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Concepts of Groundwater Occurrence and Flow Near Oak Ridge National Laboratory, Tennessee

Gerald K. Moore

Environmental Sciences Division
Publication No. 3218

ENVIRONMENTAL SCIENCES DIVISION · OAK RIDGE NATIONAL LABORATORY

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CONCEPTS OF GROUNDWATER OCCURRENCE AND FLOW NEAR OAK RIDGE NATIONAL LABORATORY, TENNESSEE

Gerald K. Moore*

Environmental Sciences Division
Publication No. 3218

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NUCLEAR AND CHEMICAL WASTE PROGRAMS
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ABSTRACT

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Previous studies of the area near Oak Ridge National Laboratory (ORNL) assumed that nearly all groundwater from precipitation and infiltration moves vertically down to the water table and then follows a combination of intergranular and fracture flow paths to the streams. These studies also generally assumed nearly linear flow paths, amounts of groundwater flow that are determined by differences in water-level elevation, large permeability differences between regolith and bedrock, and important hydrologic differences between named geologic units. It has been commonly stated for 37 years, for example, that the Conasauga Group has fewer cavities and is less permeable than the Chickamauga Group. All of these assumptions and conclusions are faulty. The new concepts in this report may be controversial, but they explain the available data.

Only the stormflow zone from land surface to a depth of 1-2 m has a permeability large enough to transport most groundwater to the streams. Calculations show that 90-95% of all groundwater flow is in the stormflow zone, 4-9% is in a few water-producing intervals below the water table, and about 1% occurs in other intervals. The available data also show that nearly all groundwater flows through enlarged openings such as macropores, fractures, and cavities, and that there are no significant differences between regolith and bedrock or between the Conasauga Group and the Chickamauga group.

Flow paths apparently are much more complex than was previously assumed. Multiple paths connect any two points below the water table, and each flow path is more likely to be tortuous than linear. Hydraulic gradients are affected by this complexity and by changes in hydraulic potential on steep hillsides. Below the water table, a large

difference in the head of two points generally does not indicate a large flow rate between these points. Groundwater storage in amounts above field capacity is apparently intergranular in only the stormflow and vadose zones. At deeper levels all effective porosity is in fractures.

The subsurface hydrology of the ORNL area is also more favorable for the containment of radioactive wastes than has been indicated by previous reports. A relatively simple solution should be possible for the problem of radionuclide migration in groundwater. The key concepts for most remedial investigations may be hydrologic isolation of contaminated materials by stopping infiltration and lateral flows in the stormflow zone.

1. EXECUTIVE SUMMARY

The groundwater flow system near Oak Ridge National Laboratory (ORNL) is complex in detail but basically consists of only three zones from land surface to the base of fresh water. The characteristics of these zones mean that contaminant transport in groundwater is a relatively simple problem for any required remedial action. The first zone is just beneath land surface and is the root zone for vegetation; it is 1-2 m thick. This layer is above the water table, and groundwater is transitory, lasting a few days to a few weeks after precipitation ends. Nevertheless, 90-95% of all groundwater flow is through this zone to discharge points at nearby springs and streams. Average hydraulic conductivity is about 8.8 m/d (220 $gal \bullet ft^{-2} \bullet d^{-1}$) in forested areas and 2.0 m/d (50 $gal \bullet ft^{-2} \bullet d^{-1}$) in grass and brush areas. The water from this zone is acidic to neutral and generally has a total dissolved solids content of less than 100 mg/L.

The middle zone extends to a depth of 20-60 m and includes intervals above and below the water table. The geometric mean of hydraulic conductivity in the vadose part of the middle zone is 0.0030 m/d (0.074 $gal \bullet ft^{-2} \bullet d^{-1}$), less than 0.002 times as large as those in the upper zone. This contrast in hydraulic conductivity is the factor that causes lateral groundwater flows in the upper zone. The water table generally is near the contact between regolith and bedrock at a geometric mean depth of 4.1 m. Recharge occurs slowly by downward percolation of water during times when there is groundwater in the upper zone. Below the water table, the middle zone generally consists of a few intervals with water-producing fractures in otherwise relatively impermeable material.

About 90% of the groundwater that reaches the water table flows through the middle zone to discharge points in seeps, springs, and streams. The groundwater occurs in complex networks of intersecting fractures. Distinctive characteristics of flow in fracture networks are the occurrence of numerous splits and joins along the flow paths and the movement of water in two directions, generally near vertical and near horizontal, along a single fracture. The first water-producing interval in the middle zone is near the water table. Other water-producing intervals occur at an average vertical spacing of about 10 m. The geometric means of hydraulic conductivity for water-producing intervals and for the intervening, relatively impermeable intervals are 0.041 and 0.00044 m/d (1.0 and 0.011 $gal \bullet ft^{-2} \bullet d^{-1}$). Water in the fractures is a slightly to moderately alkaline calcium bicarbonate type with a total dissolved solids content that generally is less than 500 mg/L.

The remaining groundwater seeps through networks of tightly compressed fractures in the deep zone. This zone extends to the base of fresh water at a depth of 150 m or more in the ORNL area. Water in the deep zone comes from shallower levels and eventually flows upward, back to shallower levels. The geometric mean of hydraulic conductivity in the deep zone is 0.00044 m/d (0.011 $gal \bullet ft^{-2} \bullet d^{-1}$), the same as in relatively impermeable intervals of the middle zone. At least some of the water is an alkaline sodium carbonate type with a dissolved solids content of about 500 mg/L.

Various approaches to remedial action are possible in the burial grounds and in other waste management areas subject to contaminant mobility in groundwater. Excavation and reburial have been mentioned as solutions, but they might only move the same problem elsewhere. A process that would make all contaminated materials less permeable than their surroundings is possible; recent studies have shown more or less successful results with in situ compaction, grouting, and vitrification. However, these processes are slow and labor intensive. Also it is difficult to determine that the objective has been completely achieved and is permanent; other remedial activities might be required at a later date. An alternative approach is the hydrologic isolation and draining of contaminated material by stopping infiltration and lateral flows in the shallow zone above the water table.

One hydrologic solution to contaminant isolation consists of (1) blocking infiltration by precipitation over the entire control area, (2) blocking lateral groundwater flow near land surface at the perimeter of the control area, (3) dewatering the interval between land surface and a depth of about 7.6 m, and (4) treating the water extracted by dewatering. These principles are not new, and various partial applications have been only partly successful in Waste Area Groupings 4 and 6. However, the principles do not represent alternative solutions. All principles must be used over the entire control area in order to be effective, but any selected procedures can be completed relatively quickly and should provide a permanent solution to the problem.

Infiltration of precipitation can be stopped by pavement or by any of various surface treatments or liner materials. The engineering objective is an average surface permeability that is less than that of unsaturated material in the middle zone. A French drain from the surface to a depth of 2 m is one method of perimeter water control. The surface seal should extend across and at least several meters beyond the perimeter drains. Interior dewatering may not be essential in many areas but is recommended (1) as a backup in case of water leakage from the surface seal or the perimeter drains and (2) as a method of quickly stopping contaminant transport out of the control area. Also, the perimeter drains otherwise may occasionally produce small amounts of contaminated water. Arrays of well points are commonly used for a drainage problem of this type. A single pump with a capacity of 38 L/min (10 gal/min) at 6 m of vacuum can extract water from an average of 35-40 well points in the ORNL area.

2. INTRODUCTION

Radionuclides and other wastes have been stored in shallow burial grounds near ORNL since 1943. The main problem resulting from the waste storage practices and from accidental spills has been the entrainment and mobility of contaminants in groundwater. This mobility is determined by the hydrologic and geochemical properties of near-surface materials. Remedial investigations, water quality monitoring, and other types of groundwater research in this area require a conceptual understanding of groundwater occurrence, recharge and discharge, flow paths, and natural water chemistry.

A number of previous investigations have been made of groundwater conditions in the area, mainly near burial grounds, and a considerable body of knowledge has been developed for local conditions and problems, such as water-level elevations and fluctuations, along-strike fracture flows, sorption of radionuclides, and bathtubbing. Also, most early interpretations of aquifer characteristics have proven valid. Unfortunately, some few data were incompletely or incorrectly interpreted, and these errors affect current plans for remedial investigation and remedial action. The errors are not necessarily the fault of previous workers. Groundwater conditions are not directly observable, fracture flow paths are extremely complex, some important data were not available for study, and some characteristics of fractured-rock aquifers are still unknown or controversial. Nevertheless, a revised conceptual model of groundwater occurrence and flow is needed for effective monitoring of contaminants, as is required by federal and state regulators.

A revision of groundwater concepts in the ORNL area is an important element of the Environmental Restoration and Facilities Upgrade Program, begun in FY 1986. This program initiated the first comprehensive study of groundwater characterization for the entire ORNL complex. The program plan requires acquisition of basic geologic and hydrologic data followed by a determination of geochemical processes and by identification and modeling of pathways and contaminant migration. The strategy includes determining the limits of the uppermost aquifer, as specified by U.S. Environmental Protection Agency (U.S. EPA 1986) under the Resource Conservation and Recovery Act (RCRA).

Basic geology and groundwater data have been acquired by drilling, testing, and monitoring of new piezometer, hydrostatic head, and water quality monitoring wells. Nearly all of the new wells have been installed in and near Waste Area Groupings (WAGs), which are areas that include one or more waste management units. Most WAGs are in the White Oak Creek drainage basin (16.9 km^2; Fig. 1). Some wells are near WAGs 3, 11, 12, 13, and 18, which are partly or entirely in nearby basins. About 40 other wells are on western Chestnut Ridge, northwest of the White Oak basin. All data are within a 4.5 km radius of the ORNL main plant (WAG 1), and this radius constitutes the study area.

2.1 PURPOSE AND SCOPE

This report is an interpretation of available groundwater data in the study area. As such, it is a theory that describes groundwater recharge, occurrence, movement, and discharge. Any theory is tested by the acquisition of future data but is a working conceptual model for numerical modeling and research. The concepts in this report should lead to an improved understanding of the relationships between groundwater flow systems and contaminant migration. The report is not intended to be a encyclopedia of groundwater information in the ORNL area. Thus, local flow systems and local problems in the WAGs are not discussed. However, the concepts in this report should aid effective planning for remedial action in the WAGs.

2.2 STATISTICAL DATA ANALYSIS

Groundwater parameters in the study area have a large numeric range, little correlation with other parameters, and abrupt spatial changes in value. Tools such as contour maps and regression analyses have limited utility in this area and can produce misleading results. The problem apparently results from nearly unique conditions along each of the many groundwater flow paths. In this situation, statistical

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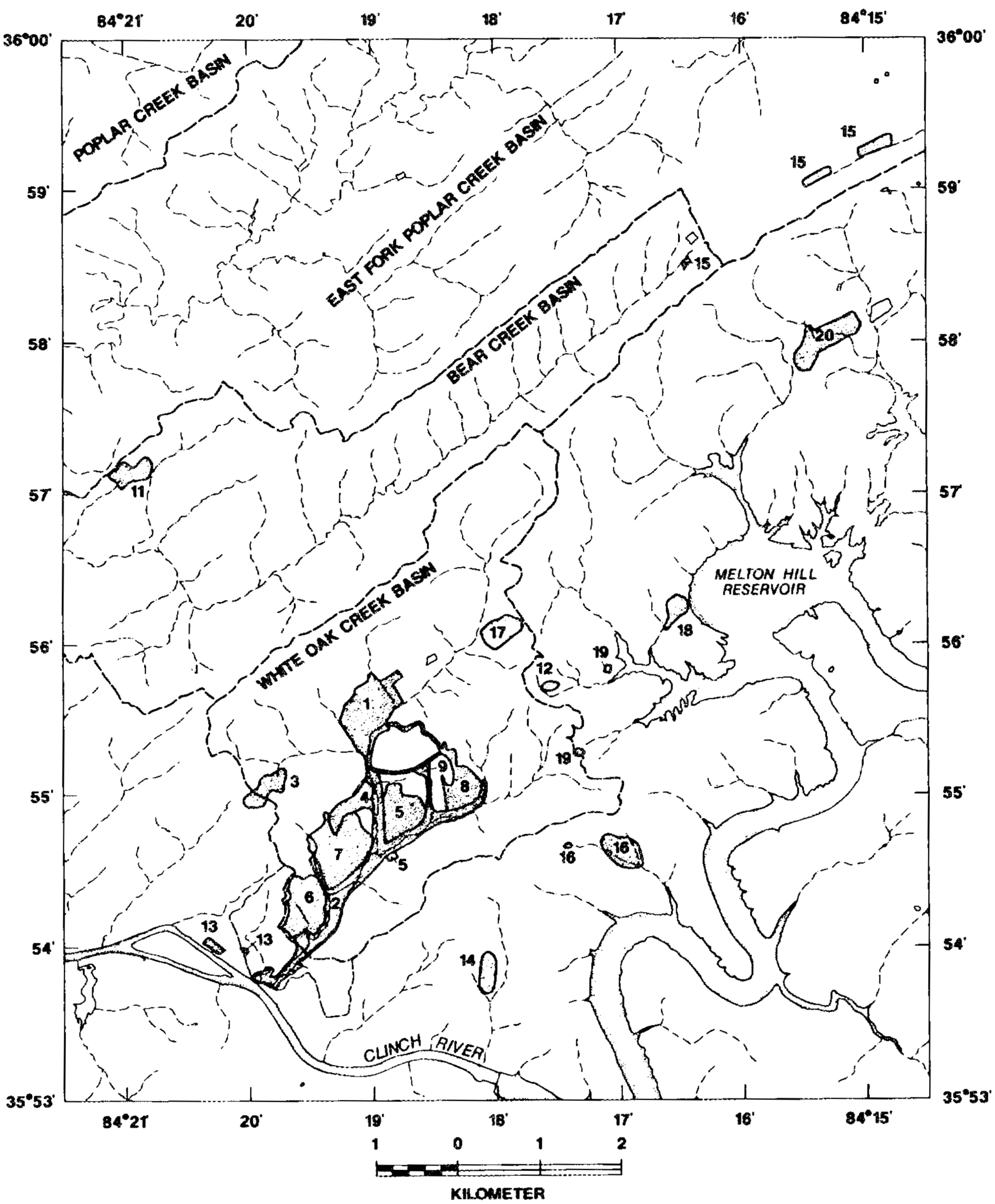

Fig. 1. Locations of Waste Area Groupings and hydrography.

analysis may be the best approach to parameter characterization. Cumulative probability graphs are used for analysis of parameter values in this report. These graphs require a fairly large amount of data but may show sample or population characteristics that otherwise would be obscure. Some deviations on cumulative probability graphs represent only an imperfect distribution of sample values, but others show a change in conditions at deeper levels in the aquifers or show some type of control on the range of parameter values. A correct interpretation of the significant deviations can contribute to an understanding of groundwater occurrence and flow paths.

The construction and use of probability graphs for analysis of geoscience data are fully described by Sinclair (1976). Basically the method consists of plotting sorted data values on cumulative probability paper; the data points are those that would be used for a cumulative histogram. If a straight line can be fitted to the data points, this line defines the cumulative normal density distribution of the population. A Gaussian population plots as a straight line on arithmetic probability paper, and a lognormal population plots as a straight line on logarithmic probability paper. The 50 percentile value of the line represents the arithmetic mean of a normal population and the geometric mean of a lognormal population. Similarly, values for the mean minus or plus one standard deviation can be read from the 16 and 84 percentile values of the line.

In the study area, all hydrologic parameters, which were plotted on probability paper, were lognormally distributed, but some parameters had a much larger data range than did others. Parameters with a relatively small range were plotted on a logarithmic scale, whereas the natural logarithms of other parameter values were plotted on an arithmetic scale. Graphically, both methods produce the same result. In the latter case, however, values such as the geometric mean must be calculated from e^x, where x is the value determined from the line.

Probability graphs provide a simple and rapid procedure for determining the type of distribution and for detecting abnormalities in the data. The disadvantages of probability graphs include the need for a fairly large amount of data and the possibility of an erroneous interpretation of any deviations from a straight-line trend.

3. PREVIOUS STUDIES AND CONCEPTS

The first study of geology and groundwater near ORNL was made by Stockdale (1951) in Bethel Valley near WAG 1. This report first noted, "In general, the water table is a subdued replica of the land surface, rising slightly below the hills, but occupying positions closer to land surface in the valleys" (pp. 50-51). The description of aquifer recharge in Stockdale's report (pp. 51-52) can be little improved, even today. Thus, he stated, "recharge of the aquifer can occur any place where water can infiltrate the soil. . . . It would appear that recharge can and does take place throughout the area, . . . [but] certain areas are more susceptible to recharge than others." The report, however, attributes local differences in recharge to a variety of factors including soil permeability, rock lithology, and degree of fracturing (pp. 51-52). It is now believed, as is discussed later, that the most important factor is the total period of time in which there is a perched water table in the stormflow zone, just beneath land surface.

Stockdale's (1951) description of aquifer discharge is also accurate. Thus, the report stated, "Records of water levels in wells adjacent to permanent streams . . . [show that] the streams are effluent with respect to groundwater" (p. 51), and it listed discharge to "springs, wet-weather seeps, the banks of White Oak Creek and its tributaries, and evaporation and transpiration" (p. 58). Stockdale correctly noted that during wet weather, perched water bodies discharge groundwater "in numerous seeps along the slopes of the valleys" (p. 58) and that during dry weather, groundwater discharge constitutes the base flow of the streams (p. 50).

The description of aquifer characteristics in the Stockdale (1951) report requires a minor update. The report stated, that "groundwater occurs under water table rather than artesian conditions" (p. 50). Later data have shown that (1) the water levels in deep wells are higher than the levels of the water-bearing fractures, (2) flowing wells occur in a few areas, and (3) the water levels in deep wells respond to earth tides and to loading caused by hydrofracture operations (Webster and Bradley, 1987, p. 65). These facts show that confined conditions occur at levels somewhat deeper than those of the shallowest wells.

Finally, Stockdale (1951, pp. 74-75) concluded that shale units in the Conasauga Group have fewer cavities and smaller permeabilities than do limestone units in the Chickamauga Group, and he recommended that future burial grounds be located in Melton Valley (pp. 75-76). As is discussed later, recent data show no significant differences in either the spatial frequency of cavities or the distributions of hydraulic conductivity for the Conasauga Group and Chickamauga group. Thus the main conclusion of the report was incorrect.

The concept of bedrock anisotropy and the movement of larger rates and quantities of groundwater in fractures parallel to geologic strike than in fractures normal to strike was first thoroughly discussed in a paper by de Laguna et al. (1958). The WAG 7 tests upon which these conclusions were based are reviewed by Webster and Bradley (1987, pp. 79-80). Similar results were later described for other areas of Melton Valley by Olsen et al. (1983) and Davis et al. (1984, pp. 77-81). As is discussed later, anisotropy may be represented by a difference in cross-valley and along-valley (strike parallel) hydraulic gradients.

