size distribution curve may be obtained from moisture-desorption data, and this curve is obtainable without much difficulty.

Russell, M. B., and Richards, L. A., 1938, The determination of soil-moisture energy relations by centrifugation: Soil Sci. Soc. America Proc., v. 3, p. 65-69.

The capillary potential of a centrifuged sample was determined by the following relationships:

$$\frac{d\psi}{dr} = r\omega^2$$

r = radium of centrifuge

ω = angular velocity

By holding ω constant this equation may be integrated between limits r_1 and r_2 to give:

$$\psi_1 - \psi_2 = \frac{\omega^2}{2} (r_1^2 - r_2^2).$$

If r_1 represents the radial distance to a free-water surface and r_2 the distance to the soil, then ψ, or capillary potential at distance r from the center of rotation, equals zero because it is customary to choose a free-water surface as a reference level and to say that ψ = D at that point.

Therefore the capillary potential of the water in the soil, ψ_2, becomes

$$\psi_2 = \frac{\omega^2}{2} (r_2^2 - r_1^2).$$

Moisture desorption curves between pF (logarithm of capillary potential) 3.0 and 4.0 for four soils were determined by centrifugation. In this procedure an inverted ceramic cup was used. A water level was maintained 5 centimeters below the upper end of the inverted cup. A 5-millimeter layer of soil was placed on the upper surface of the inverted cup. A uniform centrifugation period of 2 hours was accepted as being sufficient to reach an equilibrium moisture content. The moisture equivalent for the soils studied was found to correspond to pF values that averaged 2.6.

Sedgley, R. H., and Millington, R. J., 1957, A rapidly equilibrating soil-moisture tensiometer: Soil Sci., v. 84, no. 3, p. 215-217.

A type of tensiometer using "porvic" (polyvinyl chloride) as the semi-permeable membrane was discussed. The porvic sheets were 0.05 centimeter thick. The pores were approximately 1 micron in diameter and the volume porosity was 85 percent. This type of tensiometer has been used successfully at tensions up to 50 centimeters of mercury. Tests on this tensiometer showed a cup conductance of 11.7 cubic centimeters per square centimeter and a response-time constant of 0.025 minutes.

Sivadjian, J., 1957, Hygrophotographic method for depicting soil moisture: Soil Sci., v. 83, p. 109-112.

This paper described a hygrophotographic method for the qualitative measurement of soil moisture and heterogeneity of soil.

A plate or film of silver bromide was immersed in developer solution in the darkroom, brought into daylight, developed fully, and fixed in a solution of hypo. The plate was then washed thoroughly and bleached in a 3-percent solution of mercuric chloride. After washing again, the plate was transferred to an 0.5-percent solution of potassium iodide. The plate then became uniformly yellow and was rinsed and set aside to dry. The double salt of mercury and silver thus obtained was remarkably stable and insoluble. It was sensitive to light and changed rapidly from yellow to violet black. After exposure it was greatly sensitive to water and atmospheric moisture, which restored it to its original yellow.

These plates, when first exposed to light and then pressed against the soil, gave a qualitative measure of record of a change of moisture conditions within a soil.

Inconstant images on hydrophotographic plates were rendered permanent by reproducing them on bromide paper by the contact method in usual photographic procedure.

Slobod, R. L., Chambers, Adele, and Prehn, W. L., 1951, Use of centrifuge for determining connate water, residual oil, and capillary-pressure curves of small core samples: Am. Inst. Mining Metall. Petroleum Engineers Trans., v. 192, p. 127-134.

The need for accurate values of connate water and residual oil and for capillary-pressure curves in reservoir engineering has been recognized for many years. Several investigators have used centrifugal forces to develop pressure differences between two immiscible phases in a core. The purpose of this report was to establish the practical value of centrifugal methods in determining core properties.

If a core containing water is rotated at an appropriate speed in a tube containing oil or air around the core, the water can be displaced to the connate water value as air or oil replaces the displaced phase. A capillary-pressure curve is then obtained by establishing equilibrium at intermediate speeds of rotation and recording the speed of the centrifuge and the volume of displaced water for each equilibrium state.

A core plug approximately 1 inch in lenth and three-fourths inch in diameter was cleaned and then saturated with a 5-percent soidium-chloride brine. The porosity was determined, using the dry weight and saturated weight, and the final percent of saturation was calculated. The temperature of the centrifuge was maintained at 75°F.

The method employed to establish connate-water saturation depended on the lithology of the core and the choice of the displacing phase. For tight materials, a sealed core holder employing a sand cushion was used and oil was used as the displacing liquid for water. The cores were rotated in pairs at 18,000 revolutions perminute. Although several hours may be required for very tight cores, the time required for equilibrium for most cores was about one hour. Cores of loosely consolidated material required standard centrifuging at normal speeds for 1 hour, usually. Four to 8 cores were run simutaneously.

In capillary-pressure determinations, a special core holder was used. As the water was displaced it was measured in a pipette contained in the lower end of the holder. Readings were taken at different speeds by use of light flashes from a stroboscope while the centrifuge was allowed to operate. Thirty to 40 minutes was usually sufficient time to establish equilibrium at each point. The average saturation of the core was determined by subtracting the volume of displaced water from the original water content.

The percentage of connate water determined by the centrifuge was shown in graph form. In general, a reproducibility within a few tenths of 1 percent of the pore space was found.

Smith, Alexander, 1917, Relation of the mechanical analysis to the moisture equivalent of soils: Soil Sci., v. 4, p. 471-476.

Experimental results relating mechanical analysis to moisture equivalent were given for 12 different soil types ranging in texture from coarse sand to clay. The soils ranged from residual types to recent alluvium. They were separated into seven particle-size groups, and all of the particles that belonged to the same group were combined so that a composite sample for each of the seven groups was obtained. The moisture equivalent for each of the composite groups was determined. The results suggested that if the mechanical analysis was used as an indirect calculation of the moisture equivalent, each textural grade must be given a definite and distinct value and the sands must be considered.

Three synthetic soils were made from the different grades of soil particles. The moisture equivalents of these synthetic soils were determined and found to be very close to the calculated moisture equivalents. If the sands were grouped, the departure of the calculated moisture equivalent from the determined moisture equivalent was much greater than if definite values were given to each of the sand groups. A comparison of calculated and measured moisture-equivalent data for many soils showed variations from -5.44 to +7.28 percent.

The author pointed out that mechanical analysis notes only the amounts of the various sized particles in a given soil, whereas the moisture equivalent is a soil constant which is influenced not only by the size and amount of the different particles present in a soil, but also by the shape of the soil particles, amount of organic matter present, amount of soil colloids present, and chemical composition of the soil. Although moisture equivalents calculated from the mechanical analysis gave approximate results, the author emphasized that these are far from accurate for scientific and for soil-moisture work and that moisture equivalents should be determined for accurate results.

Smith, R. M., and Browning, D. R., 1948, Soil-moisture tension and pore-space relations for several soils in the range of "field capacity": Soil Sci. Soc. America Proc., 1947, v. 12, p. 17-21.

In order to investigate soil-moisture relationships in the field, the following information was obtained for the surface layer of several typical soil profiles: Volume weight and total pore space, maximum water content after thorough artificial soaking, water content after 2 days of free drainage following artificial soaking, water content after thorough soaking by rain, the moisture equivalent, and the tensions and corresponding moisture contents at several times under natural rainfall conditions.