The next reports to describe general principles of groundwater occurrence were those by McMaster and Waller (1965) and McMaster (1967). About 200 wells had been drilled in the ORNL area by this time, but nearly all wells were clustered near WAGs 1, 3, 4, 5, and 7, and the wells were used only for the monitoring of water levels. The conclusions in these reports are observations and inferences based on the flows of springs and streams. The descriptions of groundwater discharge to springs are valid, but some of the other interpretations have not been confirmed by later data.

McMaster and Waller (1965, p. 7) wrote, "The Knox Group of Chestnut Ridge is the principal aquifer in the White Oak Creek basin . . . Springs occurring along the base of Chestnut Ridge and in its valleys are the chief source of the base flow discharge of White Oak Creek. The Knox Group of Copper Ridge, however, is not known to discharge groundwater into this basin." In reference to the Chickamauga Group, the authors noted only, "Although a substantial quantity of water probably is stored in many small openings . . . within about 100 ft of the land surface, rates and quantities of water movement are, for the most part, relatively small" (p. 8). More information was given for the Conasauga Group. The authors said, "during the winter months, springs are common in the small valleys along the northwestern side of Melton Valley, . . . but during April and May discharge is greatly decreased as recharge is reduced and the water-bearing material is drained. During the summer months, very little water is discharged. Discharge of groundwater [from Conasauga units on the slope of Copper Ridge] is small even during the winter months; during the late summer months no discharge is known to occur" (p. 6). The McMaster and Waller report thus was the first to recognize the magnitude of groundwater discharge from the Knox Group and was the first report to describe differences in discharge from the Conasauga Group on the dip-slope and scarp-slope sides of the ridges. On the other hand, the differences between wet-weather and perennial springs are unclear, and the importance of perched water to the discharge of wet-weather springs was not recognized.

Descriptions of groundwater occurrence and flow in the McMaster and Waller (1965) report are inferences that are partly contradicted by recent data. The authors stated that in the Rome Formation, "The thin mantle of residual clay and the near-surface weathered bedrock zone having slightly enlarged openings probably account for the greater part of the water movement" (p. 6). Similarly, they believed "Groundwater in the Conasauga Group occurs principally in the weathered zone where openings along joints and bedding planes have been slightly enlarged by circulating water" (p. 6). For the Knox Group, they wrote, "The thick mantle of overburden appears to have a high infiltration capacity, minimizing overland runoff of rainfall and also serving as a reservoir of groundwater feeding underlying solution openings, many of which probably are of large size" (p. 7). They thought (p. 8) that most groundwater in the Chickamauga Group occurs in small openings in the bedrock but "rates and quantities of water movement are relatively small" (p. 8). Recent data (Moore 1988, Tables 1-2) show no significant differences in the mean hydraulic conductivities of (1) regolith and bedrock and (2) the Conasauga Group and Chickamauga group. The water-bearing characteristics of the Rome Formation and the Knox Group cannot be completely described at present, but available data indicate that these units are not greatly different than the Conasauga Group and the Chickamauga group. An exception is that large springs in the Knox Group are unexplained by present data.

More recent reports that describe general groundwater occurrence and movement are those by Webster (1976) and Webster and Bradley (1987). These studies are based upon information from WAGs 3, 4, 5, 6, and 7; nearly all data are from wells near burial grounds in the Conasauga Group. Both of these studies were mainly intended to describe the three-dimensional network of groundwater flow paths below the water table; flow paths that produce wet-weather springs above the water table were not considered. Webster (1976, p. 15) stated that the direction of groundwater movement in regolith "is based on the premise that in an unconfined aquifer system, groundwater moves by gravity downgradient in a direction generally normal to the water table contours." This premise was further developed in the report by Webster and Bradley (1987, p. 31), as follows: "To the extent that . . . [the regolith consists of decomposed rock below the water table] it would be reasonable to expect that groundwater flows between the fragments wherever a hydraulic gradient exists across the beds." This report then described tests (pp. 31-34) in which the largest concentration of a tracer moved normal to geologic strike and normal to the water table contours in two locales but moved along fractures parallel to geologic strike in other locales. The authors concluded, "the saturated interval of the regolith has the flow characteristics of a porous medium in some areas, whereas in other areas it has some of the flow characteristics" of fractured rock (p. 34). Webster (1976, p. 17) noted that below the weathered zone is "a thick zone of fractured rock where the major component of flow parallels [geologic] strike."

There are four problems with flow concepts in the Webster reports. First, Davis et al. (1984, pp. 23, 45-49) wrote that carbonate cement has been leached from rock layers in the weathered zone and that the regolith is very acidic (pH 4.0-5.5) with no calcium carbonate and small amounts of other soluble minerals. However, virtually all groundwater from shallow wells, as is discussed later, is a nearly neutral to moderately alkaline (pH 6.5-8.1), calcium bicarbonate type; total dissolved solids content is typically 90-500 mg/L. This evidence shows that virtually all groundwater below the water table, including that found in regolith, has flow paths through fresh rock. Second, in two tracer tests described by Webster and Bradley (1987, pp. 31-33) and in one tracer test described by Davis et al. (1984, pp. 77-80), the tracer eventually appeared in all observation wells. These results show lateral groundwater flows in all directions from strike parallel (and nearly parallel to water table contours) to strike normal. Third, recent data (Moore 1988, Table 1) show that the geometric means of hydraulic conductivity in regolith and bedrock are essentially the same. These results suggest groundwater occurrence in fractures for both regolith and bedrock. Finally, all estimates and calculations of storativity in recent reports, as discussed later, have values in the range $1x10^{-3}$ to $1x10^{-6}$ whereas values in the range 0.1-0.3 would be expected for porous media. Only a three-dimensional fracture network with complex vertical and lateral flow components can explain these data and these results.

Discussions of fracture aperture and permeability in the Webster reports are based on previous studies, on inference, and on a relatively few data from slug tests and aquifer tests. Few of the conclusions are confirmed by analyses of other data. Thus, for example, Webster (1976, p. 6) stated that the Knox Group and the Chickamauga group are unsuitable for waste storage because of the occurrence of cavities. Recent analyses of well logs, as is discussed later, show that the spatial frequency of cavities in limestone units of the Conasauga Group is about the same as in the Chickamauga Group; the Knox Group has a larger spatial frequency of cavities, but it is not greatly larger.

Webster (1976) also said, "A few studies of the Conasauga have reported that the greatest permeability is associated with the transition zone between the fresh and weathered rock" (p. 9), and he added, "the greatest permeability beneath the weathered zone is associated with the residue of the more soluble beds; a lower permeability is associated with fractures... that cross the bedding" (p. 16). Webster and Bradley (1987, p. 28) added that "The volume of openings represented by... fractures and solution cavities per unit volume of bedrock [in the Conasauga Group] is far less than that of the regolith and tends to decrease as depth increases." These authors concluded (p. 82) that groundwater flow in bedrock extends to a depth of more than 60 m but that total flow in fresh bedrock probably is less than in regolith and weathered rock combined. Additional data (Moore 1988, Tables 3-4) show little difference in the geometric mean of hydraulic conductivity for water-producing zones to a depth of about 20 m, regardless of whether the aquifer is regolith or bedrock. As is discussed later, the mean along-valley hydraulic gradient is lower than the mean cross-valley gradient, but this difference might be caused by a different spatial fracture frequency; it need not be caused by a difference in fracture permeability.

A number of other reports have made important contributions to hydrogeologic characterization in the ORNL area. Sledz and Huff (1981) made an intensive study of the orientation and spatial frequency of fractures on rock outcrops; this report is also a good source of information on fracture apertures and calculated porosity in cores. Rothschild et al. (1984, p. 97) produced an interesting water budget for an area northeast of WAG 8 in Melton Valley but did not separate groundwater discharges above and below the water table. Davis et al. (1984, pp. 36-59) included a definitive study of the physical and geochemical properties of soils in an area of WAG 6; these authors were also the first to recognize seasonal changes in the concentrations of chemical constituents in groundwater. Ketelle and Huff (1984, pp. 131-135) were the first to document rapid groundwater flow through a cavity system in the Knox Group; this report also includes the best available information on the hydrogeology of the Knox Group. The most accurate measurements of transmissivity and storativity were those made by Smith and Vaughan (1985) for two aquifer tests. Tucci (1985) calibrated a two-layer (regolith and bedrock) finite-difference model of groundwater flow in Melton Valley and described results of a sensitivity analysis. The model is probably not representative of the study area because a large difference in transmissivity was assumed between regolith and bedrock, but the results show that modeling is a feasible approach to the estimation of some parameter values. Dreier et al. (1987) showed that local groundwater flow in saprolite occurs along at least three different fracture sets but that flow may be concentrated at fracture intersections. Finally, Moore (1988) analyzed about 380 hydraulic conductivity values by type of aquifer material, geologic formation or group, and depth of test.

4. GEOLOGIC FRAMEWORK

The study area is a part of the Valley and Ridge Physiographic Province and a part of the Appalachian thrust belt of eastern Tennessee. Rocks of Cambrian to Ordovician age outcrop within two thrust sheets of a sequence. The Copper Creek thrust fault comes to the surface on the northwestern slope of Haw Ridge; this fault sheet underlies Copper Ridge, Melton Valley, and Haw Ridge. The White Oak Mountain thrust fault has several traces between Pine Ridge and East Fork Ridge; this fault sheet underlies Bethel Valley, Chestnut Ridge, and Bear Creek Valley. The ridges in this area are about 90 m higher than the valleys.

The parallel ridges and valleys are a result of differential erosion, which has been affected by the underlying structure of folds and faults. The strike of the rocks generally parallels the axes of the ridges and averages about N55°E; the dip varies but commonly is 30-40°SE (Stockdale 1951, p. 16). The main geologic units in the study area are the Rome Formation and the Conasauga Group, of Cambrian age; the Knox Group, of both Cambrian and Ordovician ages; and the Chickamauga Group, of Ordovician age. A small amount of alluvium occurs as floodplain and terrace deposits.

4.1 BEDROCK AND REGOLITH

The Rome Formation consists mainly of siltstone, shale, and sandstone; dolostone beds occur locally (McMaster 1963, p. 6). A distinctive characteristic is bright variegated colors including green, maroon, red, violet, purple, yellow, and brown. The Rome Formation is a relatively resistant unit that forms some of the ridges in the study area. As described by McMaster (1963, p. 8), these ridges are "typically narrow, steep sided, and broken by many [shallow but] closely spaced wind and water gaps . . . ." Regolith above Rome bedrock generally is less than 4 m thick and is a light-colored sandy silty clay containing scattered siltstone and sandstone fragments.

The Conasauga Group consists of sequences of calcareous shale, siltstone, shaley limestone, and limestone. From lithologic similarities and differences in cores, six formations were identified in the Conasauga Group of Melton Valley (Haase and Vaughan 1981). These formations include, from oldest to youngest, the Pumpkin Valley Shale, Rutledge Limestone, Rogersville Shale, Maryville Limestone, Nolichucky Shale, and Maynardville Limestone. Later work by R. H. Dreier et al. (ORNL, written communication, 1987) has shown that these formations are mappable in the subsurface. However, with the exception that cavities occur only in limestone layers, the formations are similar hydrologically, and they are not treated separately for the purposes of this report. The Conasauga Group forms valleys, slopes, and hillocks in the study area. Scarp slopes on the northwestern sides of the ridges are steep, but a line of knobs and shallower slopes occur on the dip slope that forms the southeastern sides of the ridges. A dull tan to gray and olive green color is characteristic of the Conasauga Group. Regolith is considerably thicker beneath the hillocks than in the valleys and gradually changes from a silty clay near the surface to a leached saprolite at depths of 1-15 m. Top of bedrock is commonly considered to be auger refusal or the first layer with a lime content.

The Knox Group was mapped on a part of Chestnut Ridge (Ketelle and Huff 1984, pp. 14-17) as four formations including, from oldest to youngest, the Copper Ridge Dolomite, Chepultepec Dolomite, Longview Dolomite, and Newala Formation. For the purposes of this report, however, the Knox Group is considered to be a single hydrologic unit. This unit grades from a massive, dark gray, very cherty dolostone near its base to a less massive, lighter gray, less cherty dolostone near its top (McMaster 1963, p. 10). The bedrock surface typically has a pinnacle and trough topography and is overlaid by 3-4 m of regolith in valleys and by up to 50 m of regolith beneath hills. This regolith has a characteristic orange red to deep red color and contains more or less abundant chert of silt-sized to cobble-sized particles in a matrix of 10-70% clay (Ketelle and Huff 1984, p. 32). The cherty regolith is resistant to erosion, and the Knox Group forms broad, dissected ridges, which are steepest on the northwest-facing scarp slopes. Shallower slopes occur on the dip slope and within the outcrop belt, where the topography has a rounded appearance. Sinkholes are common to abundant, and springs, some of large size, are common near the bases of the ridges. Swallow holes and other karst features are infrequent to rare.

The Chickamauga Group was mapped as eight informal units, designated "A" to "H", in a part of Bethel Valley (Stockdale 1951, pp. 21-25), and these units were described in more detail by Lee and Ketelle (1988). For the purposes of this report, however, the Chickamauga is considered to be a single hydrologic unit that consists of sequences of limestone, shaley limestone, and limy siltstone. The color is generally gray to bluish gray or olive green, but a few layers are distinctively maroon. Chert is common in the basal layers of Bethel Valley. The regolith is thin, generally less than 3 m thick, and consists of a yellowish or reddish clay with variable amounts of chert fragments (McMaster 1963, p. 14). The Chickamauga Group forms valleys and low hillocks, which are prominent on the northwestern side of Bethel Valley. Sinkholes are small and spatially infrequent; other karst features are rare.

Three piezometer wells near WAG 13 show that recent alluvium along the floodplain of Clinch River consists mainly of brown silt and clayey silt about 8 m thick. Gravel was mixed with the silt in samples from one well. Along White Oak Creek and Melton Branch, alluvium is less than 1 m thick, fine grained, and difficult to distinguish from residual and colluvial soils. Subangular to rounded terrace gravels have been commonly described as mixed with clay and silt in near-surface samples from many wells. At one location, near well 270 in WAG 6, a terrace deposit about 5 m thick was found to consist of "well-rounded pebbles and cobbles of quartzite and other resistant materials along with sand, silt, and clay . . . . The pebbles and cobbles, which are not representative of the surrounding bedrock, are thought to be remnants of an ancient floodplain of the nearby Clinch River" (Lomenick and Wyrick 1965, p. 5)

Hydrologically, the floodplain and terrace materials are considered to be a part of regolith for the purpose of this report. However, rounded gravel, pebbles, and cobbles are also reported as fill material in many bedrock cavities. As is discussed later, a coarse-grained fill cannot be explained by the hydraulic conductivities and groundwater velocities in cavities below the water table. These cavities thus might represent fossil flow paths of the same age as the terrace deposits.

A cumulative probability graph of regolith thickness in observation wells (Fig. 2) shows that these data generally represent a single lognormal population in which the geometric mean is 3.9 m and the range from the mean minus one to plus one standard deviation is 2.0-7.3 m. The data points are well fitted to the line except at the extreme low end of the graph; this one deviation represents a few wells with a regolith thickness of less than 1 m and can be ignored for most purposes. Virtually all of the wells are in the Conasauga and Chickamauga Groups; the graph is not representative of regolith thickness in the Knox Group, and it may not be representative of the Rome Formation. A classification of the data by geologic unit shows that 64% of the wells with a regolith thickness less than the mean minus one standard deviation are in the Chickamauga Group but that 74% of the wells with a regolith thickness more than the mean plus one standard deviation are in the Conasauga Group. This difference probably represents the inclusion of saprolite in the regolith thickness of shaley Conasauga units.

4.2 FAULTS

The hydrologic importance of faults in the study area is only partly understood at present. Wells penetrating the Copper Creek thrust fault have been reported as having impermeable zones of brecciated rock and gouge at the level of the fault and in a zone 5-20 m above the fault (Stockdale 1951, p. 41; Haase et al. 1985, pp. 67-69). A different situation occurs in WAG 11 near one branch of the White Oak Mountain thrust fault. Thirteen of 14 piezometer wells in this area have a regolith thickness (7.3-22 m) greater than the geometric mean plus one standard deviation for all wells. Also, there is little difference in water level elevations for wells on opposite sides of the fault, and the configuration of the water table suggests that a zone near the fault may be a conduit for groundwater flow.

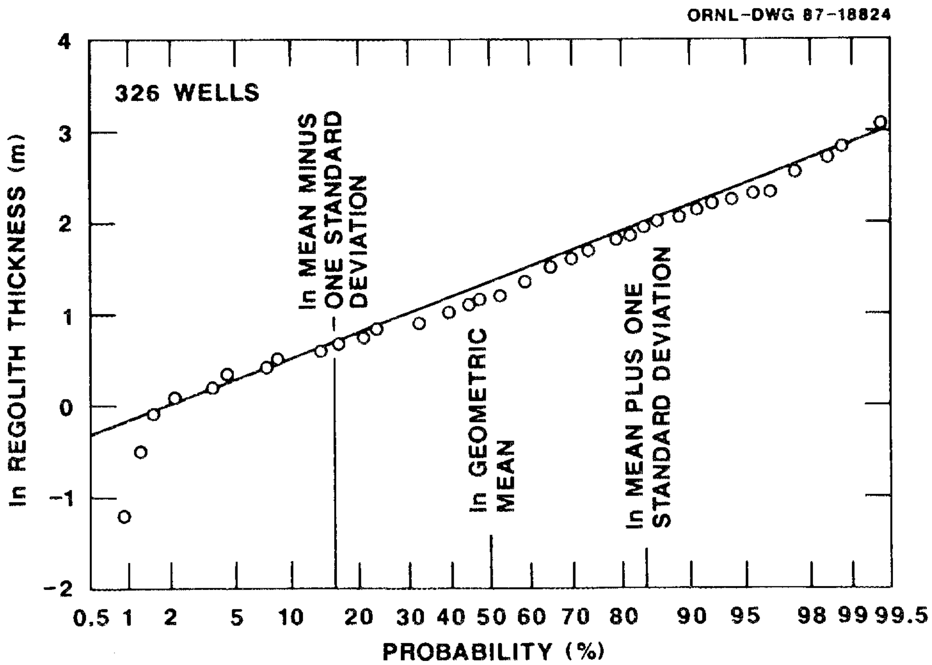

Fig. 2. Cumulative probability graph of regolith thickness in wells.

High-angle faults also occur in the region. One such feature probably occurs along a valley alinement near the Y-12 Plant, which is outside the area of this study. Wells along this alinement had an unusually large initial production of water during drilling, but these water yields soon decreased greatly [Z. C. Bailey, U.S. Geological Survey (USGS), personal communication, 1987]. This response to pumpage suggests a flow system (perhaps a fossil system) in which nearby openings in the rocks can supply more water to wells than can be sustained by more distant openings. A high-angle fault also occurs in the water gap of White Oak Creek through Haw Ridge, but the geometric mean of hydraulic conductivity for wells in this valley (0.036 m/d) is almost exactly the same as the mean for all wells (0.041 m/d) in the study area. Many smaller and unmapped faults of this type almost certainly occur, but available data do not show an association of unusually large well yields with indications of faulting, such as intensely fractured rocks, unusually thick regolith, and karst features. A tentative conclusion is that groundwater conduits can occur along and near faults in the study area but that such features are uncommon and may be rare.

4.3 OTHER FRACTURES

Groundwater flow paths are mainly through joints and other fractures in bedrock. Fractures may also be the main pathways for groundwater flow in regolith below the root zone. As is discussed later, groundwater may actually flow through a relatively few open channels within fractures, the sides of which are otherwise in contact and much less permeable.

Most fractures are short, a few cm to 1 m in length, but various joint sets form intersecting systems (Sledz and Huff 1981, p. 12). Many fractures are bedding-plane parallel or strike parallel; both sets occur in some areas. A less common, orthogonal fracture set is parallel to the dip of the beds. A fracture system consisting of these two or three sets may be presumed to occur in any locality, and other sets may also be present. Most fractures are steeply dipping (R. B. Dreier, ORNL, personal communication, 1988).

The fractures are abundant. Measurements on outcrops and on cores of the Conasauga Group showed a density of 10-15 joints per meter in shale and 6-40 joints per meter in siltstone (Sledz and Huff 1981, pp. 44-52). Some fractures have larger apertures (gap widths) than others, and the rate of groundwater flow is determined by the cube of the aperture (Witherspoon et al. 1980). Measurements made on mineral-filled joints in cores (Sledz and Huff 1981, p. 74) showed that apertures are 0.1-0.7 mm in siltstone and <0.1-0.2 mm in shale of the Conasauga Group. As is discussed later, fracture permeabilities are represented by the statistical characteristics of two populations.

4.4 CAVITIES

Cavities in bedrock are formed by solution, abrasion, and a combination of these two processes. Enlargement of fractures begins with slow solution of a part of the adjacent rock mass or of the rock cement. Openings enlarged above some critical size permit turbulent groundwater flow. Physical erosion by abrasion then increases the rate of cavity enlargement while turbulent flows remove at least part of the resulting detritus. Any remaining detritus accumulates at the bottom of the cavity and partially protects this rock surface against further erosion. Thus, most larger cavities develop mainly by abrasion and by upward stoping.

Some rock layers and lithologies along a groundwater flow path are more easily eroded than others. Thus, the cross-sectional area of a cavity may change considerably from one location to another. Mean groundwater velocity is less in the larger sections, especially near cavity walls, and solution may again become the dominant process in further enlargement of these reaches. Also, more detritus may accumulate in the larger sections. If abundant detritus is available through time, a profile may develop in which a small open cavity overlays a wedge of detritus; the axis of this wedge generally conforms to the location of the original fracture. Because most fractures are steeply dipping but not vertical, one well may intercept a cavity filled with detritus, another well a short distance away may intercept an open cavity, and a third well a short distance in the other direction may intercept a slightly enlarged section of the original fracture. The wedge of detritus is commonly thicker than the open part of the cavity, and thus most cavities may be reported as filled. This is apparently the situation in the ORNL area. Cavity fill material in this area has been described as consisting of clay, mud, mixed clay and gravel, sand, chert, limestone chips, rounded gravel, or limestone pebbles and cobbles; twigs were recovered from one cavity.

So-called solution cavities in the Chickamauga Group were first described by Stockdale (1951, p. 41). Since then, cavities have been reported in all other rock units with limey layers. A general principle of cavity occurrence is that if everything else is equal, the largest cavities are found in the purest and most massively bedded limestones. This principle is generally applicable to rock units in the ORNL area. Cavities in the Conasauga Group have been reported only in the Maryville Limestone, Nolichucky Shale, and Maynardville Limestone. All three of these formations contain limestone layers, and the cavities are presumed to occur in these layers. Similarly, two cavities in Rome Formation bedrock may occur in dolostone layers, which have been described in the upper part of this formation (Stockdale 1951, p. 17).

A few cavities in the study area have been reported as zones of high water velocity (which washed away sandpack material) in regolith, just above top of bedrock. Apparently this is a form of piping, which occasionally occurs in unconsolidated sediments near bedrock or near a point of groundwater discharge. Two cavities in the Rome Formation and three cavities in the Conasauga Group are of this type. The piping at the base of the regolith may occur along upward extensions of enlarged fractures in the bedrock.

The records of 802 wells in the study area show that only 97 wells (12%) intercept a cavity, and most of the smaller number intercept only one cavity. However, there are some distinctive differences in cavity occurrence among the geologic units (Table 1). None of the Rome Formation wells and only one of the Chickamauga Group wells (8%) intercept more than one cavity. In the Conasauga Group, 27 wells (46%)

Table 1. Number of cavities in wells grouped by geologic unit

Geologic unit	Number of wells	Number of wells with one or more cavities: One cavity	Two cavities	Three cavities	Four cavities	Five or more cavities
Chickamauga Group	13	12	1	0	0	0
Knox Group	21	5	10	2	3	1
Conasauga Group	59	32	16	9	2	0
Rome Formation	4	4	0	0	0	0
Population	97	53	27	11	5	1

intercept more than one cavity, but only 11 wells (19%) intercept more than two cavities, and none intercept more than four cavities. Wells in the Knox Group show the largest differences with the distribution of cavities in the population. More Knox wells intercept two cavities than one cavity or any larger number. A total of ten cavities were reported in one Knox well, but only four wells (19%) intercept more than three cavities. Thus, multiple cavities are rare except in the Conasauga and Knox Groups, and more than three cavities in these units are uncommon.

Only the vertical dimension of cavities (called height for the purposes of this report) can be obtained from well records. A cumulative probability graph of these data (Fig. 3) is somewhat irregular, mainly because many cavity heights were recorded as integers rather than as decimal numbers. Nevertheless, a single straight line can be satisfactorily fitted to the data points. This fit shows that cavity heights represent a single lognormally distributed population and suggests that all cavities were formed by the same processes. The geometric mean of cavity height is 0.59 m, and the range from the mean minus one to the mean plus one standard deviation is 0.18-2.0 m.

A grouping of cavity heights by geologic unit (Table 2) shows some interesting differences. The cavity with the largest height is in the Knox Group, and, although large cavities also occur in the Conasauga Group and the Chickamauga group, the geometric mean height of cavities in the Knox Group is almost twice as large as that in the Conasauga Group. One-tail Student's t-tests of log-transformed (normalized) data show that the geometric mean height of cavities in the Knox Group is statistically larger than the mean height in both the Conasauga Group and the population at the 1% level of significance. There are no statistically significant differences in the geometric mean heights of the other units.

The shape of cavities in the study area is unknown. In central Tennessee, many cavities have a semicircular (domed upward) section; this is also the stable shape for piping. However, circular to rectangular (with elongation in the direction of the controlling fracture) sections have also been observed.

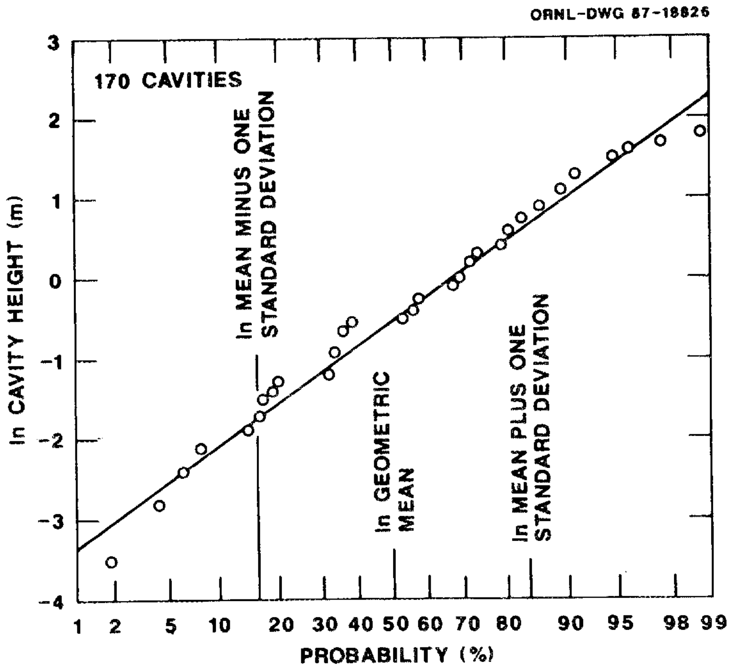

Fig. 3. Cumulative probability graph of cavity heights (vertical dimension of cavities).

Table 2. Distribution of cavity heights grouped by geologic unit

Geologic unit	Height of cavities (m)					
	Number of values	Geo-metric mean	Mean minus one standard deviation	Mean plus one standard deviation	Minimum value	Maximum value
Chickamauga Gp.	14	0.90	0.25	3.3	0.061	6.9
Knox Group	53	1.0	0.34	3.1	0.030	8.5
Conasauga Group	99	0.51	0.16	1.7	0.030	5.9
Rome Formation	4	0.81	NA[a]	NA[a]	0.30	1.5
Population	170	0.59	0.18	2.0	0.030	8.5

[a]Not applicable because of small number of wells.

The depths of cavities in the well records were denoted as the midpoints of the open intervals. A probability graph of cavity depths (Fig. 4) shows that a single straight line can be satisfactorily fitted to these points. This fit shows that the data can be considered to be samples from a single population. Thus, cavity depths are lognormally distributed with a geometric mean depth of about 14 m and a range of 5.8-32 m from the mean minus one to the mean plus one standard deviation.

An analysis of cavity depths by geologic unit (Table 3) shows that the geometric mean depth of cavities in the Knox Group is much larger than those in the Rome Formation, Conasauga Group, and Chickamauga Group. This result was expected because of the larger regolith thickness in the outcrop area of the Knox Group and is not hydrologically significant. The similarity of geometric mean values for the depths of cavities in the Rome, Conasauga, and Chickamauga was not expected. Apparently, the factors that determine cavity depth are nearly the same over the entire area underlaid by these three units. The other statistical indices suggest that the curvature of data points near the center of Fig. 4 may be caused by the fact that only 16% of cavities in the Conasauga and Chickamauga Groups are deeper than 16 m and only 16% of cavities in the Knox Group are shallower than 21 m.

The vertical spatial frequency of cavities in the study area (Table 4) was determined by dividing the total height of cavities in any group of wells by the total length of borehole in rock. The lateral spatial frequency of the cavities could have been estimated by dividing the number of wells that intercept one or more cavities by the total number of wells in any group. However, results might have been biased by differences in the average depth of wells that intercept cavities and the average depth of wells that do not. Instead, each well was assigned a weight based on its depth and on the probability (Fig. 4) of its intercepting a cavity at any level above this depth. For example, a well 20 m deep ($\ln 20 = 3.0$) was given a weight of 0.73 based on the relationship shown by the fitted line on Fig. 4. Lateral spatial frequency was then calculated by dividing the total weight of wells that intercept one or more cavities by the total weight of all

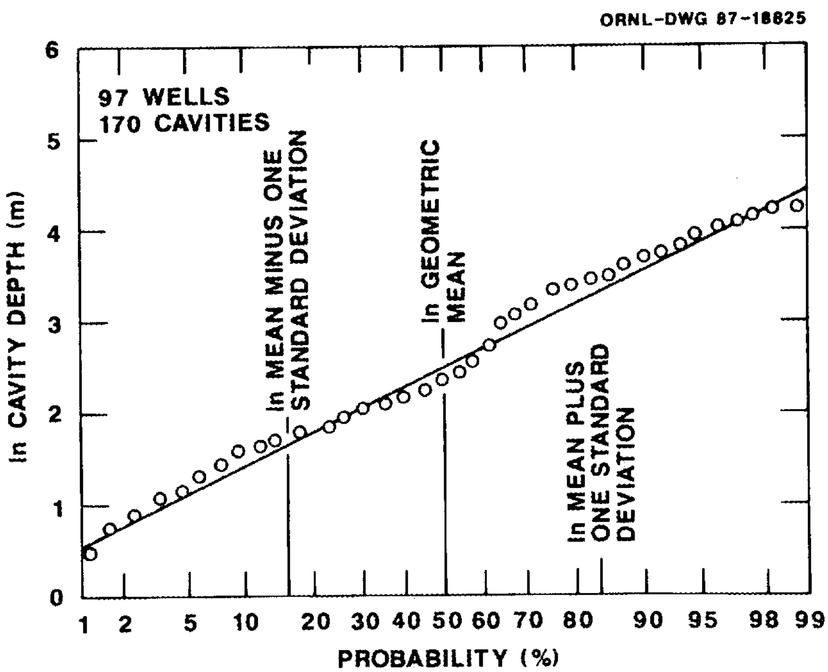

Fig. 4. Cumulative probability graph of cavity depths below land surface

Table 3. Distribution of cavity depths grouped by geologic unit

		Depth of cavities (m)				
Geologic unit	Numbers of values	Geo-metric mean	Mean minus one standard deviation	Mean plus one standard deviation	Minimum value	Maximum value
Chickamauga Group	14	9.7	5.7	16	4.8	21
Knox Group	53	34	21	53	19	96
Conasauga Group	99	8.3	4.3	16	1.2	71
Rome Formation	4	12	NA[a]	NA[a]	6.1	26
Population	170	14	5.8	32	1.2	96

[a]Not applicable because of small number of wells.

Table 4. Spatial frequency of cavity occurrence grouped by geologic unit

Geologic unit	Number of wells	Lateral spatial frequency	Vertical spatial frequency
Chickamauga Group	175	0.13	0.018
Knox Group	50	0.43	0.058
Conasauga Group	552	0.12	0.012
Rome Formation	25	0.11	0.011
Population	802	0.16	0.019

wells in the group. A lateral spatial frequency of 0.12, for example, means that if wells in the Conasauga Group were drilled deeply enough to intercept all possible cavities, 12% of these wells would intercept one or more cavities.

An analysis of the spatial frequency of cavity occurrence grouped by geologic unit (Table 4) shows that cavities in the Knox Group are three to five times as common as cavities in the other geologic units. The only other significant difference between the units is that cavities in the Chickamauga Group have about 1.5 times the vertical spatial frequency of cavities in the Conasauga Group. This difference might be caused by the larger geometric mean height (Table 2) of cavities in the Chickamauga Group, even though multiple cavities are more likely to occur in wells in the Conasauga Group.

5. GENERAL HYDROLOGY

All water in the study area comes from precipitation or is imported by pipeline from Melton Hill Lake. All water leaves the area as streamflow or is consumed by evapotranspiration. The land surface is permeable, and nearly all precipitation infiltrates. About 57% of this water is then evaporated or transpired. The remaining water moves laterally to nearby streams and springs. All groundwater is discharged; none is known to leave the area as underflow.

5.1 PRECIPITATION AND EVAPOTRANSPIRATION

Mean annual precipitation in the period 1954-83 was 1330 mm (52.2 in.) for stations near ORNL; the minimum and maximum amounts in this same period were 897 and 1900 mm (35.3 and 74.8 in.) of water (Webster and Bradley 1987, p. 13). The 1986-88 water years were unusually dry; only 876 mm (34.5 in.) of precipitation in 1986 and 1060 mm (41.7 in.) in 1987 were recorded by U.S. Geological Survey at a station in WAG 5. The wettest months generally are January through March, and the driest months are August through October; in these periods, mean monthly precipitation at the Oak Ridge Station of the National Oceanic and Atmospheric Administration (NOAA) is 130-160 mm (5.3-6.2 in.) and 74-96 mm (2.9-3.8 in.), respectively. The monthly extremes for the Oak Ridge Station are 340 mm (13.3 in.) for January 1954 and 13 mm (0.5 in.) for August 1953 (NOAA 1974, p. 378). The 1986 water year was dry; only 22 mm (0.9 in.) of precipitation was measured in WAG 5 during January, and 3 months (December, April, and June) had less than 51 mm (2.0 in.) of precipitation. October 1987 was another dry month; 18 mm (0.69 in.) of precipitation was recorded in WAG 5. The average frequencies of occurrence for various precipitation intensities over periods of 30 min to 24 h are shown by McMaster (1967, Fig. 3).

Droughts lasting 7 d occur about 17% of the time, but droughts lasting 15 d occur, on an average, only 1.8% of the time (McMaster 1967, Fig. 5). In the 1986 water year, the longest droughts at the WAG

5 station were 18 d in January, 15 d in April-May, 15 d in June, and 17 d in July. A 20-d drought at this station occurred in July-August 1987.

An average 764 mm (30 in.) of water is consumed by evapotranspiration. Based on pan evaporation measurements at Norris, Tennessee, about 75% (571 mm or 22.5 in. of water) of the evapotranspiration occurs during a 6-month period from April through September [Tennessee Division of Water Resources (TDWR), 1961 p. 18]. The growing season, when potential evapotranspiration is highest averages 220 days, from April 1 to November 5 (NOAA 1974, p. 373). A water balance graph for Rogersville, Tennessee, shows that potential evapotranspiration exceeds precipitation for 5 months, from May through September, and that the main period for replenishment of the soil moisture deficit is October 1 to November 10 (TDWR 1961, Fig. 4).

5.2 STREAMFLOW

Mean annual runoff for streams in the ORNL area is 566 mm (22.3 in.) of water, not including an estimated 180 L/s of effluent that is imported from Melton Hill Reservoir and discharged to White Oak Creek and Melton Branch. Excluding the imported water, nearly all streamflow is discharged groundwater. Increases and decreases in streamflow are accompanied by changes in the total length of flowing channels. Numerous source areas appear after periods of intense or prolonged precipitation and disappear after a few days to a few weeks of dry weather.

Streamflow and runoff depend upon amounts and changes in precipitation, evapotranspiration, and groundwater storage. Average quarterly runoff from the Oak Ridge area (McMaster 1967, p. 10), as a percentage of mean annual runoff, is shown below.

Quarter	Percentage of annual runoff
October-December	17
January-March	49
April-June	23
July-September	11

A seepage run (closely spaced measurements of streamflow) made by U.S. Geological Survey on June 28 in the 1988 water year showed that natural streamflow in the White Oak Creek basin was near zero; nearly all of the 165 L/s of streamflow below the confluence of White Oak Creek and Melton Branch was effluent from imported water (H. H. Zehner, USGS, personal communication, 1988). Groundwater flow at elevations above the valley floor continued, as was shown by water levels in observation wells, but this water was captured by evapotranspiration before reaching the stream channels.

5.2.1 Overland Flow

Small amounts of overland flow occur in the study area as a result of precipitation on urban facilities, wetlands, water bodies, and barren landscape features. Roofs, roads, and parking lots constitute about 25% of the land cover at the ORNL main plant (WAG 1) and ORNL services area (WAG 17). Most of the collected runoff from these areas is carried to the streams by storm sewers. Smaller concentrations of urban facilities are in WAGs 8, 9, and 18. Forested wetlands occupy a few small areas near the channels of White Oak Creek and Melton Branch. Seasonal wetlands are the wet-weather seeps and springs, which occur on concave slopes, and the downstream drainage channels. The main water bodies are White Oak Lake, the perennial channels of White Oak Creek and Melton Branch, and holding ponds in WAGs 1 and 8. Barren landscape features include dirt roads, other bare soil areas, and rock outcrops. The combined total area of overland flow is about 2% of the White Oak Creek basin.