These data indicated that with artificial soaking, the soil-moisture contents were always higher after 2 days of drainage than were the laboratory moisture equivalents. The moisture tensions after 2 days were between 25 and 125 centimeters of water. With less thorough wetting by rain, the tensions developed more rapidly but never approached 500 centimeters of water after 2 days drainage.

Pore-space analysis from the natural-tension curves indicated that the moisture equivalent fits rather satisfactorily near a 500-centimeter tension in the field results. The only way for natural field capacity to be equal to or lower than the moisture equivalent was apparently by a lack of thorough wetting. Medium- or heavy-textured soils that lack complete wetting readily can have field capacity values equal to or lower than the moisture equivalent, but with sandy soils having many middle-sized and coarse pores, the normal field capacity after a few days of drainage was invariably greater than the moisture equivalent.

By considering the degrees of saturation and the true moisture tensions that occur in the range of field capacity, such subjects as soil-water movement, available water supplies, aeration, and temporary water storage can be more clearly understood. The real keys in explaining many of the relationships were the distribution of pore sizes and the degree of filling of the soil pores.

Smith, W. O., 1932, Capillary flow through an ideal uniform soil: Physics, v. 3, p. 139-146.

An ideal soil is an assemblage of spheres packed to a definite porosity. For statistical purposes this soil may be represented by grains placed in hexagonal array, with adjacent grain centers equidistant and at a distance (2r + d), where r is the grain radius and d is a spacing constant adjusted to suit the observed porosity. Three sets of capillaries extending continuously throughout the packing furnish the channels through which fluid crosses the soil body. The velocity through the mean capillary was calculated by a method essentially due to Slichter, and from this quantity, the sectional area of the mean capillary, and the number of capillaries per centimeter squared, the quantity of fluid per second, f, crossing a soil of section area s and length L, was found to be

$$f = 0.00809(1 - P)^{2/3}\left[0.9850/(1 - P)^{2/3} - 1\right]^2 \Delta p D^2 s/\eta L,$$

P = porosity,

D = diameter of grains,

Δp = pressure difference under flow,

η = viscosity of the fluid.

Intrinsic permeabilities computed from this equation were compared with tabular values for carefully sized glass spheres and quartz sands observed by Green and Ampt. A further comparison with corresponding values calculated from Slichter's equation was given.

The problem of capillary flow in soils as compared in the above noted table differed from that of capillary rise reported in the author's earlier paper. The packing employed here was used and experimental agreement was obtained. In the earlier paper, however, the effective area and perimeter used to calculate the height of rise was a weighted mean, in which each of the pore sections occurring in the unit cell was weighted according to its frequency of occurrence. The reason for this was explained as follows: The capillary rise problem involves either filling or draining of pores; for equilibrium it depends only on the perimeter and area of the pore cells in the meniscus. All types of cells will be involved because all share in the filling or draining of the soil. In the flow problem, however, only the capillaries that exist continuously throughout the packing are capable of transmitting fluid after the assemblage is initially filled and steady flow takes place.

Smith, W. O., 1933, Maximum capillary rise in an ideal uniform soil: Physics, v. 4, p. 184-193.

In an assemblage of uniform spheres similar to an ideal soil, the height at which a liquid will be supported above a free water surface has been found experimentally to depend on the way in which this equilibrium was obtained. If the liquid level is falling throughout this media, the height attained is the maximum obtainable; if the liquid is rising by capillarity, the height attained is the minimum. If the liquid level is forced artificially to a given height between these two extremes, the intermediate height will be maintained.

The above generalized picture of maximum capillary rise in an ideal soil is best understood by considering it statistically to be a regular hexagonal arrangement of spheres in which three principal pore cells (triangular, rectangular, and rhombohedral) occur in a definite relative frequency. These cells are so arranged that three sets of capillaries run continuously throughout the media and range from a minimum to a maximum cross section. This cyclic repetition of capillaries of the hexagonal packing thus gives rise to multiple positions of equilibrium; a maximum produced by drainage, a minimum caused by capillarity, and intervening positions if the liquid is forced between these extremes by external means. Pore diameter determines the highest point to which the liquid will be supported at any particular equilibrium because the highest rise occurs when the minimum cross section holds the meniscus and the least rise should occur when the maximum cross-section holds the meniscus. However, the minimum rise is affected by a wetting complication. Rings of liquid form around the point of contact of each pair of soil grains just above the soil meniscus and on coalescence merge and carry the soil meniscus to a height intermediate to the two extremes. The rings are formed as vapor condenses on the grains ahead in the free pore space. This process continues until the ring ceases to grow and an equilibrium is reached.

The conditions controlling capillary rise are determined thus. Using these conditions and an approximation developed for the capillary equation in an earlier paper, the author calculated the minimum rise. On comparison with the data given in the earlier paper, close agreement was evident.

Smith, W. O., Foote, P. D., and Busang, P. F., 1931, Capillary rise in sands of uniform spherical grains: Physics, v. 1, p. 18-26.

Uniform spheres packed in regular array form a noncylindrical cyclic capillary, characterized by a maximum, minimum, or possibly intermediate position of capillary rise. In practice, shperes may be packed to a variety of porosities (P), thus requiring a mixture of regular and irregular pilings arranged in a very distorted pattern. However, the meniscus is also distorted to conform in a general way with the distortions of the lattice. Accordingly, positions of maximum and minimum rise may be expected.

The meniscus for maximum rise tends to pass through the plane of centers of neighboring spheres. Slight deviations from this condition owing to rise at sphere contact are thought by the authors to be of minor importance. Any piling may be treated statistically as a hexagonal array with a spacing $2r + d$ where d is computed to give the observed porosity. In such a system three types of cell occur with a definite frequency, and these cell types are assumed present in the meniscus with the same frequency distribution. Hence it is possible to evaluate $pr/a = \rho ghr/\sigma$ where

p = perimeter,

r = grain radius,

a = area of pore opening,

g = acceleration of gravity,

σ = surface tension,

ρ = density,

h = capillary rise.

The final formula so derived reduces to

$$\frac{pr}{a} = \rho\frac{ghr}{\sigma} = \left[\frac{2}{0.9590/(1 - P)}\right] - 1,$$

where

P = porosity.

This agreed with experiments made with several sizes of grains, porosities, and liquids. The minimum rises were also determined but a satisfactory interpretation in terms of a model has not been effected.

The maximum rise was determined by allowing liquid to fall very slowly in a specially designed vessel containing glass spheres (radius 0.316 centimeter) and Ottawa sand (radii 0.443 and 0.0135 centimeters).

Porosities were measured by observing the volume occupied by a given weight of spheres of known density.

Liquids used were lubricating and crude oil, carbon tetrachloride, acetylene tetrachloride, toluene, benzene, xylene, and water.

In all instances, the minimum rise was determined by allowing the liquid to rise slowly in a previously wetted sand until equilibrium was reached.

Spangler, M. G., and Pien, Wei, 1952, Distribution of capillary moisture at equilibrium in stratified soil: Highway Research Board Spec. Rept. No. 2, p. 120-125.

The theoretical aspects of free energy applied to stratified soil were discussed and the equilibrium moisture contents at various elevations above a free water-table surface in two experimental stratified soil columns were compared with theoretical moisture distributions.

Two columns of stratified-loess and glacial-till soils were placed in glass tubes with a diameter of 4.9 centimeters and a length of 110 centimeters. The lower ends of the tubes were immerssed in a vessel of water which was open to the atmosphere and water was allowed to rise by capillarity. After equilibrium was attained, the moisture contents were measured and compared with sorption curves obtained with a soil tensiometer.

The rate of rise in the soil in the upper part of the tube was greatly influenced by the kind of soil in contact with the water table. The wetting front in the tube which had glacial till in the bottom took much longer to reach the top of the column than it did in the tube with loess at the bottom. Theoretical curves for soil-moisture content versus height above the water surface were constructed from the sorption curves. The theoretical and actual values were in reasonably close agreement throughout the height of the soil columns. The abrupt changes in moisture content at the interfaces between the various soil strata were indicated by theoretical considerations and were fully developed in the experimental soil columns.

Stahl, R. F., Martin, W. A., and Huntington, R. L., 1943, Gravitational drainage from the unconsolidated Wilcox sand: Am. Inst. Mining Metall. Petroleum Engineers Trans., v. 151, p. 138-146.

In some volumetric oil fields the performance of the reservoir during the latter part of its producing life is governed largely by downward movement of liquid due to gravity. The effect of capillarity and of gravitational forces has been related to this phenomena but little data have been reported showing the history of saturation distribution during the drainage period, This was the purpose of the study.

The drainage tests were run in two 8-foot vertical tubes, of 2-inch and 4-inch standard wrought-iron pipe, respectively. Each tube was equipped with a water jacket, which could be kept virtually isothermal (± 1°F). Core nipples for taking samples of sand were inserted at 6-inch intervals. The bottom of the tube was fitted with a short 3/4-inch nipple, into which was inserted a 1/4-inch nipple perforated by many very small holes. To prevent loss of fine sand from the tube, a 100-mesh copper screen was soldered into position covering the perforations. A pressure chamber for filling the tube was equipped with a calibrated glass tube for determination of the amount of liquid that flowed. The tube was packed with a Wilcox sand by tamping and vibrating the tube with blows of a hammer. (This sand was unconsolidated and consisted of well-rounded grains of quartz of average diameter--0.0055 inch.) Porosity was calculated to be 31.9 percent. All subsequent packings provided substantially the same porosity.

The sand-packed tube was filled with liquid, which was forced up through the base of the tube. Liquid was discharged from the top of the tube in an amount equal to or more than the volume required to fill the pore space. The tube was then allowed to drain for several hours and again refilled similarly. After this second filling, which was necessary in order to assure consistent results, the tube was ready for the drainage runs.

At completion of drainage, or at any time when cores were desired, the plugs in the core stations were pulled and samples were taken by means of a cork borer. All saturation data were claculated in percentage by volume of reservoir fluid per unit volume of pore space. Water, heptane, and Wilcox crude oil were used.

A comparison of the equilibrium-drainage curves for the above liquids were shown in graph form.

Historical data on the drainage sequence was obtained by stopping each run when the tube was partly drained and by quickly removing core samples in order from top to bottom before any appreciable changes could take place in the tube. The cavities caused by taking these cores were filled with fresh sand before the tube was refilled with liquid. The runs checked within ±5.0 percent for the final equilibrium saturations. However, for samples taken at partial drainage, the checks were poorer since the cork borer evidently forced some of the crude away from the sampler as it penetrated the highly saturated sand.

The effect of temperature on drainage was determined. Using Wilcox crude oil, the rate of flow was proportional to the temperature in the early stages, but towards the end of the producing life, the rate at 130°F fell below that at 100 or 115°F. The same relationship held true for 3 runs on water at 64, 100, and 130°F.

Staple, W. J., and Lehane, J. J., 1954, Movement of moisture in unsaturated soils: Canadian Agr. Sci. Jour., v. 34, p. 329-341.

Measurements of soil-moisture movement under isothermal conditions from near saturation to an air-dry condition were reported principally for surface samples of the Wood Mountain clay loam, a soil having a field capacity of 22 percent and a wilting point of 9 percent.

Air-dry soil was packed into cylinders that were 9 to 15 inches long. Water was added to the top of the soil column and the soil was sampled every 1 to 4 days for a period of 2 weeks. Different soil densities were obtained in the soil cylinders by changing both the quantity of soil added between tampings and the amount of tamping. The moisture movement at different moisture contents was calculated from the density of the soil and the moisture contents at different depths and time intervals. The moisture movement was converted to capillary conductivity by dividing the amount of moisture moving past each level by the pressure gradient at that depth. Moisture gradients were converted to pressure gradients from capillary-tension data obtained by the pressure-plate and pressure-membrane methods.

Capillary conductivity increased from a minimum near a moisture content of 15 percent to high values at a moisture content above 23 percent. Conductivity increased with soil density, with the greater increases occurring at the higher moisture contents. Conductivity dropped sharply near the field capacity at 22-percent moisture but appreciable movement continued to a moisture content below 15 percent.

Stearns, H. T., Robinson, T. W., and Taylor, G. H., 1930, Geology and water resources of the Mokelumne area, California: U.S. Geol. Survey Water-Supply Paper 619, p. 151-172.

Undisturbed soil columns of water-bearing materials were obtained during the fall when the water table was at its lowest level. A 4-foot-square pit was dug within about 4 feet of the water table. A 16-gage galvanized iron cylinder of 18-inch diameter and 36-inch length was driven into the bottom of the pit. Soil was excavated from around the cylinder as driving progressed. A plate of 12-gage galvanized iron was jacked under the cylinder and then joined to it by soldering. After small-diameter observation wells were installed in the soil columns, and artificial water table was established at a high stage. Measured volumes of water were then withdrawn and added alternately and the water levels were observed.

Two 1½-inch holes were drilled to the bottom and on opposite sides of the soil column. Observation-well casing was installed in each of these holes, one length perforated over nearly the full length, the other perforated only in the lower 6 inches. The perforations were of 1/16-inch diameter and were in parallel rows one-half inch center to center. Water was added to the soil column by pouring it into the casing which had only the lower 6 inches perforated. The water was removed from the cylinder by drawing it up into a glass tube or rubber hose. The amount of water added or removed was measured in a glass graduate. To determine if the water table was level in the cylinder, the water level in both small diameter wells was measured with a steel tape.

The time intervals allowed for the water level to reach equilibrium were too short--only one was over 50 minutes. Later experiments indicated that an interval of several days would provide resultant specific yield that was larger and more nearly the actual specific yield. Greater specific yields were also obtained for a rising water table than for a falling water table. A consistent variation in the water level in response to changes of water temperature in the cylinders was found in all tests. All measurements were corrected to a common temperature.

In order that more of the capillary fringe could be included, in later tests 42-inch soil columns were used. During these tests, measurements of the depth to water were made with an electrical device called a micro-hydrogage. This gage measured depths to water to thousandths of a foot by means of a slow-motion screw and vernier assembly.

Stewart, G. L., and Taylor, S. A., 1956, Field experience with the neutron-scattering method of measuring soil moisture: Soil Sci., v. 83, p. 151-158.

Research has shown that the neutron-scattering phenomenon may be used to determine soil moisture. Hydrogen nuclei influence the scattering of neutrons and because most of the hydrogen nuclei in soils occur in water, the correlation of the scattering of neutrons in the soil with moisture content is possible. Because this measurement is independent of the physical or chemical nature of soil, variables such as temperature, soil texture, and salt concentration can be ignored. This report discussed the results of 2 years of field experience using this method.