5.2.2 Subsurface Flow

The subsurface flow of water that eventually is discharged to streams occurs both in a shallow zone just beneath land surface and in a deeper zone below the water table. Groundwater flow in the shallow zone is called "stormflow" for the purposes of this report. The word "interflow" could have been used, but this term is commonly applied to the perennial flow of perched groundwater at somewhat deeper levels. Stormflow in the study area is transient, a few days to a few weeks after periods of precipitation, but it continues longer near valleys than on hills and ridges. The processes that recharge both zones are infiltration and downward percolation.

A relatively small number of infiltration tests have been made on forested soils near ORNL (Watson and Luxmoore 1986; Wilson and Luxmoore, in press). Based on these data, a cumulative probability graph (Fig. 5) shows a lognormal distribution. The geometric mean infiltration rate is 8.8 m/d (14 in./h), and the range from the mean minus one to the mean plus one standard deviation is 3.2-23 m/d (5.2-38 in./h). One-hour precipitation intensity does not exceed 1.9 m/d (3.1 in./h) in the study area (McMaster 1967, p. 8). The infiltration data thus show that virtually all precipitation is readily absorbed by forest soils. Infiltration tests have not been made in grassed areas or in mixed grass, sedge, and brush areas. However, evidence of overland flow (lodged vegetation and matted detritus) has not been observed in these areas except in the seasonal wetlands. Average infiltration rate is apparently larger than precipitation intensity during storms and might be in the range 1-3 m/d (1.6-4.9 in./h).

Infiltration tests have also been made in areas where A- and B-horizon soils had previously been removed (Luxmoore et al. 1981, p. 688; Davis et al. 1984, p. 72). A cumulative probability graph of these data (Fig. 6) shows a lognormal distribution, a geometric mean infiltration rate of 0.025 m/d (0.041 in./h), and a range from the mean minus one to plus one standard deviation of 0.0082-0.080 m/d (0.013-0.13 in./h).

Other infiltrometer data were obtained at depths of 2.4-3.0 m and 6.0-12 m in boreholes on Chestnut Ridge (as summarized in Ketelle and Huff 1984, pp. 75-77). A cumulative probability graph of these 38 values has several deviations, but the overall linearity of the points shows that the data are lognormally distributed. The geometric mean of hydraulic conductivity is 0.0030 m/d, and the range from the mean minus one to the mean plus one standard deviation is 1.5×10^{-4} to 0.061 m/d. These values are smaller than those from infiltration tests in the C-horizon soils, and there are several possible explanations.

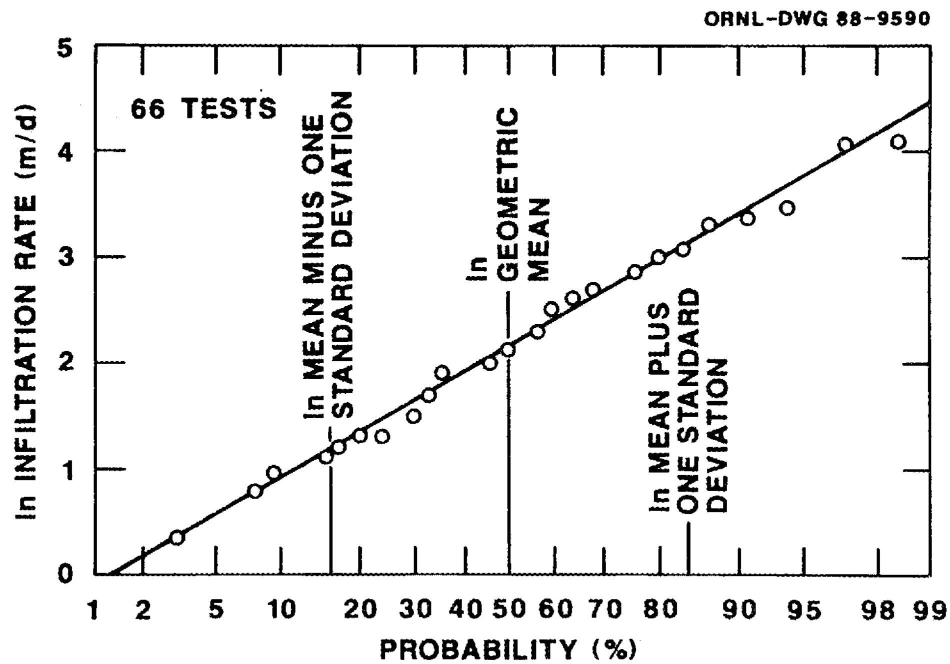

Fig. 5. Cumulative probability graph of infiltration rate in forested areas.

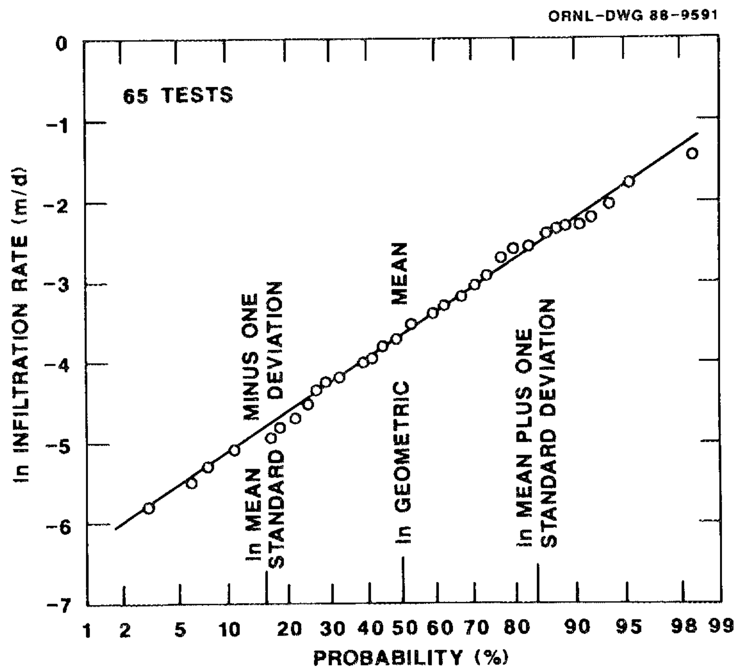

Fig. 6. Cumulative probability graph of infiltration rate in C-horizon soils.

First, hydraulic conductivity might decrease at larger depths in the vadose zone. Second, a thin stormflow zone might have been created by grass roots before infiltration tests were run on the C-horizon soils. Third, well construction procedures might have sealed some openings adjacent to the boreholes on Chestnut Ridge. A calculation of vertical water flux, as described later, suggests that the vadose zone is anisotropic and that vertical permeability may be only about 0.1 times as large as lateral permeability. Thus, the first two explanations, above, are more likely than the third. If the borehole infiltrometer data are correctly interpreted, the result shows that regolith permeability below B-horizon soils is 2900 times smaller than that of the surface layer in forested areas and an estimated 670 times smaller than that in grass and brush areas. Additional data are needed to clarify permeability differences between the surface layer and lower levels in the regolith. It is clear, however, that the average permeability of the surface layer is at least 1000 times larger than that at deeper levels. This contrast in permeability is the factor that causes lateral flows of groundwater in the surface layer.

Analyses and a model of tritium concentrations in headwater streams near WAG 6 show that stormflow is about 95% of total streamflow at the time of hydrograph peaks; stormflow discharged to wet-weather streams accounts for about 65% of total tritium release from this burial ground (D. K. Solomon, ORNL, written communication, 1988). Near major streams, the water table is within the stormflow zone, and additional groundwater is discharged to these streams. The shallow stormflow zone also constitutes the water reservoir for nearly all evapotranspiration. This zone is too thin to supply significant quantities of water to wells, but lateral stormflow can be intercepted by shallow trenches.

The remaining subsurface flow is water that percolates downward to the water table and moves slowly through fractures to discharges at seeps, springs, and streams. Nearly all groundwater movement below the water table can be described as slow seepage.

6. GROUNDWATER OCCURRENCE

Groundwater flow occurs in a shallow zone just beneath land surface and in a deeper zone that extends from the water table to the base of fresh water. Transient lateral flows of water probably are rare in the intervening vadose zone. The shallow zone generally corresponds to the vegetation root zone. Bouwer (1978, p. 264) noted that root depths vary from about 0.5 m for grasses and shallow-rooted crops to 1-2 m for most field crops to 3 m for small to medium trees. However, roots are concentrated in the upper layer of soil for all types of vegetation. Similarly, as noted previously, infiltration permeability is much smaller at depths of 2.4-3 m than at land surface. The shallow stormflow zone extends to depths of 1-2 m; the uppermost part of this zone is probably the most permeable, but the perched water table generally is in the lower part of the zone.

The water table generally occurs near the regolith and bedrock contact but is considerably above or below this contact in some areas. Brine that probably is connate occurs at depths of about 150 m in Melton Valley. Elsewhere, however, brine does not occur in wells at depths of 120 m in Bear Creek Valley, and only two wells at this depth produce an alkaline sodium carbonate water, which may be a transitional type (Z. C. Bailey, USGS, personal communication, 1988). Neither brine nor a sodium carbonate water type has been identified in wells up to 75 m deep in Bethel Valley. Thus the base of fresh water has not been determined in most of the study area.

6.1 STORMFLOW ZONE

Very little stormflow is intergranular in the shallow groundwater zone. The porosity of clay soils is generally about 50%. However, Watson and Luxmoore (1986, p. 581) found that macropores and mesopores, which together occupy only 0.2% of the soil volume, account for 96% of the infiltration. Macropores and mesopores are not completely understood but are connected voids that may have various causes, including biochanneling, cracking, and aggregation of soil particles. As described previously, the average infiltration rate in the stormflow zone is probably in the range 1.0-8.8 m/d. Infiltration rate is commonly about half the saturated hydraulic conductivity (Bouwer 1978, p. 237) because permeability is reduced by trapped air in materials above the water table. However, infiltration tests in the study area were run with procedures that may have approximated saturated soil conditions (Luxmoore et al. 1981, p. 689; Wilson and Luxmoore 1986, p. 578). For the purposes of this report, saturated hydraulic conductivities in the stormflow zone are assumed to be equal to infiltration rates, and these values control lateral stormflow through macropores and mesopores.

The estimated 50% porosity of the clayey stormflow zone represents total water storage capacity of this layer, but only a part of the stored water contributes to hydrologic processes. The field capacity of a clay is commonly about 40% (Bouwer 1978, pp. 260-261). Thus about 10% of layer volume is water that will drain under the influence of gravity. Drainage pathways in the stormflow zone probably are macropores and mesopores, but most water comes from smaller pores. Drainage is rapid at first and then progressively slower in the absence of precipitation and infiltration. Soil water content at the wilting point of vegetation is about 30% for a clay soil (Bouwer 1978, p. 265). Thus, an additional 10% of soil volume in the root zone is water that is available for transpiration. Soil moisture contents below 30% are possible by surface evaporation, but vegetation wilting is uncommon in the study area. Typical water content in the stormflow zone may range from 30% to 50% of soil volume; effective porosity or storativity of this zone is about 0.10.

6.2 VADOSE ZONE

The vadose zone consists of unsaturated material between the stormflow zone and the water table. As described previously, the geometric mean of infiltration rate in the vadose zone is 0.0030 m/d. For the purposes of this report, saturated hydraulic conductivity is assumed equal to infiltration rate. However, the downward percolation of water in the vadose zone is controlled by vertical permeability, which may be considerably smaller than the infiltration rate because of anisotropy.

The more than 30% clay content of regolith in the vadose zone probably means that nearly all groundwater percolation occurs through secondary openings. The nature and configuration of these openings have not been completely determined, but Luxmoore et al. (1981, pp. 687-688) mentioned the common occurrence of fractures in C-horizon regolith of the Conasauga Group. Dreier et al. (1987, pp. 52-55) measured a density of about 200 fractures/m in C-horizon saprolite of the Conasauga Group. Also, the standard deviations of hydraulic conductivity data (as shown by slopes of the probability graphs) for the vadose zone (Fig. 6) and the shallow aquifer (see Section 6.3) are nearly identical, strongly suggesting two samples from a single population. From these facts, fractures are presumed to be the only important secondary openings in regolith below the stormflow zone. In areas where the water table is below the top of bedrock, cavities can occur in the vadose zone. The spatial frequency of enlarged fractures and the water-bearing characteristics of both fractures and cavities are described later.

The effective porosity of the vadose zone is probably about the same as that of the stormflow zone. Drainage of the vadose zone under the influence of gravity is slower than in the stormflow zone because of the smaller hydraulic conductivity.

6.3 SHALLOW AQUIFER

Water-bearing fractures are ubiquitous below the water table, but enlarged fractures and cavities are common only at shallow depths. Enlarged openings generally are the targets for wells and constitute the water-producing intervals in wells. As described later, the geometric mean of hydraulic conductivity for the water-producing intervals is about 100 times larger than that of intervals with smaller fractures. The shallow aquifer thus can be described as consisting of water-producing intervals in otherwise relatively impermeable material.

The water table is the level at which water stands in shallow wells. It is presumed to be the same level as in a fracture at that point. Actually, most wells intercept a water-producing interval below the water table, and water rises in a well above the level of this interval. Groundwater in fractures near the water table apparently occurs under unconfined conditions, but water levels in deeper wells respond to earth tides and to loading forces caused by hydrofracture operations (Webster and Bradley 1987, p. 65). Also, the water levels in deeper wells are above the levels of the water-bearing fractures, and flowing wells occur in a few areas (Webster and Bradley 1987, p. 44, for example). Confined conditions at deeper levels are apparently caused by relatively impermeable material between the water-producing intervals. There is probably a gradual change from almost completely unconfined to confined conditions with increasing depth in the shallow aquifer.

Well depths in the Conasauga Group and the Chickamauga group approximately represent the depths of water-producing intervals because a large majority of the wells were drilled to the shallowest of these intervals. A cumulative probability graph of well depths (Fig. 7) shows that these data have little deviation from a straight line representing a lognormal population except at the upper end of the graph. This one deviation is not easily interpreted, but a cumulative probability graph of bedrock thickness in wells (Fig. 8) has a distinctive break in the slope of the population line at a probability of 75%. The two slopes on these graphs seem to indicate some type of control on the depths of water-producing intervals. The nature of this control is unknown, but the graphs show that very few wells intercept a water-producing interval below a depth of 30 m ($\ln 30 = 3.4$). This depth is assumed to represent the approximate base of the shallow aquifer in the area underlaid by the Conasauga Group and the Chickamauga group. Enlarged fractures and a few cavities occur at depths up to 60 m in the Knox Group, and the base of the shallow aquifer in these areas has not been accurately determined.

The geometric mean of well depth (Fig. 7) and thus the mean depth of the first water-producing interval in the Conasauga Group and

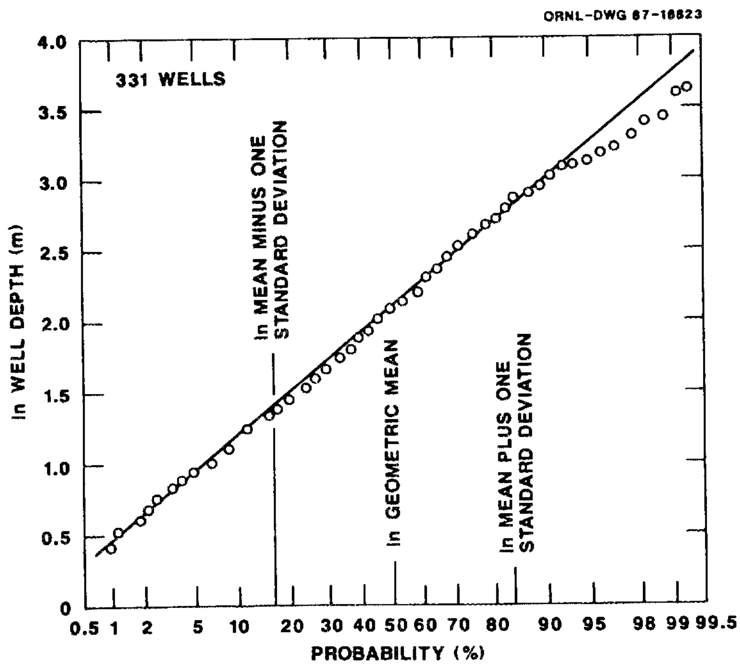

Fig. 7. Cumulative probability graph of well depths.

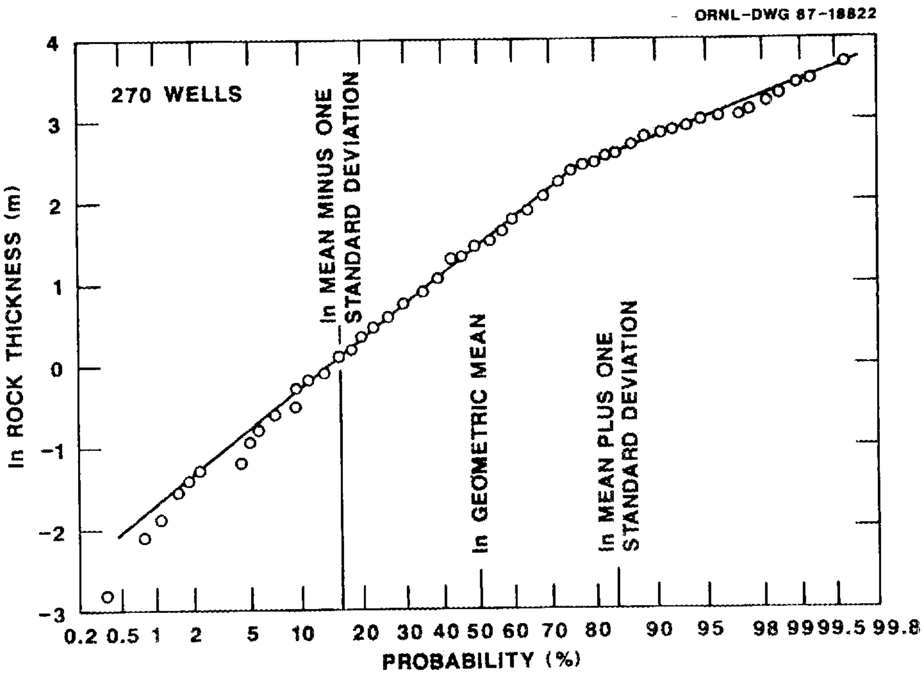

Fig. 8. Cumulative probability graph of bedrock thickness in wells.

the Chickamauga group is 8.2 m; the range from the mean minus one to the mean plus one standard deviation is 4.1-16.8 m. The probability graph of rock thickness (Fig. 8) shows that the geometric mean thickness of rock above the shallowest water-producing interval is 4.3 m. Similar information is not presently available for the Rome Formation or the Knox Group.