The portable neutron-scattering equipment consisted basically of a counting unit and a probe unit. The counting unit was a model 2111 pee-wee proportional alpha counter used in conjunction with either of two BF_3 chambers enriched with B^{10}. One was filled at 35 centimeters of Hg pressure with 11 percent B^{10} and operated from 1750 to 1950 volts. Its sensitive volume was 12.5 centimeters in length and 1.5 centimeters in diameter. A larger (35 centimeter in length and 5.1 centimeters in diameter) and more sensitive chamber was used when this one proved to be somewhat inadequate. Two neutron sources were used; one consisted of an 8 millicurie. Ra D-Be capsules, 9.5 millimeters in length and 7 millimeters in diameter, (with a neutron flux of 60,000 neutrons centimeter^{-2} second^{-1} per capsule); the other weaker source consisted of 1 milligram $RaCO_3$: Be capsule with a neutron flux of 1.5×10^4 neutrons centimeter^{-2} second^{-1} and dimensions of 7.5 millimeters in length and 7.0 millimeters in diameter. The neutron meter was calibrated in the field using material ranging from inorganic sand to clay soils.

Undisturbed moisture samples approximately 14.8 centimeters long were taken in 1-inch auger holes using a 2-centimeter brass tube. A cutting edge of flat spring steel welded inside the tube cut a sample slightly smaller than the tube. The thickness of sample was limited by a spatula inserted in a slit cut into the sampling tube. Samples were taken at 10-centimeter intervals to a depth of approximately 160 centimeters. Unit weight and moisture content were determined gravimetrically.

After the undisturbed moisture samples were collected, the hole was enlarged to accommodate an aluminumn access tube with an outside diameter of 5.35 centimeters and a length of 180 centimeters. Neutron-meter readings were then taken after steady state was reached at approximately 10-centimeter intervals as the probe was pulled up the access tube. Depths were measured from the soil surface to the center of the sensitive zone of the detector tube.

The standard used consisted of a 1,000-milliliter graduated cylinder placed securely in the center of a container 60 centimeters in height and 20 centimeters in diameter. The space between the cylinder and the glass walls was filled with water. The hydrogen in this given quantity of water provided a standard source of counts which could be compared with counts from the soil moisture. Readings were taken before or after each profile or at intervals during the day.

The calibration curve and its equation as determined from linear regression were shown. For the 30- to 125-centimeter depths, $R = 0.36N-0.15$; where R represented the ratio of volume of water in the sample to its bulk volume and was called the water ratio, N was the neutron count ratio of soil to standard. The relative association of count ratio to water ratio was given by the correlation coefficient (r) which for the above calibration curve was 0.95. This curve did not include the surface soil because near the surface (0 to 30 centimeters) there was a strong tendency for the moisture content as measured by the neutron method to be less than true value, apparently because neutrons escape from the soil mass and thus were not detected. If each depth was calibrated separately, the correlation would be improved. The 15-centimeter depth had too low a value, so in using this method, the surface soil should be calibrated separately, whereas all other points usually can be determined by using the calibration curve which included depths 30-125 centimeters. Considering the 30- to 125-centimeter depths, the variation was 9 percent of the predicated regression value. This included errors of sampling when measuring R, weighing and drying errors, and other unknown errors. A large part of error in calibration might be attributed to moisture sampling alone. When the instrument was calibrated carefully, this method was found to be more accurate than resistance methods, and only slightly less accurate than the gravimetric method.

Stolzy, L. H., 1954, The effect of mechanical composition and clay mineral types on the moisture properties of soils: Michigan State Univ. Ph. D. Dissert.

A study was made of the moisture characteristics of 38 Michigan soils. Moisture properties were determined on cores and bag samples taken from each soil layer and field-capacity measurements were made on the different layers after they were artificially saturated and allowed to drain free for 36 to 48 hours.

The soil cores were taken into the laboratory and tension from 0 to 1 atmosphere were determined on the tension table and by the porous-plate method. Tensions from 3 to 27.19 atmospheres were determined on air-dry samples less than 2 millimeters thick by the pressure-membrane apparatus. Moisture equivalents, mechanical analyses, and wilting-point determinations were also made on the soil samples.

The Norelco X-ray spectrometer was used to determine the types and amounts of the clay minerals montmorillonite, illite, or kaolinite occurring in Michigan soils. Illite was the predominant clay and montmorillonite was the least common--less than 20 percent.

The data for the different layers of each soil were tabulated and the moisture-release curves were drawn. The field capacity, moisture equivalent, and the wilting point were indicated on the release curves. The drop in moisture tension from 1 to 3 atmospheres, especially in the A_p horizon (plowed surface horizon), indicated that soil structure was still a factor to be considered in moisture studies above 1 atmosphere.

The relationships of field capacity to moisture equivalent, to 0.06-atmosphere tension, and to 0.33-atmosphere tension, were studied. The relationships of field capacity to moisture equivalent for Michigan soils were similar to those found by other soils investigators in different parts of the country. Samples with field capacity values below 12 percent have a much lower moisture equivalent. Those with moisture equivalent of 12 to 22 percent approach but are still lower than field capacity. Samples with moisture equivalent above 22 percent have lower field capacities. The 0.06-atmosphere tension was the best measure of field capacity on samples below 12-percent moisture while a tension between 0.06 atmosphere and 0.33 atmosphere would be the best measure of field capacity above 12 percent.

The permanent wilting percentages were determined on the stems of tomato plants and then compared with the 5-, 8-, and 15-atmosphere tensions. The permanent wilting percentage approached most nearly the 5-atmosphere tensions with the line of best fit falling between 5- and 8-atmosphere tensions.

The percent of available water in the different zones of surface soils varied from 4- to 16-percent moisture when the clay content of the soil sample was less than 28 percent. This decreased with higher percentages of clay. Subsurface samples with clay content of less than 18 percent had from 4 to 10 percent available moisture whereas available water for subsurface samples with clay content higher than 18 percent decreased with increasing percentage of clay. (Modified from author's abstract.)

Stone, J. F., Kirkham, Don, and Read, A. A., 1955, Soil-moisture determination by a portable, neutron-scattering, moisture meter: Soil Sci. Soc. America Proc., v. 19, no. 4, p. 419-423.

The aim of this paper was to describe the operation of a portable, neutron-scattering, moisture-measuring device which differed from previously reported devices of this type as follows: A fast neutron source in the form of an annulus was placed about the center of a slow neutron-detecting tube, glow transfer tubes were used for absolute neutron-count determinations, and a calibrating volume of paraffin, which was also used as a neutron shield, was incorporated as a part of the source-detector carrying case, to permit simple field checking and standarization of the device.

The positioning of the annular neutron source compared to end positioning increased the number of slowed electrons which return to the detector for a given soil-moisture percentage, and, hence, increased its sensitivity to detect moisture. The volume of soil sampled was also reduced so that changes in moisture for 3- to 6-inch-depth increments could be detected.

A special sampling tube, a tip fastened on the end of a metal tube of the same tube stock, was designed to take 3-inch samples 1.87 inches in diameter to a depth of 30 inches.

The soil-moisture meter was calibrated by using the paraffin of the shield as a standard. The shield was then removed, the probe lowered into the hole, and readings taken at 3-inch increments starting with the neutron source at 4.5 inches depth and continuing to 28.5 inches. The ratio of counts for the soil to standard for a given time interval (30 seconds or 1 minute) was taken as a measure of soil moisture. Three readings were taken at each depth. Gravemetric samples were taken at 4 locations, each within 1 foot of the access pipe. The calibration was made on 3 soils ranging in type from sand to clay and in moisture content from 9 to 50 percent.

To determine the possibility of obtaining more discrimination between varying moisture regions in the soil profile, all but 2 inches of the center part of the detector tube were shielded by cadmium 1/64-inch thick.

Using this shield, an experiment was conducted to see if a wet-dry front could be detected in a 32-gallon can fitted with an access pipe and filled with dry and wet sand. The moisture content was made 45 percent in the bottom and 3.5 percent in the top. For the profile below 6 to 9 inches, the equipment employing the cadmium shield generally gave the soil moisture per unit soil bulk volume within the range of the standard deviation of gravimetric determinations. Graphs of the wetted-front laboratory experiment showed that the front could definitely be picked up by the shielded probe and that closer agreement with true moisture contents of the dry and wet portions of the column was obtained.

Tanner, C. B., Bourget, S. J., and Holmes, W. E., 1954, Moisture-tension plates constructed from Alundum filter discs: Soil Sci. Soc. America Proc., v. 18, no. 2, p. 222-223.

An Alundum porous-plate unit was described that could be used in tension systems up to a moisture tension of 0.15 atmosphere. It was reported to be more economical and convenient to assemble and was more convenient to operate than similar tension apparatus. Advantages over "blotter-type" tension tables were its use at higher tension and its ease of being placed into operation. The primary advantage of this plate over the asbestos tension table, ceramic and fritted-glass plates used in tension systems, and the porous plates used in pressure systems, was the economy of the apparatus and the convenience and simplicity of construction which required no machining or other special fabrication techniques. Basically, the system consisted of an Alundum plate backed with rubber. A valve stem was connected to the rubber and tension was applied from the valve stem.

Terwillinger, P. L., Wilsey, L. E., Hall, H. N. and Morse, R. A., 1951, An experimental and theoretical investigation of gravity drainage performance: Am. Inst. Mining Metall. Petroleum Engineers Trans., v. 192, p. 285-296.

The purpose of the paper was to give experimental data on the performance of a constant-pressure, gravity-drainage system over a wide range in producing rates and to present a method by which accurate prediction of the performance of a gravity-drainage system could be made.

The equipment used was a lucite tube 13 feet long and 2 inches in diameter, vertically mounted and packed by mechanical vibration, with a clean silica sand. To eliminate boundary flow along the wall of the tubing, the lucite was heated and under pressure caused to conform to the outside surface of the sand grains. Saturation distributions of 0.25 normal sodium-chloride brine in the columns were measured at 5-centimeter intervals by an electrical conductivity method. Capillary pressures were determined by allowing the column to drain near the top of the system through a siphon open to the atmosphere. When equilibrium was reached (when no change in saturation could be noted), the conductivity profile was measured and production started at a constant rate through a metering pump connected to the bottom of the system. A constant nitrogen pressure of 50 pounds per square inch was maintained at the top of the column.

The conclusions indicated that recovery to gas breakthrough by gravity drainage was inversely proportional to rate, and that the gravity drainage reference rate (maximum theoretical rate of gravity drainage) was not significant in recovery unless the relative permeability, capillary-pressure characteristics, and displacing fluid viscosities were identical for the systems compared.

Close agreement between experimental and calculated drainage performance proved that steady-state relative permeability and static capillary-pressure data can be used to describe fluid displacement behavior.

The author showed many graphs to substantiate his data; among them, those showing change in saturation with time.

Thomas, M. D., and Harris, Karl, 1926, The moisture equivalent of soils: Soil Sci., v. 21, p. 411-424.

A study of the moisture-equivalent method as it was affected by the amount of material centrifuged, by the texture and chemical treatment of the soil, and by the nature of the outside boundary, led to the following conclusions:

(1) If the sample size of very coarse and very fine soils was increased, the amount of water retained was reduced to a slight extent whereas with soils of intermediate texture greater effects were evident.

(2) The moisture gradient in the soil mass, opposing the centrifugal force of the machine, also indicated a similar maximum with intermediate textures. Very fine-grained soils had a nearly uniform moisture distribution at equilibrium.

(3) A period of many hours centrifuging was often required to establish capillary equilibrium in heavy clays and in very fine silts. The slow movement of water through a silt was shown graphically.

(4) When 10- to 25-gram samples were centrifuged, silt of about 10-micron average diameter retained more water than silt of 5-micron average diameter. This excess of water decreased with the increasing size of the sample.

(5) The very fine silts had a lower apparent specific gravity than the heavy Trenton clay, and also retained as much or more water. Capillaries in the silt which the centrifuge was unable to empty possibly were filled entirely with water, whereas in the clay the corresponding interstices contained some colloidal material.

(6) As the capillaries of the outside boundary were reduced in size, the adjacent soil became drier as reflected in the moisture content throughout the whole soil block. This effect conformed quantitatively to the thermodynamic theory of capillary equilibrium.

(7) When the replaceable base of a clay was entirely sodium, the impermeability, and probably also the equilibrium moisture-retaining power, were greater than when the replaceable base was potassium, ammonium, calcium, aluminum, or hydrogen. The collodial swelling was enhanced by the sodium in the complex.

Thornton, O. F., Marshall, D. L., 1947, Estimating interstitial water by the capillary-pressure method: Am. Inst. Mining Metall. Petroleum Engineers Trans., v.170, p. 69-80.

Several methods are used for estimating the interstitial water saturation at given points within oil and gas reservoirs. This paper discussed the capillary-pressure method and related it to results obtained by other means.

In this method, the core sample was placed in the laboratory under capillary pressure existing at the depth the core was obtained. This apparatus used a metal cylinder with a removable pressure-tight cap through which pressure could be applied. The base of the cylinder was closed with a membrane which was permeable to water but impermeable to air at specified pressure differentials. Cores were cut to convenient size, extracted with carbon tetrachloride and acetone to remove oil, and dried to constant weight. The samples were saturated under vacuum with treated formational water. After the samples were saturated and weighed, they were placed on the membranes in the core holders and a constant air pressure was applied through an opening in the top cover. When equilibrium was indicated by the constant weight of the core samples, the water saturation was computed by difference between the dry weight and the weight of the partially saturated core. The pressure difference between the air and water phase, or capillary pressure, was equal to the gauge pressure of the air.

The author compared interstitial water as obtained by retort distillation at above 400°F, as calculated from the salt content (obtained by titration) of the core of known salinity, and as obtained by the capillary-pressure method. The percentage of interstitial water by distillation was higher than the close agreement obtained by the capillary and salinity methods.

On comparing the interstitial water obtained by the capillary method and that calculated from the electric log, some agreement was achieved provided the log was obtained and interpreted as follows: Use the resistivity recorded with normal electrode arrangement and 16-inch spacing, assume no invasion, assume no bed thickness correction, assume formation factor based on m (slope of line representing the relationship between formation factor and porosity) of 1.5, assume resistivity of water equal to 0.030 ohm-meters, and assume average porosities.

The agreement between capillary pressure and other methods indicated that the capillary-pressure method properly applied yielded results sufficiently accurate for most engineering purposes. However, it was somewhat time consuming and could not be applied where cores were not available.

Van Bavel, C. H. M., Hood, E. E., and Underwood, Newton, 1954, Vertical resolution in neutron method for measuring soil moisture: Am. Geophys. Union Trans., v. 35, p. 595-600.

This paper reported the results of an investigation of vertical resolution, the thickness of the layer of soil that significantly determines the counting rate that is observed in measuring soil moisture.

The equipment consisted of a fast neutron source and a slow neutron-detecting device in close proximity. It used a 10-millicurie Ra-Be source and a BF_5 filled counter (2.5 x 30 centimeters). Rates ranged from 16 to 2 counts per second. The sensitive volume of the counter was partially shielded by pushing a sheath rolled from 1/16-inch metallic column over the tube. Before each logging, a standard run was made in a defined volume of water.