Hydraulic conductivity values (Fig. 9) obtained from slug tests on wells can be interpreted to obtain information on the lateral spatial frequency of water-producing intervals in the shallow aquifer. Out of 413 test results, 6-33 of the smallest hydraulic conductivity values can be fitted to a cumulative probability graph for a population that represents unenlarged fractures, such as the intervals that are not water producing in the shallow aquifer. These small values are anomalous among data that otherwise represent the permeabilities of enlarged fractures and are interpreted to mean that about 1-8% of all wells do not intercept enlarged fractures in the shallow aquifer. The lateral spatial frequency of water-producing intervals is thus about 0.92-0.99.

The vertical spatial frequency of water-producing intervals in the shallow aquifer can be calculated from the average spacing and thickness of these intervals. Some data on vertical spacing are available for 29 well pairs in the Conasauga Group and the Chickamauga group. The deep well of each pair was drilled to the first water-producing interval below the level of the shallow well, and the difference in well depth represents the spacing from the bottom of one water-producing interval to the bottom of the next lower interval. The geometric mean difference in well depth is 10.4 m, and the range from the mean minus one to the mean plus one standard deviation is 8.2-12.6 m. The average thickness of the water-producing intervals can be calculated by dividing the geometric mean of transmissivity (0.16 m^2/d) for these intervals by the geometric mean of hydraulic conductivity (0.041 m/d). The average thickness of water-producing intervals thus is 3.9 m, and the average vertical spatial frequency is 0.38. If these data are representative of the area underlaid by the Conasauga Group and the Chickamauga group, there are an average of three

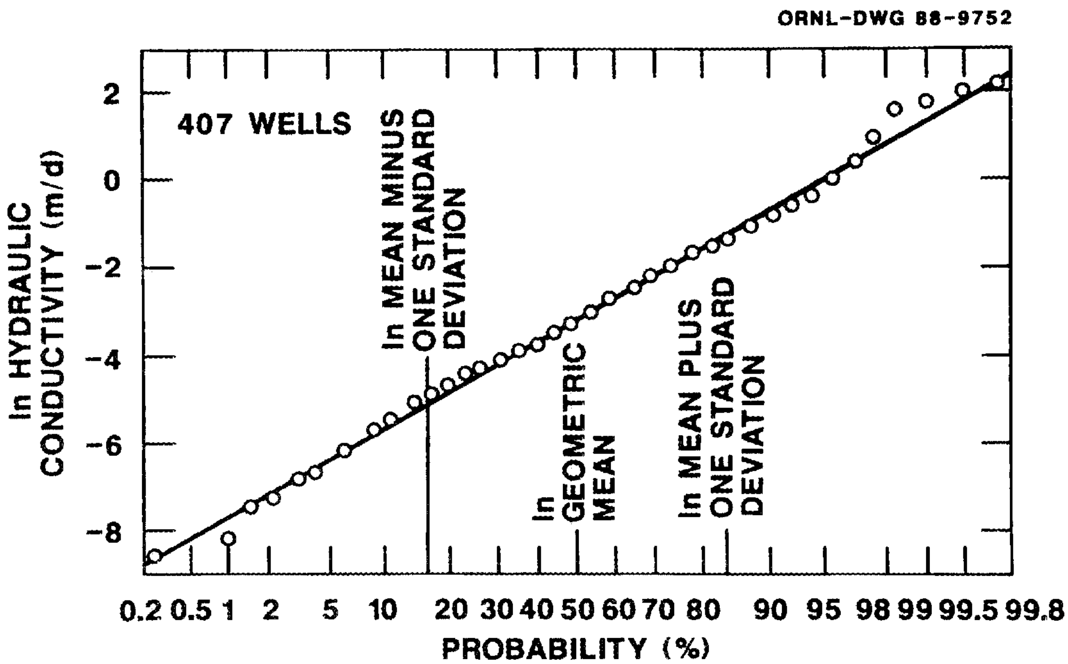

Fig. 9. Cumulative probability graph of hydraulic conductivity for water-producing intervals in the shallow aquifer.

water-producing intervals between land surface and the base of the shallow aquifer. Data are not available on the vertical spacing of water-producing intervals in the Rome Formation and the Knox Group.

Hydraulic conductivity data from water-producing intervals in the shallow aquifer (Fig. 9) are well fitted to a straight line except for a few points at the upper end of the graph. This fit shows that the data are described by a single lognormal population. The geometric mean of the population is 0.041 m/d, and the range from the mean minus one to the mean plus one standard deviation is 0.0058-0.29 m/d. A possible problem with this graph is that a large majority of the wells were not developed before the slug tests were run. Also, twelve wells have hydraulic conductivity values that plot above the line at the upper end of the graph. The possible significance of this deviation is discussed later.

The geometric mean of hydraulic conductivity for 50 water quality monitoring wells, which were thoroughly developed before slug tests were run, is 0.080 m/d, nearly twice the geometric mean value of the wells in Fig. 9. About half of the developed wells are in the Conasauga Group (WAG 6), and the rest are in the Chickamauga Group (WAG 1). Development removes mud and other detritus from the fractures, thereby increasing aperture and hydraulic conductivity near the well, but slug test results are not necessarily representative of the original, natural permeability. Also, test results may have been affected by the fact that screens or open-hole intervals commonly are longer in the water quality monitoring wells and that a better selection of water-producing intervals may have been made. Nevertheless, these results show that hydraulic conductivity values in the study area may have an accuracy of only about 50%.

Approximately the same number of wells are screened in regolith and in bedrock (Table 5), and there are only small and statistically insignificant differences in the geometric means of hydraulic conductivity for water-producing zones in these types of aquifer material. The similar hydraulic conductivities strongly suggest similar modes of groundwater occurrence in regolith and bedrock. Most

Table 5. Relationship of hydraulic conductivity ty type of aquifer material

Aquifer type	Number of values	Hydraulic conductivity (m/d) Geometric mean	Mean minus one standard deviation	Mean plus one standard deviation	Minimum value	Maximum value
Regolith[a]	182	0.044	0.0064	0.31	0.00019	48
Bedrock	225	0.042	0.0061	0.29	0.00020	7.6
Population	407	0.041	0.0058	0.29	0.00019	48

[a]Includes 77 wells screened across regolith-bedrock contact.

groundwater flow in the regolith may occur through fractures that extend upward from bedrock.

A comparison of hydraulic conductivity values by geologic unit (Table 6) shows some differences, most of which are unimportant. Results from the relatively small number of tests in the Rome Formation are probably not representative of this unit. One-tail Student's *t*-tests of log-transformed (normalized) data show that the geometric means of hydraulic conductivity for the Conasauga Group, Knox Group, and the Chickamauga group are not statistically different from the geometric mean of the population at the 1% level of significance. Similar tests show that both the Conasauga Group and the Knox Group have statistically larger geometric means of hydraulic conductivity than the Chickamauga Group. One of these tests is not conclusive, however, because 10% of the hydraulic conductivity values for wells in the Chickamauga Group are in the upper 10% of all values, but only 5% of the wells in the Conasauga Group have hydraulic conductivity values in this range. The differences between the Conasauga Group and the Chickamauga group in Table 6 apparently result from data distribution and are not real.

A comparison of hydraulic conductivity values between bedrock lithologies (Table 7) in the shallow aquifer shows only small differences except for dolostone, which occurs in the Knox Group and in a few layers of the Rome Formation. A one-tail Student's *t*-test of the log-transformed data shows that the geometric mean of hydraulic conductivity for dolostone is statistically larger than the geometric means of the population and the other rock types at the 1% level of significance. Similar tests show that the geometric means of hydraulic conductivity for the other lithologies are not statistically different than the geometric mean of the population. A different comparison shows that 9.4% of the wells in limestone have hydraulic conductivity values in the upper 10% of all values but that only an average 3.5% of hydraulic conductivity values are in this range for wells in shale, mixed shale and limestone, and mixed limestone and shale. Thus the small differences between the hydraulic conductivities of these units apparently result from differences in data distribution.

Table 6. Relationship of hydraulic conductivity to geologic formation or group

Geologic unit	Number of values	Hydraulic conductivity (m/d) Geometric mean	Mean minus one standard deviation	Mean plus one standard deviation	Minimum value	Maximum value
Rome	7	0.13	NA[a]	NA[a]	0.0027	7.3
Conasauga	241	0.050	0.010	0.25	0.00020	2.1
Knox	36	0.078	0.0045	1.3	0.00028	48
Chickamauga	123	0.029	0.0037	0.23	0.00019	7.6
Population	407	0.041	0.0058	0.29	0.00019	48

[a]Not applicable because of the small number of wells.

Table 7. Statistical characteristics of hydraulic conductivity values for bedrock lithology of the water-producing intervals

Lithology	Number of values	Hydraulic conductivity (m/d): Geometric mean	Mean minus one standard deviation	Mean plus one standard deviation	Minimum value	Maximum value
Dolostone	19	0.15	0.019	1.1	0.0019	6.1
Limestone	85	0.026	0.0028	0.25	0.00019	7.6
Limestone and shale	71	0.050	0.011	0.23	0.00065	2.6
Shale and limestone	31	0.037	0.0058	0.24	0.00022	0.59
Shale	40	0.031	0.0070	0.14	0.0010	0.41
Population	302	0.040	0.0063	0.26	0.00019	7.6

A classification of hydraulic conductivity values by well depth (Table 8) shows no statistically significant differences in the geometric means except at depths of 20-30 m. However, the geometric mean of hydraulic conductivity for wells deeper than 30 m is biased because 17 (63%) of the wells are in areas underlaid by the Knox Group. The geometric mean of hydraulic conductivity for deep wells in the Conasauga Group and the Chickamauga group is only 0.0053 m/d.
A general decrease in the hydraulic conductivity of water-producing intervals at deeper levels is also shown by 26 well pairs in the Conasauga Group and the Chickamauga group. The geometric mean depth of the shallow wells in the pairs is 7.5 m and that of the deeper wells is 19 m. The geometric mean of hydraulic conductivity for the shallow wells is 0.062 m/d whereas the geometric mean for the deeper wells is 0.016 m/d. In four cases, on the other hand, the hydraulic conductivity for the deeper well in a pair is larger than that of the shallow well.

Some of the water-producing intervals in the shallow aquifer are cavities. Slug tests were run on 25 cavity wells; the geometric mean of hydraulic conductivity for these wells is 0.085 m/d, which is about twice as large as the geometric mean for all tested wells. However, a one-tail Student's t-test of the log-transformed data shows that the geometric mean of hydraulic conductivity for cavities is not statistically larger than that of the population at the 1% level of significance. Five cavity wells are among the group of 12 hydraulic conductivity values that plot above the fitted line at the upper end of the probability graph (Fig. 9). This deviation shows that a few cavities and a few enlarged fractures have hydraulic conductivity values somewhat larger than would be expected in the population. However, these 12 values constitute only about 3% of the population, and they are not greatly larger than would be expected, as shown by the small deviation from the line.

Cavities presumably represent locations where there are large flows of groundwater or where large flows have occurred in the past. Unless hydrologic conditions or flow paths have changed, it is not logical that cavities have about the same average hydraulic

Table 8. Statistical characteristics of hydraulic conductivity values for selected classes of well depth

Well Depths (m)	Number of values	Hydraulic conductivity (m/d)				
		Geo-metric mean	Mean minus one standard deviation	Mean plus one standard deviation	Minimum value	Maximum value
0-5	74	0.044	0.0070	0.28	0.00019	1.3
5-10	166	0.054	0.011	0.26	0.00060	8.9
10-15	59	0.052	0.0069	0.39	0.00020	7.6
15-20	44	0.039	0.0040	0.38	0.00056	48
20-30	37	0.014	0.0017	0.12	0.00022	2.9
> 30	27	0.036	0.0029	0.45	0.00067	6.1
Population	407	0.041	0.0058	0.29	0.00019	48

conductivity as other water-producing intervals in the shallow aquifer, as is the case in the study area. It is possible that the large majority of cavities were formed in the geologic past, at the same time as the terrace gravels. If so, the cavities may be remnants of fossil flow paths. This hypothesis will require further study but is supported by the common occurrence of gravel (among other materials) as a cavity fill. A few cavities and enlarged fractures may also have larger hydraulic conductivities than those that have been measured in slug tests. This would seem to be the case for water-producing intervals near large springs, especially in the Knox Group.

Other intervals in the shallow aquifer contain fractures that have a different distribution of hydraulic conductivity values. Twenty-six (79%) of the 33 lowest date values in Fig. 9 are from wells less than 30 m deep. These low values are anomalous because they can be successfully merged with results of packer tests in coreholes up to 400 m deep and with results of slug tests on hydrostatic head-monitoring wells, which have depths of about 30, 60, and 120 m. A cumulative probability graph of the merged data (Fig. 10) is somewhat irregular, but a single straight line can be satisfactorily fitted to the points. The statistical characteristics of this population apparently describe hydraulic conductivities both in the deeper aquifer and in the parts of the shallow aquifer that do not represent the water-producing intervals. The geometric mean of hydraulic conductivity in this population is 0.00044 m/d, and the range from the mean minus one to the mean plus one standard deviation is 1.8×10^{-5} to 0.011 m/d.

The ranges of hydraulic conductivity values in Figs. 9 and 10 have a large overlap, and only the highest and lowest values can be automatically assigned to one population or the other. This overlap suggests that fracture apertures are a continuum in the study area. Hydraulic conductivity data are separated into two populations because (1) the resulting distributions have different standard deviations, as shown by the different slopes of the population lines; (2) the larger hydraulic conductivity values occur only at shallow depths; and (3) the geometric mean of hydraulic conductivity for water-producing intervals

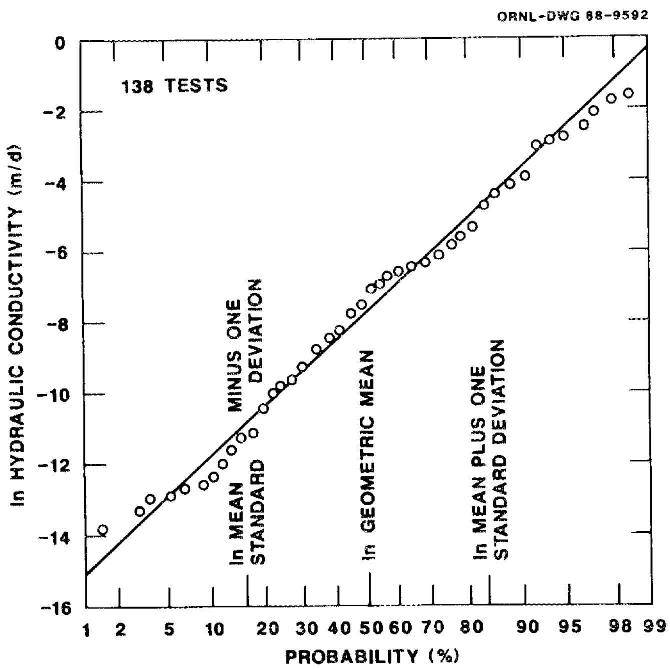

Fig. 10. Cumulative probability graph of hydraulic conductivity for intervals that are not water producing.

is about 100 times larger than that of the other intervals. Nevertheless, the data overlap shows that terms such as "enlarged fracture" and "water-producing" interval are relative and do not necessarily reflect a large difference in local permeabilities within the aquifer.

Recent studies of fractured-rock aquifers have shown that groundwater flow commonly occurs along tubes or channels and that the sides of a fracture are otherwise tightly compressed (Witherspoon et al. 1980, for example); the older concept envisioned flow between parallel plates. Studies of this type have not been made in the ORNL area. However, some driller's logs refer to "1/8th-inch streams" or "1/4th-inch streams" of water flowing into wells during construction; these descriptions seem to refer to tube flow. If tube flow is a factor in the study area, water-producing intervals include multiple tubes because the average thickness of these zones is 3.9 m, as was calculated above.

In fractured rock, storativity is commonly assumed to be equal to effective porosity (volume of fractures that drain by gravity in a unit volume of rock). In the ORNL area, storativity probably has a lognormal distribution and a large range, but various previous studies have estimated the mean value of this parameter. Webster and Bradley (1987, Table 6) estimated storativity of the aquifers at $1x10^{-4}$ to $1x10^{-5}$ using aquifer test data for several wells. Recent slug tests on 450 piezometer wells suggest that the range in storativity is $1x10^{-2}$ to $1x10^{-6}$ but that the mean may be about $1x10^{-4}$. More accurate measurements were made by Smith and Vaughan (1985, pp. 141, 144), who used two aquifer tests and the data from six observation wells in each test; they obtained geometric mean values for aquifer storativity of $1x10^{-3}$ and $4x10^{-3}$. It is probably significant that the average storativity of 0.25% determined by Smith and Vaughan (1985) is approximately the same as the 0.21% volume of macropores and mesopores calculated from soil infiltration tests (Watson and Luxmoore 1986, p. 581). Thus, mean effective porosity for the shallow aquifer may be 0.0025. In this case, each 10 m of aquifer thickness has a water storage capacity of 2.5 cm.

6.4 DEEPER AQUIFER

The deeper aquifer occurs below any water-producing intervals and generally has the same characteristics as intervals that are not water producing within the shallow aquifer. The geometric mean of hydraulic conductivity is 0.00044 m/d, and the range from the mean minus one to the mean plus one standard deviation is 1.8×10^{-5} to 0.011 m/d (Fig. 10). It is important that this range covers about 2.6 orders of magnitude, and, as discussed above, that there is a large overlap with the range of hydraulic conductivity values for water-producing intervals in the shallow aquifer. In the deeper aquifer, intervals having relatively large hydraulic conductivity values could be called water producing, in comparison with intervals having relatively small values.

All water in the deeper aquifer occurs under confined conditions. This water comes from shallower levels and eventually returns to shallow levels before discharge to springs and streams. However, the smaller geometric mean of hydraulic conductivity in the deeper aquifer, as compared to water-producing intervals in the shallow aquifer, indicates that rates and quantities of groundwater flow are much smaller than in the shallow aquifer.

Storativity has not been measured in the deeper aquifer but probably is smaller than in water-producing intervals of the shallow aquifer because mean aperture is smaller. A typical effective porosity and storativity in the deeper aquifer might be in the range 1×10^{-5} to 5×10^{-4}.

7. RECHARGE AND WATER LEVELS

Precipitation and infiltration first replenish any soil moisture deficit within the root zone of vegetation. After field capacity has been reached, continued infiltration produces a saturated layer and a perched water table near the base of the stormflow zone. At this time, both lateral groundwater flow through the stormflow zone and vertical percolation through the vadose zone begin. A majority of the water that enters the stormflow zone during the growing season is consumed by evapotranspiration, but virtually all deeper percolation reaches the water table and recharges the aquifers.