The first series of experiments utilized a 55-gallon barrel filled with sand or soil at uniform moisture content. The moisture content and bulk density were carefully determined throughout the entire barrel. The source-counter assembly was moved inside a thin-walled aluminum pipe, 7.5 centimeters in diameter, placed centrally in the barrel and the counting rate was determined at 5-centimeter intervals over the entire vertical extent of the barrel. A minimum of 6,400 counts were taken, thus giving a standard error of less than 2 percent.

The objective of the first set of experiments was to determine the depth below the surface and the height above the bottom at which the maximum counting rate would be attained, and the effect of moisture content thereon. On plotting the depth of source below the surface against the counting rate, a plateau-like curve was obtained at differing depths and for different moisture contents. By subtracting the length of the plateau in centimeters from the height of the barrel, the vertical resolution was obtained. The vertical extent of the soil layer contributing 95 percent of the measured counting rate ranged from 65 centimeters with very dry material (4.4-percent water by volume) to 45 centimeters with saturated material (38-percent water by volume). Partial shielding of the counting tube with cadmium foil made it possible to increase the vertical resolution at the expense of the counting rate. This meant less accuracy or a long counting time.

In a second experiment, bags filled with wet sawdust (approximately 100-percent moisture by volume) were stacked around the barrel to determine whether or not the barrel was larger in the horizontal direction than the "sample" being measured. The results indicated that the bags of wet sawdust around the barrel apparently did not increase the counting rate appreciably when the moisture content was 28.4 percent by volume and increased the counting rate only 6 percent at a moisture content of 9 percent by volume. However, for moisture contents below 5 percent, the barrel was not "infinite" in the horizontal direction and the counting rate was, therefore, too low.

The conclusions regarding the use of this equipment were as follows:

(1) The method was simple in practical application;

(2) The equipment performed dependably and was not excessively expensive;

(3) An accuracy of about 3 percent in moisture content by volume required about 3 minutes of counting;

(4) The verical resolution appeared to be smaller than would be desired for most work;

(5) Partial shielding of the counter was shown to be rather ineffective.

Veihmeyer, F. J., 1927, Some factors affecting the irrigation requirements of deciduous orchards: Hilgardia, v. 2, no. 6.

Early in 1919 a system of soil sampling was begun in the Santa Clara Valley, Calif. to study the moisture behavior of soils in response to different irrigation practices in the prune orchards. Moisture-equivalent determinations were made on many samples and the plan was to record percentages of moisture as ratios of the moisture equivalent, or of any of its related soil-moisture constants. However, wide variations were found in the moisture equivalents after repeated determinations were made on the same sample of soil. In spite of the variations in moisture equivalent and consequently in the wilting coefficients that were calculated from them, reasonably close agreement was obtained between the wilting coefficient and the field moisture content determined at wilting point for the trees. The author determined also that the movement of moisture from moist soils in contact with dry soils in columns over a 4½-month period was small. The results of these studies indicated that when the soil was not in contact with a free water surface the capillary movement of moisture from a moist soil to a drier soil was probably too limited in extent and in rate to be important for plant use.

Veihmeyer, F. J., and Hendrickson, A. H., 1931, The moisture equivalent as a measure of the field capacity of soils: Soil Sci., v. 32, no. 3, p. 181-193.

Field capacity, as determined in this study, indicated the amount of water held in the soil after the excess gravitational water had drained away and after the rate of downward movement had materially decreased. The moisture equivalent was found to be a close measure of the field capacity of deep, well-drained soils that had no decided changes in texture or structure with depth and that had moisture equivalents ranging from about 12 to 30 percent. Below 12 percent, however, the moisture equivalent was less than field capacity. Field-capacity values were not determined for soils that had moisture equivalents greater than 35 percent.

The moisture equivalent of the medium-textured soils gave a reasonable value closely related to texture even though the structure of the sample used in making the moisture-equivalent determination was changed from that found in the field.

The authors stated that a rather definite soil-moisture content existed which, if measured within 2 to 3 days after a rain or irrigation, could be assumed to be field capacity providing no discontinuity of structure or texture occurred in the soil strata and providing the water table was absent.

Veihmeyer, F. J., Israelsen, O. W., and Conrad, J. P., 1924, The moisture equivalent as influenced by the amount of soil used in its determination: California Univ. Agr. Expt. Sta. Tech. Paper 16, 61 p.

Moisture equivalents determined by the Briggs and McLane procedure were obtained for samples of six soils. As the size of the sample was increased, the percentage of water retained decreased, the difference being greater with small samples than with large ones. Clay soils became impermeable to water during centrifuging when the amount of soil placed in the centrifuge cup was increased. Some clay soils would re-absorb the supernatant water when the centrifugal force was relieved.

Thirty-gram samples of Yolo clay loam were used to test the effect of the duration of centrifuging on the percentage of moisture retained. In general, a small decrease in moisture content occurred with increased time of centrifuging.

The effect of the length of moistening periods of 2, 6, and 24 hours upon the percentage of moisture retained by various weights of samples was determined. The percentage of moisture retained increased with time of wetting, and was signigicantly higher after the 24-hour wetting. For all periods of wetting, the percentage of retained moisture decreased when the amount of soil used in the determination was increased.

Changing the weight of soil per unit of cross-section had a direct relation to the percentage of moisture retained and further suggested that no variation occurred in the percentage of moisture retained normal to the radii of the centrifuge.

To determine the amounts of moisture held in thin layers of samples, the centers of which were at different distances from the axis of rotation, samples were sliced approximately normal to the radius and the slices dried separately. These slices were at least 2 millimeters thick and were made by a special microtine. Results indicated that within a 60-gram block of soil, the moisture content increased from the inner to the outer surface, but that the rate of increase was less than that resulting from reducing the size of sample from 60 to 5 grams.

Separate determinations were made of the apparent specific gravity of the volume of different weights of soil after centrifuging. The paraffine-immersion method and the direct micrometer method were used in making these volume determinations. The apparent specific gravity was found to increase as the weight of the sample was increased. For successive thin layers of soil in a 60-gram sample, the apparent specific gravity increased significantly from the inner surface outward to a point about 6 millimeters within the sample, beyond which it remained almost constant to the outer surface.

Pressures of different magnitude were applied artificially to different amounts of soil. The results indicated that an increase in the magnitude of the compressing force decreased the percentage of moisture retained against the centrifugal force.

The percentage of the pore space occupied by water in centrifuged samples of different soils and in different weights and successive layers of soils indicated that the amount of water retained against the centrifugal force was not limited by the pore space. The outer layer of a centrifuged block of fine-textured soil might be saturated or nearly so, but the limit of the amount of water retained by the soil against the centrifugal force was not determined by the pore space of the entire block of soil. If precision was desired, the measurement of the sample by weight rather than by measuring cup was necessary. Subjecting the moist soil to centrifugal force until the moisture was in equilibrium with this force resulted in the establishment of an equipotential region throughout the block.

Veihmeyer, F. J., Oserkowsky, J., and Tester, K. B., 1927, Some factors affecting the moisture equivalent of soils: Internat. Cong. Soil Sci. Proc., v. 1, p. 434-442.

Repeated moisture-equivalent determinations of thoroughly mixed samples of the same soil did not give consistent results. As a consequence, a detailed study of the moisture-equivalent procedure was made selecting soils that had a wide range of moisture equivalents. A quantity of air-dry soil, equivalent to 30 grams of oven-dry soil, was used. The soils were completely submerged in water, were allowed to drain for 30 minutes, and then were centrifuged at 1,000 times the force of gravity for a period of 30 minutes.