The lateral downslope movement of water in the stormflow zone means that water content within this zone decreases to field capacity sooner on the ridges than in the valleys. Near major streams, the water table is in the stormflow zone; at valley edges the zones are separate but aquifer recharge continues longer than on hills and ridges. Wells on floodplains in the study area have a shallow depth to water and generally have a small seasonal fluctuation in water level.

A majority of all aquifer recharge occurs during the nongrowing season and soon thereafter, from about November 5 to April 30. During periods of intense precipitation in the growing season, some recharge reaches the water table, and water levels rise in wells or show a slower rate of decline for a few days. However, the water levels in all wells decline, although at a variable rate, throughout the growing season because most precipitation is captured by vegetation in this period of time.

Annual low water levels in wells are reached in the fall, commonly in October and almost always between September and early December. Annual high water levels follow periods of prolonged or intense precipitation and occur sometime between December and early June. Also, however, some wells reach an annual high water level 6 weeks or more before other wells. In the near normal period of early 1987, for example, observation wells in the ORNL main plant area (WAG 1) recorded high water levels in the middle of March whereas most wells in Melton Valley (WAGs 5, 6, and 7) continued rising slowly until May.

A cumulative probability graph of depth to water in October (Fig. 11) for wells in the Conasauga Group and the Chickamauga group shows that two or three lines with different slopes are required to fit the data points but that one line is a good fit to 82% of the data. For the main group of values, the geometric mean depth to water is 4.1 m, and the range from the mean minus one to the mean plus one standard deviation is 1.7-10.0 m. Data points for about 2% of the wells fall below the fitted line at the lower end of the graph. This deviation apparently indicates that depths to water of less than 0.75 m are somewhat less than would be expected in a lognormal population; this deviation can be ignored for most purposes. Data points for about 15% of the wells plot with a shallower slope at the upper end of the graph. Nearly all of these wells are deeper than the geometric mean of well depths, and about 80% of the wells are deeper than the mean plus one standard deviation. These deeper wells are fairly good evidence that the change in slope on the graph results from a change in conditions at deeper levels in the aquifer. This change may be related to the maximum difference in elevation and to the hydraulic gradients between ridges and valleys.

A cumulative probability graph of seasonal change in water level (Fig. 12) for wells in the Conasauga Group and the Chickamauga group shows that the data are well fitted to a straight line except at the low end of the graph. The geometric mean of change in water level is 1.48 m and the range from the mean minus one to the mean plus one standard deviation is 0.74-2.94 m. Some but not all of the wells that plot below the fitted line at the lower end of the graph are close to streams and ponds. Unusually small water-level fluctuations in the other wells cannot be explained. Only about 3% of all wells have anomalous fluctuations, however, and these data can probably be ignored for most purposes.

If seasonal changes in the water table were to occur within a material with an effective porosity of 0.10, such as the stormflow zone, the average seasonal change in ground-water storage would be

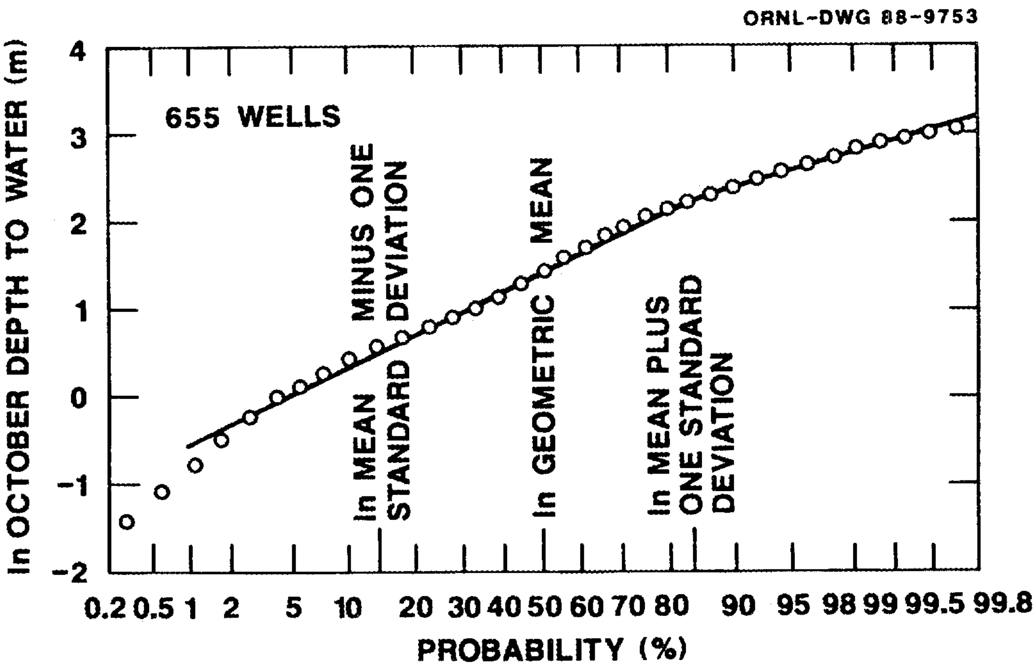

Fig. 11. Cumulative probability graph of depth to water in October.

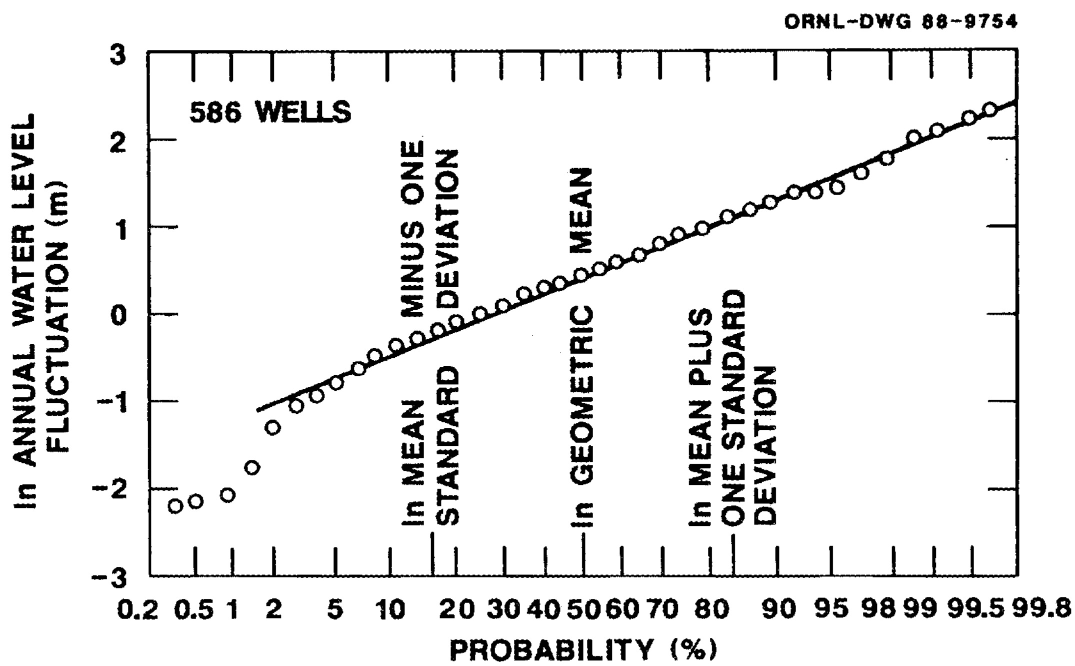

Fig. 12. Cumulative probability graph of seasonal fluctuation in water levels.

about 15 cm of water. This amount of water is unreasonable because it is 27% of all streamflow, and hydraulic conductivities in the shallow aquifer are much smaller than those in the stormflow zone. The effective porosity of fractured bedrock is about 0.0025, and an average water level change of 1.48 m represents a change in water storage of only 0.37 cm of water. Combined with other evidence, this is fairly good proof that the effective porosity of the shallow aquifer near the water table is that of a fractured rock and not that of granular material.

8. FLOW PATHS, HYDRAULIC GRADIENTS, AND FLOW RATES

Nearly all groundwater flow paths (Fig. 13) are short. Most stormflow travels only about 50-100 m from the points of infiltration to wet-weather seeps and springs. The remainder of the stormflow is discharged at downslope springs and streams, distances that generally do not exceed 200 m. Flow paths for some of the groundwater that reaches the shallow and deeper aquifers are longer but probably do not exceed 300-400 m. Exceptions occur, especially along valley axes. Dye was detected in one tracer test (Ketelle and Huff 1984, p. 133) across a travel distance of 1.5-1.8 km.

Water from precipitation and infiltration accumulates in the stormflow zone because, as discussed previously, average hydraulic conductivity is more than 1000 times larger than in the underlying vadose zone. This accumulation forms a transient, perched water table, which slowly declines as water is drained by lateral and vertical flows. A large majority of the perched water moves laterally downslope in the stormflow zone. The remainder percolates downward through the vadose zone to the water table.

Lateral hydraulic gradients in the stormflow zone cannot exceed surface slopes, which range from near 0 in flat areas to about 0.4 on the steepest hillside. Surface slopes in the upper half of this range are uncommon. Also, lateral hydraulic gradient near a stream can only slightly exceed cross-valley slope, which averages about 0.016 in floodplains and other areas of low relief. A larger gradient on an upslope hillside represents a larger groundwater flow rate and causes the stormflow zone to fill and overflow, as occurs at wet-weather springs. For this reason, nearly all wet-weather springs occur on concave slopes. The average of lateral hydraulic gradients is probably about 0.05-0.1.

Representative velocity and flow rate in the stormflow zone can be calculated from typical parameter values, by using simple equations derived from Darcy's law:

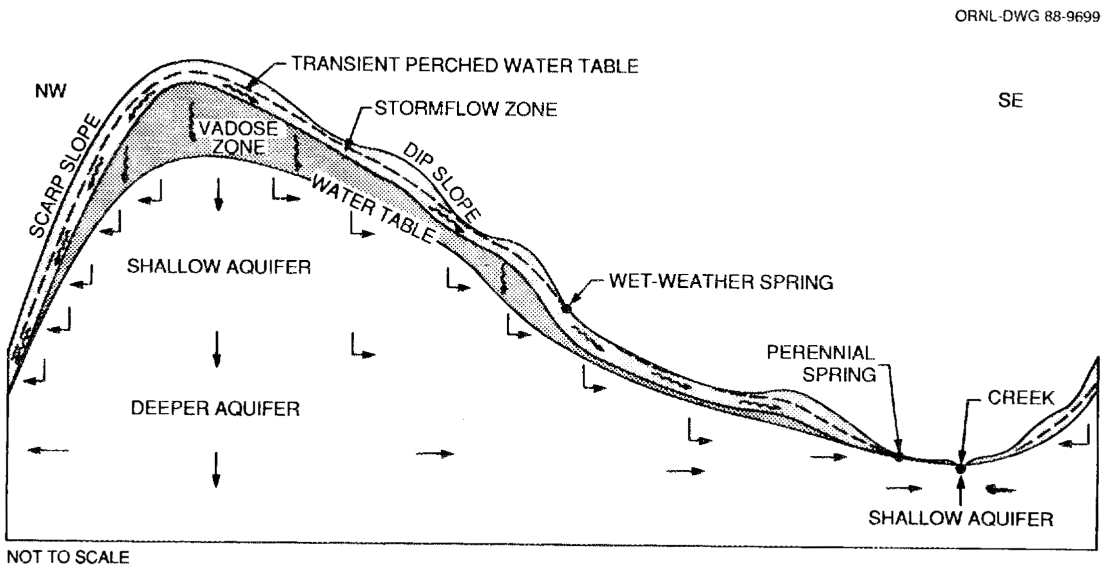

Fig. 13. Section showing subsurface zones and directions of groundwater flow.

$$v = -Ki/n ,$$

and

$$Q = -KiA ,$$

where v is average linear velocity, Q is flow rate, K is hydraulic conductivity, i is hydraulic gradient, n is effective porosity, and A is cross-sectional area. If mean hydraulic conductivity is 4.0 m/d, if average hydraulic gradient is 0.075 and if effective porosity is 0.10, then average linear velocity is 3.0 m/d. If saturated thickness is 1.0 m, then groundwater flow rate through a section 305 m wide is 64 L/min, and discharge into a stream from both sides of the channel would be 128 L/min for a reach 305 m long.

The capacity of the stormflow zone to transmit infiltration can be checked by assuming a flow tube 1 m wide and 200 m long (length of the longest flow path). About 56 cm of precipitation are not consumed by evapotranspiration. Therefore, the flow tube transmits about 110 m^3/year of water. If hydraulic conductivity is 4.0 m/d and saturated thickness at the lower end of the flow tube is 1.0 m, a hydraulic gradient of 0.077 results in annual discharge of the same total volume of water. In the vicinity of wet-weather springs, there is convergence of flow tubes because drainage volumes resemble semicylindrical or lemniscate sections. This convergence suggests that the area of the spring grows as discharge increases and contracts as discharge decreases.

Most groundwater movement in the vadose zone consists of near-vertical percolation down to the water table. If hydraulic conductivity is 0.0030 m/d, if hydraulic gradient is 1.0, and if effective porosity is 0.0025, then average linear flow rate for this percolation is 1.2 m/d. If the average thickness of the vadose zone is 3 m, then an average 2.5 d is required for water to percolate from the stormflow zone to the water table.

Rapid lateral groundwater movement apparently occurs in a few areas where cavities are at or above the water table. One tracer test in the Knox Group (Ketelle and Huff 1984, pp. 131-135) showed a water velocity of about 200-300 m/d between a swallow hole and a discharge point farther downstream. Another tracer test in the Chickamauga Group showed a groundwater velocity of about 20-80 m/d between an excavated cavity in limestone and a sump in a reactor building at ORNL main plant. Parts of the flow paths for both tracer-test results are above the water table. It is hypothesized that lateral flow velocities larger than 1-3 m/d are rare in the ORNL area and are more likely to occur above the water table, mainly during storm events, than below the water table.

Flow paths in the shallow aquifer generally trend down the slope of the land surface but have numerous deflections and are characterized by splits and joins along multiple routes. Linear flows apparently occur only in a few areas, where small folds or faults control flow directions in water-producing intervals of the shallow aquifer. Webster (1976, p. 8) summarized previous studies showing that some structures of this type may persist along strike for as much as 300 m and might be longer.

In other areas of the shallow aquifer, flow-path characteristics are determined by intersecting fracture networks. Within a fracture, groundwater may flow downdip and laterally in either or both of two directions. Changes in flow direction can occur at fracture intersections. A simple representation of flow paths is a stairstep pattern in both plan and section views. However, simple flow paths are uncommon because splits and joins also occur at fracture intersections. Tracer tests in three different areas (Webster and Bradley 1987, p. 32; Davis et al. 1984, p. 77) eventually detected small concentrations of the tracer in all observation wells, including wells along strike but in opposite directions from the injection well. Thus, a contaminant entering the aquifer at one point may eventually occur in all fractures within a semicylindrical volume of the aquifer.

Apparent hydraulic gradients in the shallow aquifer, as calculated from water-level elevations in observation wells, range up to 0.3-0.4 on steep hillsides. However, calculations based on shallow wells in areas with low relief show only a median cross-valley gradient of 0.016 and a median along-valley gradient of 0.0070. These smaller hydraulic gradients apparently are representative of groundwater flow rates toward discharge in the streams, although water levels are within the stormflow zone in some of the areas. The large gradients on hillsides probably result from changes in potential (hydraulic head) as a result of the near-vertical movement of groundwater from one level to another. A decrease in hydraulic gradient may be common near the base of a ridge, and this change may explain the presence of some springs. Gradients that control the lateral rate of groundwater flow in the shallow aquifer probably do not exceed 0.05 and might be as small as 0.01 in areas of low relief.

Hydraulic heads at deeper levels in the shallow aquifer commonly are different than those near the water table. At any point below the water table, hydraulic potential is a vector sum of head losses along multiple flow paths, including paths leading to and from this point. Between any two points, including points at shallow and deeper levels, differences in hydraulic potential represent head losses along multiple flow paths that connect the points; these flow paths are more likely to be tortuous than linear.

The distribution of hydraulic conductivity values with depth can be interpreted to show that approximately the same amounts of groundwater move vertically and laterally within the upper 20-30 m of the aquifer. The average hydraulic conductivity for water-producing intervals is about 100 times larger than that between these intervals. However, the cross-sectional area for vertical flow between water-producing intervals is larger than that for lateral flows within the intervals. Thus, for example, if the water producing intervals average 3.9 m thick and if two lateral flow tubes are 1 m wide, then the combined area for lateral flow is 7.8 m^2. If the average length of the flow tubes is 250 m, the cross-sectional area for vertical flow between the tubes is 250 m^2. These relationships mean that the head loss along a vertical flow path may be two to four times larger than for the same distance along a lateral path within a water-producing interval. Webster and Bradley (1987, pp. 61-66) describe water level differences that show an average vertical hydraulic gradient (for a linear flow path) of about 0.1 for several well clusters in WAG 5. This vertical gradient is two to ten times larger than the lateral gradients discussed above.

Water-level differences of up to 5 m between shallow and deeper wells in a pair have been commonly interpreted as showing relative amounts of groundwater flow between aquifer levels (Webster and Bradley 1987, pp. 62-69, for example). Instead, differences in water level show only relatively impermeable (or long and tortuous) flow paths and the resulting differences in hydraulic potential between the levels. In the case of a permeable connection between water-producing intervals in a well pair, water levels would rise and fall together and would maintain a nearly constant head difference. The well pairs in the study area have water-level changes that do not fit this description.

Previous workers agree that the shallow aquifer is anisotropic and that fractures with traces parallel to the strike of the rock layers are more permeable than transverse fractures. Tucci (1985, p. 5) stated "the reported ratio of strike-normal to strike-parallel hydraulic conductivity values ranges from 1:3 to 1:20." All evidence for anisotropy is from breakthrough times in tracer tests and from drawdown patterns in observation wells during aquifer tests. Within aquifer volumes this large, the results might show only that strike-parallel fractures are more numerous or longer than transverse fractures. Nevertheless, the effect is the same: a larger hydraulic gradient is necessary for a given rate of groundwater flow in the cross-valley direction than in the along-valley direction. Tucci (1985, pp. 10, 15) determined that an anisotropic ratio value in the range 1.5-3.0 was appropriate for a digital groundwater flow model. In comparison, the calculated ratio of along-valley and cross-valley gradients (0.007 and 0.016) near streams in Melton and Bethel valleys is 1:2.3.

Groundwater in the shallow aquifer flows into the stormflow zone near major streams (Fig. 14), and the two zones may also be adjacent in other areas, such as steep hillsides. Regolith is thin on steep hillsides of the Conasauga Group and the Chickamauga group, and any vadose zone must also be thin.

Typical velocities and flow rates in the shallow aquifer can be calculated from average values of the parameters. If hydraulic conductivity is 0.041 m/d in the water-producing intervals and

0.00044 m/d in the other intervals, if effective porosity is 0.0025, and if cross-valley hydraulic gradient is 0.050, then average linear velocity is 0.8 m/d in the water-producing intervals and 0.01 m/d in the other intervals. If the combined thickness of water-producing intervals is 12 m and the combined thickness of the other intervals is 18 m, total groundwater flow toward the stream (from both valley sides) is about 11 L/min through sections 305 m wide. All but about 1.5% of this flow is through the water-producing intervals.

The amount of flow in the shallow aquifer can be checked by considering a flow tube 1 m wide and 250 m long (the assumed average length of flow paths). Discharge from the flow tube would be 9.1 m^3/year. This amount of water is equivalent to an annual recharge of 3.6 cm of water along the length of the tube. The calculated flow rate is reasonable because a similar recharge rate for the entire shallow aquifer represents about 6.5% of the streamflow.

The recharge rate can be used to estimate vertical hydraulic conductivity in the shallow aquifer. If vertical hydraulic gradient is 0.1 and if all recharge water were to move vertically downward (none were to flow laterally), then the average vertical hydraulic conductivity would be 0.00099 m/d. This value is unreasonable because it is about 2.2 times larger than the geometric mean of hydraulic conductivity for intervals that are not water producing in the shallow aquifer. The calculated value for vertical hydraulic conductivity would be smaller if some groundwater moves laterally, and it would be larger if flow paths are nonlinear (and the hydraulic gradient is smaller). The results of the calculation thus suggest that the shallow aquifer loses water by lateral flow above the levels of springs and streams. If so, the most likely explanation is that fractures transport water (above the water table) from the shallow aquifer into the stormflow zone.

The occurrence of a thin vadose zone above the shallow aquifer shows that average annual recharge is almost exactly balanced by discharge; this fact requires some consideration and explanation. If the shallow aquifer were capable of transmitting more water to the streams than is received as recharge, the water table would be only slightly above the level of the streams. The water table is 30-50 m below land surface in a few areas on Chestnut Ridge, and these areas may represent a situation in which potential discharge exceeds recharge. Conversely, if the shallow aquifer were to receive more recharge than could be transmitted vertically and laterally, the water table would be in or near the stormflow zone. Locally, very shallow water levels in a few observation wells above the valley floors may reflect this situation. However, a thin vadose zone occurs in a large majority of the study area. It is hypothesized that fractures are kept open only by flows of groundwater and otherwise are sealed by mineral deposits. Over a long period of time, most fracture apertures may become adjusted to the available annual flux.

The rate of recharge to the shallow aquifer can be used to estimate the amount of time that recharge occurs. As calculated previously, discharge from a flow tube 1 m wide would be 9.1 m^3/year. If the flow tube is 250 m long, if vertical hydraulic conductivity of the vadose zone is 0.0030 m/d, and if vertical hydraulic gradient is 1.0, recharge of the same amount of water would occur in 12 d. However, an average recharge period of only 1 d/month is unreasonable. If the vertical hydraulic conductivity of the vadose zone were 0.00030 m/d (0.1 times as large as lateral hydraulic conductivity), recharge would occur an average 10 d/month. The actual periods for recharge of the shallow aquifer cannot be determined, but the calculated values suggest that the vadose zone is anisotropic.

Groundwater flow paths in the deeper aquifer have more and longer vertical segments than those in the shallow aquifer, and the hydraulic gradient for lateral flow is assumed to be smaller. Water from the deeper aquifer flows upward into the shallow aquifer and then into the stormflow zone near major streams. Hydraulic conductivities are small in the deeper aquifer, but this zone is much thicker than the shallow aquifer. If the deeper aquifer is assumed to average 200 m thick, if hydraulic conductivity is 0.00044 m/d, and if hydraulic gradient is 0.025, then the groundwater flow rate toward a stream through two sections 305 m wide is 0.93 L/min. This flow rate is only 9% of the rate in the shallow aquifer.

The accuracy of all calculations used to estimate groundwater flow rates depends on the accuracy of the parameters. Nearly all hydraulic conductivity values are the geometric means of more than 100 measurements and can be presumed to be accurate. Saturated thicknesses are reasonable and should be accurate within 25%. The parameter with the largest potential error in all determinations of flow rate is hydraulic gradient. It is simple to measure and compare water-level elevations in wells, but the significance of these data is poorly understood, as is indicated by the previous discussions. Flow path lengths are also important because a larger number of shorter flow tubes can discharge the same total volume of water with a smaller hydraulic gradient.

The combined total flow rate of groundwater in the shallow and the deeper aquifers is calculated to be about 9% as large as that in the stormflow zone. This result assumes that hydraulic gradient in the shallow aquifer (0.050) is somewhat larger than that measured across valley areas (0.016) but smaller than the apparent gradient for steep hillsides (0.3-0.4). If a cross-valley gradient of 0.016 were used in the flow calculation, combined groundwater flow from below the water table would be only 1.3% of stormflow, and this result is unreasonable. A hydraulic gradient smaller than about 0.03 may occur only in floodplain areas. Tentative conclusions are that 90-95% of all groundwater flow occurs in the stormflow zone and that 90% of the remaining flow occurs in the shallow aquifer.

9. CHEMICAL WATER QUALITY

The composition of groundwater is controlled by many factors, including the chemical content of recharge waters, interactions with regolith and bedrock, residence times, and mixtures or dilutions with waters from other flow paths. These factors and the complexity of flow paths below the water table commonly result in large differences in the concentrations of dissolved constituents in nearby wells. Also, the concentration of the constituents in water at some wells is not constant but varies throughout the year.

Only a few incomplete chemical analyses are available for water from the stormflow zone. Davis et al. (1984, p. 46) showed analyses of chemical extracts from soils in WAG 6 and described the material as highly leached and strongly acidic; except for one sample, the range in pH was 4.0-4.7. A water sample from well T6-7 in WAG 7 was reported to contain 3.4 mg/L of Ca, 2.7 mg/L of Na, and 1.2 mg/L of Mg. Another shallow well (1062) in lower WAG 2 has shown a range of in situ measurements for specific conductance of 35-56 micromhos/cm at 25°C over a period of 6 months. Samples of surface water collected at a flume in WAG 6 (Davis et al. 1984, p. 157) may include some water from the shallow aquifer but show a mean specific conductance of 120 micromhos/cm, a mean alkalinity as $CaCO_3$ of 120 mg/L, a mean Ca concentration of 46 mg/L, and a mean SO_4 concentration of 13 mg/L. According to these data, water in the stormflow zone is acidic to nearly neutral; the water apparently is a calcium bicarbonate type, but other important ions are Mg, Na, and SO_4. Total dissolved solids are probably less than 100 mg/L.

Ample data are available to characterize the important properties and major ions in water from the shallow aquifer in the Conasauga Group and the Chickamauga group; similar data are unavailable for the Rome Formation and the Knox Group. Cumulative probability graphs (Figs. 14-21) and a statistical summary (Table 9) show that water from the shallow aquifer is typically a nearly neutral to moderately alkaline calcium bicarbonate type. Caution should be used in interpretation of the lowest and the highest values in Table 9. These

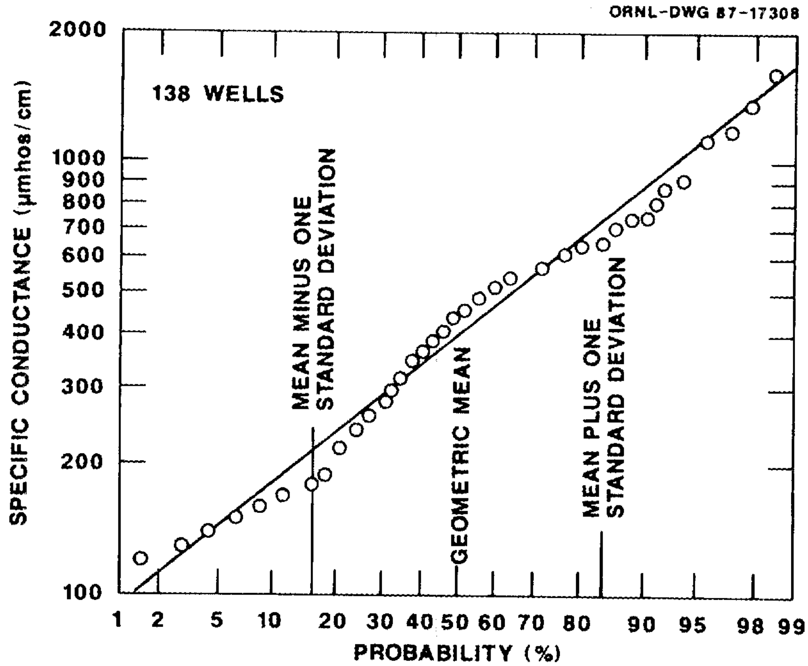

Fig. 14. Cumulative probability graph of specific conductance of water in the shallow aquifer.

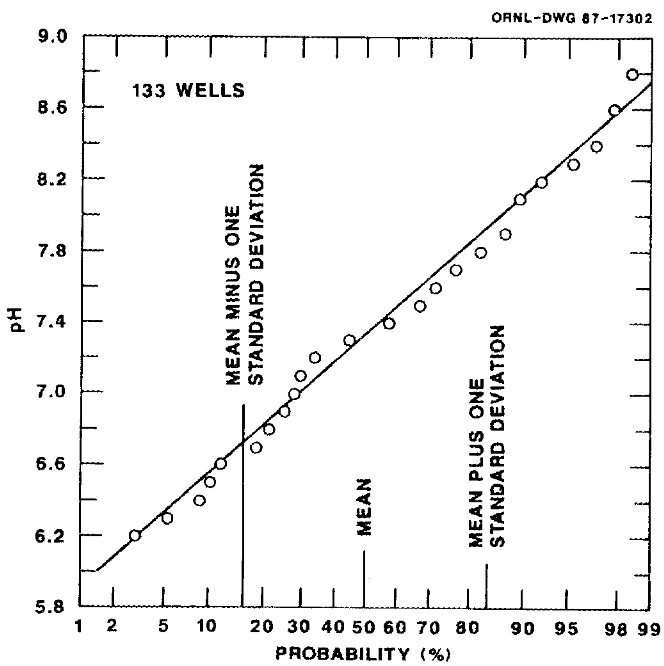

Fig. 15. Cumulative probability graph of the pH of water in the shallow aquifer.

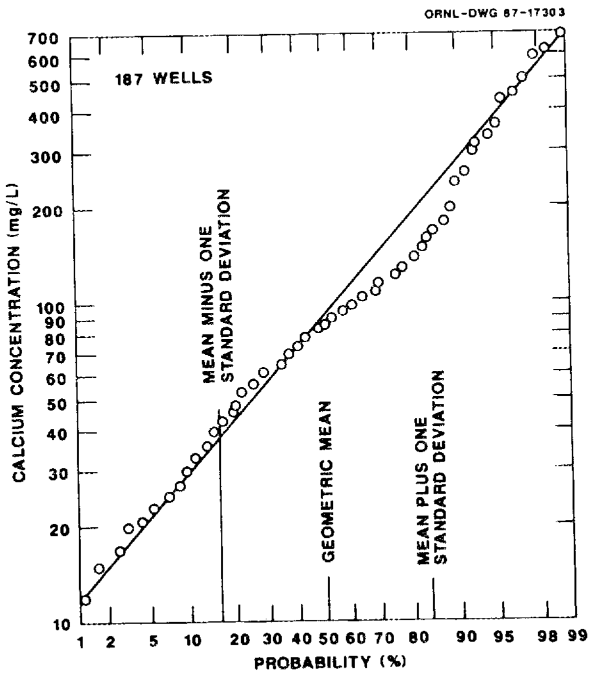

Fig. 16. Cumulative probability graph of calcium concentration in water from the shallow aquifer.

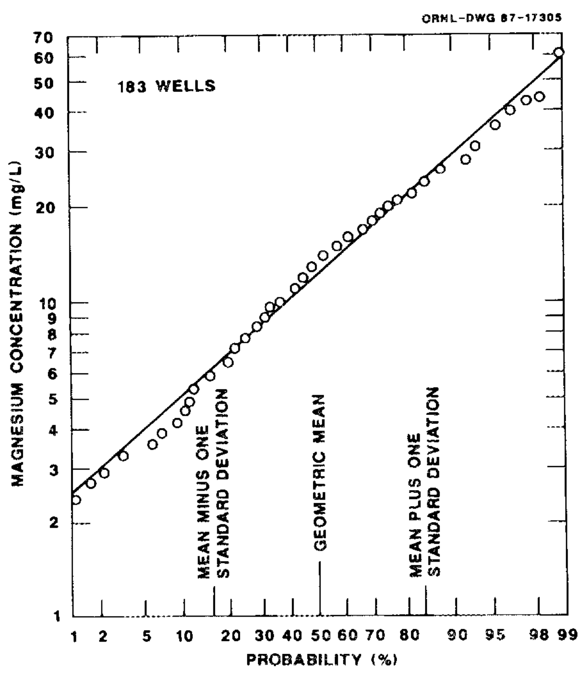

Fig. 17. Cumulative probability graph of magnesium concentration in water from the shallow aquifer.

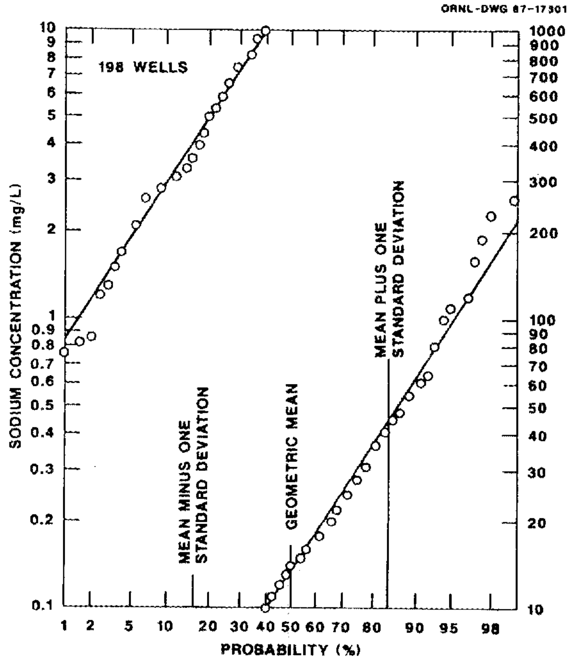

Fig. 18. Cumulative probability graph of sodium concentration in water from the shallow aquifer.

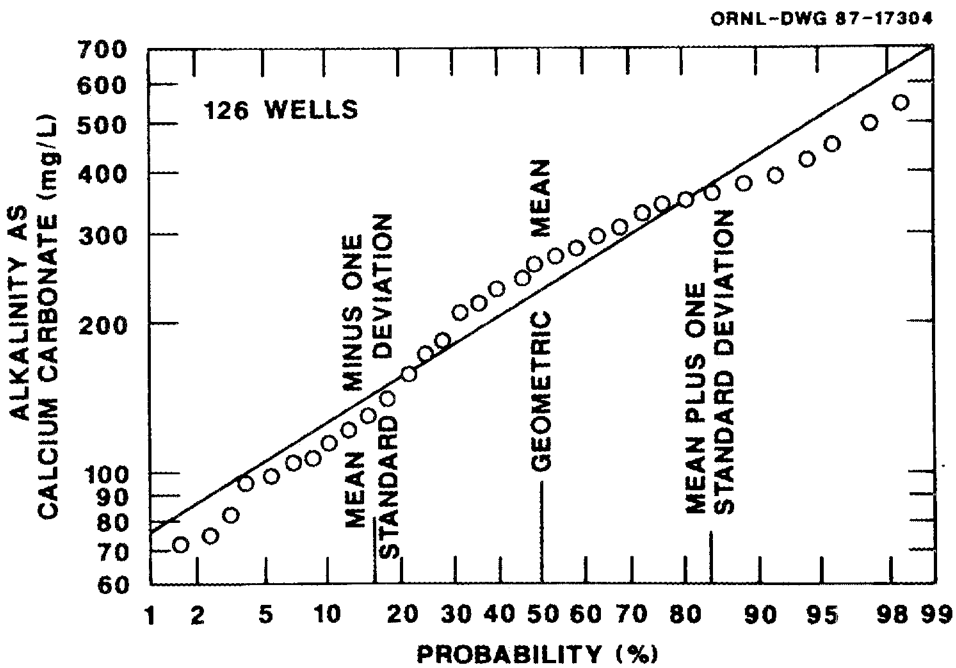

Fig. 19. Cumulative probability graph of alkalinity concentration in water from the shallow aquifer.

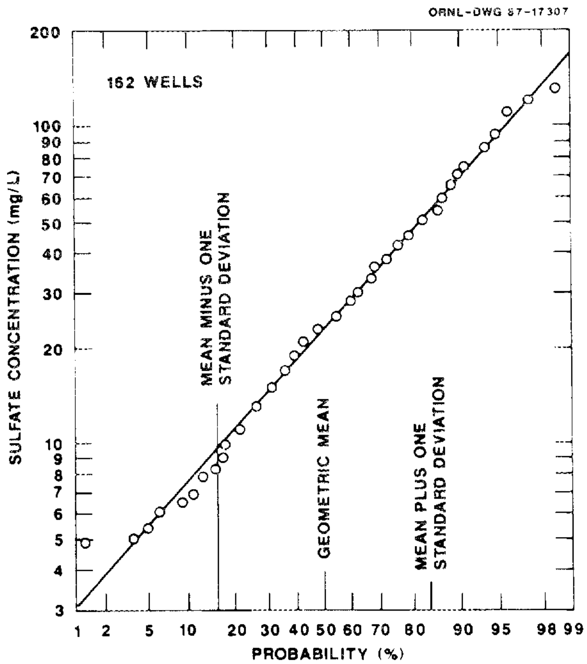

Fig. 20. Cumulative probability graph of sulfate concentration in water from the shallow aquifer.

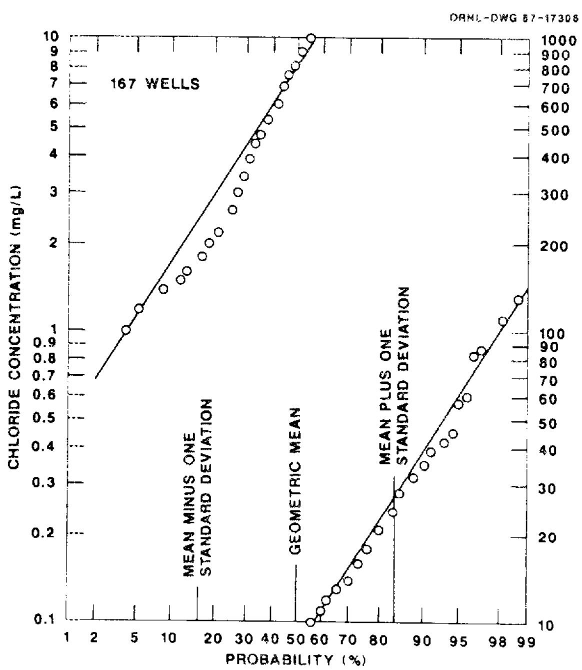

Fig. 21. Cumulative probability graph of chloride concentration in water from the shallow aquifer.