Moisture-equivalent tests made on samples stored in a moist condition were in some instances different from moisture equivalent tests made on air-dried samples. Air-drying the moist samples apparently restored the soil to a condition that gave about the same moisture equivalent as the original air-dried samples.

A study of the comparison of oven-dry and air-dry samples used for moisture-equivalent determinations indicated that any changes affecting the moisture equivalent that take place in oven-drying will be restored on air-drying the samples. However, the authors emphasized that this restoration could be a slow process.

Machine grinding, pulverizing the samples with mortar and pestle, pulverizing the sample by rolling, and forcing the soil through a sieve with the fingers gave different moisture equivalents. Variable results were obtained between operators when the samples were pulverized with mortar and pestle and pulverized by rolling. Consistent results were obtained between operators when the soil was forced through the sieve with the pressure of the fingers. Standardization of the machine-grinding method of preparing the samples seemed impractical.

The results from a study on the effects of puddling of a soil sample on the moisture equivalent indicated that no unusual precaution needed to be taken to guard against jarring the samples when they were saturated.

The difference in the moisture equivalents obtained by wetting the samples from below and by completely submerging the samples in water was not significant. A study of the effect of the length of time that the soil was kept moistened before centrifuging showed that in general the moisture equivalents increased with increased duration of the saturation period but that the differences between the moisture equivalents of samples saturated for 9 to 12 hours and 24 hours and for 24 hours and 46 to 54 hours were very small.

The differences in moisture equivalents obtained when the temperature of the air in the room varied were small. However, a small decrease in the moisture equivalent was observed when the temperature at which the soil was stored was increased.

Observing all the precautions found to be necessary as a result of this investigation, moisture-equivalent determinations were made on 28 soils. The data indicated that the moisture-equivalent determinations, when properly made, could be used with confidence.

The authors stated that the points to be emphasized in the moisture-equivalent procedure were as follows:

(1) Moisture equivalents should be made on air-dried samples only.

(2) Soil should be crumbled and forced through a 2-millimeter sieve with fingers only. Large lumps could be broken by mechanical means if care were taken not to pulverize the soil.

(3) No special care needed to be taken to smooth the surface of the soil in the cups. Tapping the cups lightly to level the surface of the soil gave satisfactory results.

(4) For usual purposes, samples could either be wet from below or completely submerged.

(5) Samples should be saturated for about 20 to 28 hours.

(6) Temperature during saturation, as well as during centrifuging, should be about 20°C.

(7) No special care needed to be taken to balance the load in the centrifuge if the proper speed was maintained.

Vomcil, J. A., 1954, In situ measurements of soil bulk density: Agr. Eng., v. 35, p. 651-654.

This paper presented a method of determination of bulk density that did not require the removal of a large sample but one that took advantage of the properties of a radioactive material for measuring purposes.

This method employed a two-probe system instead of the usual neutron meter employing a one-probe system. By this system the density of a layer 2 or 3 inches thick and within 3 inches of the surface could be measured.

The probes were two aluminum tubes, with 1 inch outside diameter and a wall thickness of 0.064 inch, fastened together near one end by two lengths of 2- by 3-inch oak wood so the distance between them was 12 ±1/64 inch. The cobalt 60, consisting of an 8-millimeter section of 1-millimeter wire contained in a hollow aluminum wafer, was housed in one probe. A second wafer, riveted to the first by an off-center rivet, served as a lid. Lead plugs 2 inches long were fitted above and below this source. The other probe contained a Geiger-Mueller detector tube in a side-window shield with a 10-foot coaxial cable.

Calibration was necessary to relate the system to soils of unknown density before it was used. The calibration was made in a steel cylindrical tank 16 inches in diameter and 30 inches in depth. Two aluminum tubes with inside diameter of 1.1 inch and wall thickness of 0.035 inch were fixed in the tank so that the probes could be lowered in them. The tank was filled with soil whose bulk density was determined by its mass and volume. Transmittancies of a number of soils at different bulk densities and moisture contents were determined in the calibration chamber. The gamma-ray intensity transmitted was plotted against wet bulk density. Eight readings at 15-second intervals were made giving a probable error of ±1.5 percent.

For field use two properly sized holes spaced 12 inches apart and parallel to one another were bored to 4 inches below the desired depth of measurement. The soil from these holes was saved for moisture determination. Using these data the wet densities could be corrected to dry-bulk densities. The measurement of a series of counts was then made at the desired depth and the average of these compared on the calibration graph for conversion to wet-bulk density. For purposes of spacing and alining the prepared holes, a guide constructed of two stainless-steel tubes was used. The maximum depth of measurement depended on the length of the probe.

To investigate the border effects, a wooden box 2 by 2 by 7 feet was used and steel partitions inserted perpendicularly to the length in two positions. Then several different soil and metal arrangements were tested. The differences in counting rates due to differing border conditions all were found to be less than the probable error of measurements. Comparing these with bulk density by the core-sampler method, differences of only .02 to .03 grams per centimeter were found between the methods.

To correct for decay changes in the cobalt 60 and for other changes in the system, monthly calibration was necessary. The lead shield and a container of water (same container and depth each time) were found to provide continuing calibration for the system.

A lead shield 1½ inches in thickness and long enough to cover the source was used to cover the probe when it was out of the hole.

Wenzel, L. K., 1933, Specific yield determined from a Thiem pumping test: Am. Geophys. Union Trans. v. 14, p. 475-477.

In 1931, a pumping test was made in the Platte River valley, in Nebraska, to determine by Thiem's method the permeability of the water-bearing sand and gravel that underlie the area. The well was pumped continuously for 48 hours and measurements of the discharge were made every 50 minutes. Measurements of water-level changes were made in 80 observation wells located within the area surrounding the well (maximum radius 1,200 feet).

This paper presented a method for determining specific yield based on the Thiem equation.

Using the measurements of depths to water in observation well, draw-down curves were plotted for each well and profiles along a section parallel to the direction of natural ground-water movement were constructed of the cone of depression after 2, 6, 12, 24, 36, and 48 hours of pumping. The total volume of sediments unwatered in each of the successive periods and the volume unwatered in each ring were computed. The quantites of water that percolated towards the well through each of the successive concentric cylindrical sections were computed from the formula Q=PIAT where

Q = quantity of water in gallons a day,
P = permeability,
I = the slope of cone of depression,
A = the area of the cylinder in square feet,
T = the period of pumping in days.

A cylindrical section of water-bearing material with a radius of 50 feet was used as the control section. Each value of specific yield was obtained by computing the difference between the quantity of water that percolated through the 50-foot cylinder and the quantity that percolated through a larger cylinder, and dividing this difference in quantity of water by the volume of material unwatered between the 50-foot cylinder and the larger cylinder.

The values of specific yield computed for different rings about the pumped well and for several periods were given in tabular form. The average value of specific yield determined for the last 12 hours of pumping was 24.8 percent. A sample of the material that was unwatered during the test was examined in the hydrologic laboratory and the specific yield determined from values of moisture equivalent and porosity. Applying moisture equivalent directly as specific retention, the specific yield of the sample was computed to be 24.5 percent, which compared favorably with the average value determined from this pumping test.

Wenzel, L. K., 1942, Methods for determining permeability of water-bearing materials with special reference to discharging well methods: U.S. Geol. Survey Water-Supply Paper 887, 192 p.

In connection with an investigation of the ground-water resources of Nebraska and Kansas, four pumping tests were made to determine the permeability of water-bearing alluvial materials along the Platte and North Platte River valleys in Nebraska. This report primarily described the investigation and outlined the general methods of determining permeability in the field and in the laboratory. However, specific yield (or storage coefficient under water-table conditions) was discussed in relation to the pumping tests.

The value for the specific yield of a material, as ordinarily determined in the laboratory, is probably not attained in the field except when the water-bearing material is permanently unwatered. The slow drainage of water-bearing material in the vicinity of a pumped well causes the water table to decline rapidly at first and then more slowly as drainage proceeds. Many investigators have found that although saturated materials may yield a very large percentage of water in a few hours that they may continue to yield small amounts of water for several years. This was substantiated by specific-yield data computed from pumping-test records collected near Grand Island, Gothenberg, and Scottsbluff, Nebr.

For the test near Grand Island, specific yield was calculated by the non-equilibrium formula to be 21.7 percent after 48 hours of pumping. By the Thiem method, the calculated value of specific yield was 9.2 percent after 6 hours of pumping, 18.5 percent after 36 hours pumping, and 20.1 percent after 48 hours of pumping. From this, a much longer period of pumping would be required before true specific yield would have been reached and the value was estimated to lie between 22 and 23 percent.

At Gothenberg and Scottsbluff, the specific yield determined by the non-equilibrium formula was 3.24 percent and 1.8 percent, respectively.

At Gothenberg, computed values for both permeability and specific yield increased during the period of pumping owing to the slow drainage of unwatered material. Specific yield increased about 85 percent over an 11-hour pumping period.

White, W. N., 1932, A method of estimating ground-water supplies based on discharge by plants and evaporation from soil: U.S. Geol. Survey Water-Supply Paper 659A., 106 p.

Methods of measuring discharge by transpiration were investigated by experiments based on the daily fluctuations of the water table in the Escalanta Valley in southwestern Utah. The amount of ground water discharged daily by plants was computed by the formula $q = y\ (24r \pm s)$ in which q is the depth of ground water withdrawn in inches, y is the specific yield of the soil in which the daily fluctuations of the water table take place, r is the hourly rate of rise of the water table in inches from midnight to 4 a.m., and s is the net fall or rise of water during the 24-hour period in inches. Specific yield was the only item that could not be determined readily from automatic records of the water-table fluctuations. To determine specific yield, a pit 4 by 6 feet was dug to a level above the water table equal to the height of several cylinders. The cylinders were of 16-gauge galvanized steel, 12 or 18 inches in diameter, and 18, 36, or 54 inches in length. The lower rim of the cylinders was slightly leveled in order to provide a cutting edge. As the cylinders were driven downward from the bottom of the pit, the pit was deepened keeping the bottom about 6 inches above the cutting edge of the cylinders. This reduced the disturbance of the soil column and made driving easier owing to the reduction of friction on the outside of the cylinder. Driving was stopped within a few inches of the water table, the soil around the cylinder was removed, and a metal plate was shoved under the cylinder and soldered to it, thus confining the column of undisturbed material in a watertight vessel. After the bottom was attached, small wells were sunk in the material enclosed within the cylinder and the effects on the water level in these wells produced by adding or withdrawing measured quantites of water were observed. Equilibrium was not established until about 24 hours after the addition of water and about 48 hours after its removal. Pronounced changes in barometric pressure between the time of addition or withdrawal and subsequent water-level observations seriously interfered with the results, requiring that observations only be made in fair weather. Occasional corrections were necessary to compensate for temperature effects on the water table. Data were presented in tabular form indicating for clay and clay loams a range in average specific yield from approximately 1 to 7 percent.

Wilcox, J. C., and Spilsbury, R. H., 1941, Soil Moisture Studies II: Some relationships between moisture measurements and mechanical analysis: Sci. Agr., v. 21, p. 459-478.

Studies of the moisture-holding capacity and other soil-moisture measurements, and of the relationships between these measurements and the textural properties of soil were reported.

Determinations of the moisture-holding capacity under field conditions, the mechanical analysis, the wilting coefficient, and the settling volume were made on 100 soils. The soils ranged from coarse sand to heavy clay. Variations in the moisture values were closely related to differences in both the mechanical analysis and the settling volume. The results indicated close correlations between the field capacity and the percentage of sand, and between the wilting coefficient and the percentage of colloid. The mechanical analysis and the settling volume were satisfactory for use in determining the field capacity and wilting coefficient. The authors' equations for estimating field capacity based on the mechanical analysis are listed below:

Field capacity in percent = (1) $- 0.276\%$ sand $+ 34.28$,

= (2) $- 0.301\%$ sand $- 0.0452\%$ clay $+ 36.435$.

These equations held promise in the area from which the soil samples were taken. However, they might not apply to other areas where climatic conditions are somewhat different. The first of the equations was the most accurate.

Work, R. A., and Lewis, M. R., 1934, Moisture equivalent, field capacity, and permanent wilting percentage and their ratios in heavy soils: Agr. Eng., v. 15, p. 355-362.

This paper reported the results of studies of soil-moisture constants and their ratios as determined for slightly pervious clay soils at Medford, Oreg. Soil samples for moisture determinations were generally taken in 1-foot increments and one moisture-equivalent determination was made for each soil sample. With certain samples, some indication of puddling in the centrifuge was evident and the moisture equivalent values for these samples seemed high. Accordingly, sand mixes were made of these samples and moisture-equivalent determinations were made on the sand-soil mixtures. This method involved determining and plotting the moisture equivalent of four sand-soil mixes. The 4 points were plotted and the line through them extended to 100 percent to give an estimate of the moisture equivalent of the soils.

In general, the maximum available moisture content decreased with depth. However, the moisture equivalents for the Medford soils increased with depth to 3 feet and were not equal to field capacity. The field capacity was 0.83 of the moisture equivalent in the 0 to 3 feet average of two heavy soils. The field capacities on Meyer clay adobe soils usually decreased with depth.

Wyllie, M. R. J., and Gregory, A. R., 1955, Fluid flow through unconsolidated porous aggregates: Indus. and Eng. Chemistry, v. 47, no. 7, p. 1379-1388.

The Kozeny equation as modified by Carman relates the permeability of a porous medium to its specific surface area and porosity. As derived, it applies only under viscous flow conditions and in unconsolidated porous media. This study supported the previous views that the Kozeny-Carman constant is a parameter dependent upon the porosity and particle shape of porous aggregates and did not contradict the limited published data which advocate its successful application to fluid flow through consolidated porous media.

The media used in this study were composed of spheres, cubes, cylinders, triangular prisms, and disks. These were packed in a lucite test section, 5 inches in diameter, composed of five 1½-inch interchangeable segments held together by bolts and sealed with O-rings. Pressure connections to each section were made through piezometer grooves cut into the face of each section. Wire-gauze supporting screens in turn supported by ½-inch metal grids were fitted across the top and bottom of the test section. Resistance measurements were made. Fluids (oil, water, or aqueous solutions of glycerol) were passed through the test section from bottom to top. To eliminate the difficulty of producing homogeneous isotropic packing of a mixture of spheres, a method was finally devised in which the spheres, after being mixed thoroughly in a funnel-shaped reservoir, were permitted to fall into a cylindrical mixing chamber. In the mixing chamber was a lattice of 1-inch balls suspended on thin steel rods. As the mixture of spheres fell through the mixing chamber, a random packing of aggregates was obtained.

AUTHOR INDEX

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SUBJECT INDEX

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