Table 9. Statistical summary of important constituents and properties in water from the shallow aquifer (specific conductance in micromhos/cm at 25°C, alkalinity as mg/L of $CaCO_3$, and other units, except pH, in mg/L)[a].

Constituent or Property	Number of values	Mean[b]	Mean minus one standard deviation	Mean plus one standard deviation	Minimum value	Maximum value
Specific conductance	138	400	220	740	110	1930
pH	133	7.3	6.7	7.9	6.2	9.2
Calcium	187	91	38	220	12	1600
Magnesium	183	12	6.2	24	2.1	137
Sodium	198	14	4.1	45	0.7	420
Alkalinity	126	230	140	370	55	847
Sulfate	162	22	9.5	54	4.0	376
Chloride	167	7.9	2.4	27	0	200

[a]Specific conductance in micromhos/cm at 25°C, alkalinity as mg/L of $CaCO_3$, and other units, except pH, in mg/L.

[b]Arithmetic mean for pH; geometric mean for others.

values may be representative of water from the stormflow zone, the deeper aquifer, or mixtures of these waters with those in the shallow aquifer.

Available data are inadequate to show the distribution of total dissolved solids in water from wells in the shallow aquifer. However, there is an excellent correlation of total dissolved solids (TDS) with specific conductance (SC) within the range 190-930 micromhos/cm at 25°C:

$$(TDS) = (SC)/1.62 - 42.7.$$

The correlation coefficient is 0.995. Based on this relationship, the geometric mean of total dissolved solids in the shallow aquifer is 204 mg/L, and the range from the mean minus one to the mean plus one standard deviation is 93-414 mg/L.

The geometric mean of calcium content in water from wells in the Chickamauga Group is 110 mg/L whereas that from wells in the Conasauga Group is 64 mg/L. Similar differences probably also occur for other constituents and properties among lithologies and the named geologic units; comparisons of this type have not been made. Also, available data are inadequate to characterize the minor elements and ions in the water. All constituents, including contaminants, are expected to be lognormally distributed.

The properties and concentrations of chemical constituents in water from the shallow aquifer are not constant. Davis et al. (1984, pp. 95-97, 159-179) showed relatively small differences between wells 8.8-15 m deep in a small area of WAG 6 but relatively large changes through time (1980-82). As described below, large differences may also occur between relatively nearby wells.

An experiment to monitor the temperature and specific conductance of water in the shallow aquifer was begun in March 1988. Monthly measurements are made in situ by lowering a probe to the midpoint of each well screen. The measurements are not the same as would have been obtained after well pumping and purging, but they have been shown to be repeatable and are thought to be representative of groundwater near the wells. Preliminary results show that some wells have very little change in temperature and specific conductance of water over a period of at least several months, whereas the water in other wells may change by several degrees centigrade or by more than 50% in specific conductance from one month to the next. Examples of the specific conductance records are shown in the table below. All wells are in WAG 6. Well 652 is 28 m deep, well 640 is 16 m deep, well 645 is 8.1 m deep, and well 644 is 5.2 m deep.

	Specific conductance of water in micromhos/cm at 25°C			
Date	Well 652	Well 640	Well 645	Well 644
3-3-88	300	1004	625	380
4-8-88	299	1108	674	321
5-5-88	299	1145	613	363
6-9-88	283	1145	675	332
6-29-88	301	1074	743	203
7-28-88	300	1041	745	156
8-31-88	300	1038	756	349

Well 652 has a fluctuation of only 6% in specific conductance of water over about 6 mos, but the other fluctuations are 12% in well 640, 19% in well 645, and 59% in well 644. These and other records show that shallow wells generally have the most fluctuation in both water temperature and specific conductance and that deeper wells have the least.

An example of specific conductance measurements for two lines of paired shallow and deeper wells on June 1, 1988, is shown on Fig. 22. Locations of wells and section lines are indicated on the map. The sections show location and length of each well screen and the specific conductance (micromhos/cm at 25°C) of water at the center of each screen. Groundwater in this area probably flows laterally across the valleys and upward to discharge into White Oak Creek. However, there is no apparent relationship between well screen locations and amounts of specific conductance. This result apparently indicates nearly unique conditions along each of many aquifer flow paths and suggests that the interpretation of other data obtained by water quality monitoring will be complex.

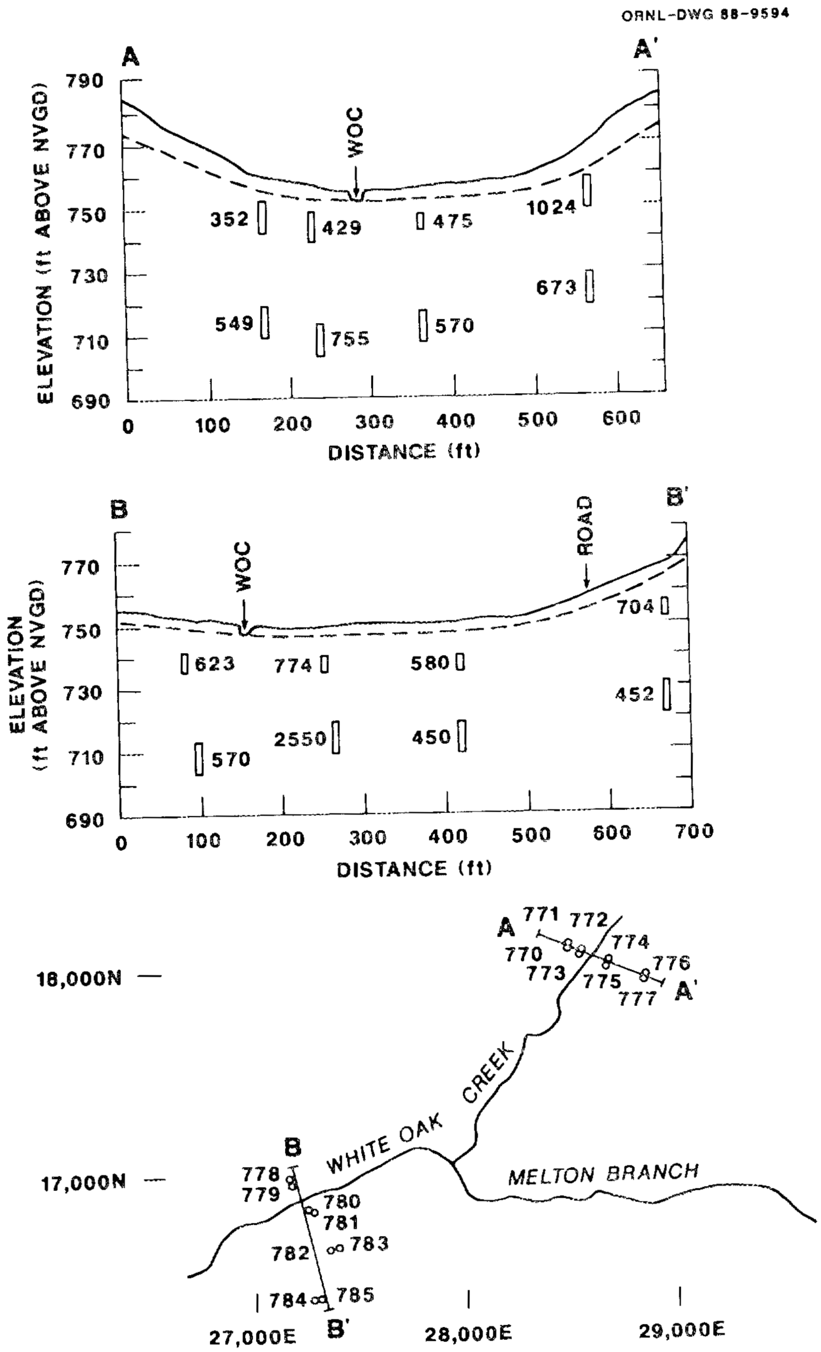

Fig. 22. Location map and sections showing specific conductance of water along lines of paired shallow and deeper wells.

A distinctly different groundwater is reported by Webster and Bradley (1987, Table 10) from six wells that are in the Conasauga Group of Melton Valley and are 30-61 m deep. This is a sodium carbonate or sodium bicarbonate type with a pH of 8.5-10.5, a sodium content of 60-300 mg/L, and a calcium plus magnesium concentration of less than 12 mg/L. These characteristics apparently result from ion exchange (calcium and magnesium for sodium) along deeper flow paths. Total dissolved solids are less than 500 mg/L, according to these few data. Other water types may also occur in the deeper aquifer, but available data are inadequate for characterization.

A much higher concentration of dissolved salts is found in water from wells that are in Melton Valley and are 150-450 m deep (C. S. Haase and John Switek, ORNL, written communication, 1987); total dissolved solids contents as high as 300,000 mg/L have been reported. This water is acidic and has (1) a high percentage weight of chloride; (2) an equivalent weight of sodium less than that of chloride; (3) enriched calcium, magnesium, strontium, and bromide contents; and (3) relatively low concentrations of bicarbonate, sulfate, and nitrogen. In terms of membrane-filtration theory, these waters are membrane concentrated and connate.

10. CONTAMINANT TRANSPORT

Subsurface contamination may occur by spills and leaching at land surface; by leakage from pipes, drains, buried tanks, and sumps; and by the leaching of buried materials, including wastes. Contaminants may thus be introduced into the stormflow zone, the vadose zone, or the shallow aquifer. Soluble and liquid constituents (and probably colloidal particles) are entrained in groundwater in the source areas and generally move toward points of discharge at springs and streams. Contaminant transport occurs along flow paths within each zone of groundwater flow and along paths that connect the zones. The hydrologic characteristics of all materials below the surface thus are important in an evaluation of alternatives for remedial action.

All precipitation infiltrates the land surface in nearly all of the study area. The majority of this water replaces the soil moisture deficit in the stormflow zone and is later consumed by evapotranspiration. However, prolonged or intense precipitation forms a perched water table near the bottom of the stormflow zone. Groundwater then moves laterally through the stormflow zone at a velocity of about 3 m/d and percolates through the vadose zone to the shallow aquifer. The perched water table and the resulting groundwater flow are transient beneath hills and ridges but may be nearly perennial at valley edges. About 90-95% of all groundwater flow is in the stormflow zone.

Other characteristics of the stormflow zone are important for contaminant transport. First, this zone is shallow and thin; lateral groundwater flow can be intercepted by drainage lines and trenches. Second, groundwater storage is apparently intergranular, and more filtration and sorption of contaminants may occur in the stormflow zone than at deeper levels. Third, many flow paths are relatively short and converge; any pollution plume would tend to be localized. Finally, it is important that infiltration and recharge occur along the entire lengths of the flow paths in the stormflow zone and that water percolates down to the water table whenever and wherever there is a perched water table in the stormflow zone.

In the vadose zone and in the shallow and deeper aquifers, groundwater occurrence is only in fractures and a few cavities. There is much less oxidation, filtration, and sorption of any contaminants than would be assumed by only a consideration of the thicknesses of these zones. Also, the average linear velocity for vertical percolation of water through the vadose zone is about 1.2 m/d. If a contaminant were introduced into the stormflow zone and if the water content of that zone exceeded field capacity, the contaminant would reach the water table in an average of about 60 h. This downward percolation can be stopped only by removing the source or by dewatering the stormflow zone.

The possibility of a rapid lateral transport of contaminants in cavities above or at the water table is also important. As described previously, cavities of this type have been documented but apparently are uncommon. In the Conasauga Group and the Chickamauga group, cavities of this type occur only in limestones and apparently occur only in areas where the water table is below the top of bedrock. Even in areas that meet this description, most cavities do not transmit unusually large amounts or velocities of groundwater.

The water table of the shallow aquifer is within the stormflow zone near major streams, and water flows from the shallow and deeper aquifers into the stormflow zone in these areas. The vadose zone may also be missing on steep hillsides. There is less filtration and oxidation of any organic contaminants in these areas. Otherwise, the principles of contaminant transport are similar to those in the separate zones.

Lateral groundwater flows in the shallow aquifer are relatively slow, an average 1 m/d, and only 5-10% of all streamflow moves through this zone to discharge locations. However, the complex flow paths mean that any contaminant introduced at a point would soon occupy all fractures within a semicylindrical volume of the aquifer and that pollution in upland areas would continue to spread both radially and vertically. In valleys, on the other hand, the convergence of lateral and vertical flow paths toward discharge locations would tend to localize a pollution plume.

An important consideration for remedial action is that the shallow aquifer yields only small amounts of water to wells. Any polluted aquifer volume can be easily dewatered, and only small water volumes would be produced for treatment. Average well yield is about 1.1 L/min of water with a drawdown of about 6.1 m. Under steady-state conditions and with an annual recharge of 3.6 cm of water, the area of influence for a pumping well would extend a radial distance of 70 m, and the hydraulic gradient toward the pumping well would be 0.085.

Only about 1% of all groundwater follows flow paths through the deeper aquifer. This water comes from the shallow aquifer and eventually flows upward, back into the shallow aquifer. Fractures in the deeper aquifer are tightly compressed; filtration and sorption may be somewhat more effective than in the enlarged fractures and cavities of the shallow aquifer. Webster and Bradley (1987, pp. 34, 59, 75-78) detected little radionuclide contamination below a depth of 30 m in Melton Valley. Well yields are very low in the deeper aquifer, and large aquifer volumes could be dewatered at an average pumping rate of 0.2 L/min for each well.

Radionuclides, like other elements, cannot be made to disappear; they can only be moved, diluted, concentrated, or changed in physical or chemical form. If radionuclides are present below land surface, the only effective method of preventing the slow spread of contaminated groundwater is isolation. Isolation can be achieved by installing physical barriers to groundwater flow, by making the wastes insoluble or impermeable, or by ensuring that the surrounding materials are dry. Monolithic barriers may or may not be a long-term solution because the same forces that have fractured the rocks (probably earth tides) may fracture the barriers. Also, some low level wastes cannot readily be made impermeable or insoluble. Fortunately, the subsurface hydrology of the ORNL area is favorable for the draining and drying of materials surrounding the wastes. The principles are relatively simple: the blocking of both infiltration and lateral flow in the stormflow zone. Once recharge of the shallow aquifer ceases, the water table will decline (probably several meters) to a steady state hydraulic gradient. If additional drainage is required, arrays of pumping wells will be needed, but only small quantities of water will be produced.

11. CONCLUSIONS

Groundwater occurrence and flow are in three or four zones; each zone has distinctive characteristics that affect groundwater velocity, flow rate, contaminant transport, and remedial action. The stormflow zone is 1-2 m thick and is just beneath land surface. About 90-95% of all groundwater flows through this zone to discharge at springs and streams. The vadose zone generally separates the stormflow zone from the underlying shallow aquifer. However, the vadose zone is absent near major streams and may be missing on steep hillsides. Also, effective porosity and the standard deviation of hydraulic conductivity in the vadose zone are the same as those of the shallow aquifer. The vadose zone thus may constitute an unsaturated upper part of the shallow aquifer. Groundwater flow in the vadose zone generally consists of near-vertical percolation to the water table. In a few areas, however, lateral flows of water occur through cavities in limestone. About 5-10% of all groundwater flows through the shallow aquifer to discharge locations. This zone extends to a depth of 20-60 m. Nearly all flow in the shallow aquifer moves through a few water-producing intervals that consist of enlarged fractures or cavities. About 1% of all groundwater follows flow paths through the deeper aquifer. This zone extends to the base of fresh water at a depth of 150 m or more. Water in the deeper aquifer comes from the shallow aquifer and it eventually flows upward, back into the shallow aquifer. Water-producing intervals in the deeper aquifer are much less permeable than in the shallow aquifer.

All precipitation infiltrates the land surface in nearly all of the study area. Exceptions are overland flow in areas with urban features, a few forested and seasonal wetlands, a few water bodies, and a few small areas of barren land. Infiltration first replenishes any soil moisture deficit in the stormflow zone and then accumulates and forms a perched water table. Both lateral groundwater flow in the stormflow zone and downward percolation (resulting in recharge of the shallow aquifer) occur whenever and wherever there is a perched water table. These flows are transient on hills and ridges but may be nearly perennial at valley edges. The stormflow zone also constitutes the reservoir for nearly all evapotranspiration; the majority of precipitation and infiltration are consumed by these processes. Discharge of water from the stormflow zone is mostly at wet-weather springs and streams and thus is transient. The remainder of the discharge is at major streams and is nearly perennial. All groundwater in the shallow and deeper aquifers flows into the stormflow zone near discharge locations. Groundwater storage is apparently intergranular in the stormflow zone, but the pathways for vertical percolation and for lateral flow are predominantly macropores and mesopores. The detailed shape and pattern of these larger openings is unknown. Lateral flow paths in the stormflow zone are assumed to generally follow the slope of land surface and to converge near discharge locations.

Beneath the water table, groundwater in excess of field capacity occurs only in fractures and cavities. Water in a fracture may flow downdip and laterally in either or both of two directions. Changes in flow direction, splits, and joins occur at fracture intersections. Groundwater flows in the fracture networks thus are characterized by complex multiple flow paths. At any point in the aquifer, hydraulic potential is a vector sum of head losses along the multiple flow paths leading to and from this point. Between any two points (including points at shallow and deeper levels), head losses occur along all flow paths that connect the points, and these paths are more likely to be tortuous than linear. These factors complicate the determination of vertical and lateral hydraulic gradients. In general, flow paths trend toward lower elevations, are longer at deeper levels, and converge near discharge locations. Increases and decreases in streamflow are accompanied by changes in the length of flowing channels. Numerous wet-weather springs appear after periods of precipitation and disappear after a few days to weeks of dry weather. Nearly all of the increases and decreases in groundwater discharge are from the stormflow zone. Seasonal water-level changes in wells represent only small changes in groundwater storage in the shallow aquifer and even smaller changes in hydraulic gradient. During long dry periods, groundwater discharge ceases because the water is captured by vegetation (near discharge locations) before reaching the streams.

Groundwater to depths of at least 20-30 m is acidic to moderately alkaline and a calcium bicarbonate type. Other important constituents are magnesium, sodium, sulfate, and chloride. Total dissolved solids generally are less than 500 mg/L. However, the concentrations of the chemical constituents are not constant through time and are not spatially correlated among nearby wells. Relatively large changes in the specific conductance of water in some shallow wells occur over a period of 1-2 months, for example. The interpretation of data obtained from water quality monitoring wells will be complex. An alkaline, sodium carbonate water type has been identified in a few wells at depths of 30-61 m in Melton Valley, and a brine occurs at depths below about 150 m in this area. Elsewhere, the base of fresh water has not been determined.

Most problems of contaminant transport by groundwater in the ORNL area have been caused by relatively large amounts of infiltration and lateral flow in the stormflow zone and by the percolation of water from this zone to the water table. Remedial action will require hydrologic isolation of contaminants as could be accomplished by blocking infiltration and stormflow in control areas. It may or may not be necessary to pump and treat contaminated groundwater from the shallow aquifer. If so, only relatively small rates and volumes of water will be involved.

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