Practical Handbook of

ENVIRONMENTAL SITE CHARACTERIZATION AND GROUND-WATER MONITORING

SECOND EDITION

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Monitoring and Sampling the Vadose Zone

Thomas Ballestero, Beverly Herzog, and Glenn Thompson

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accuracy of the method is not high for detecting small changes in water content, especially for dry soils. Like gamma-ray logging, this method requires care in the handling of the radioactive source.

Other Methods

There are a variety of other methods for obtaining information on soil moisture content. These methods range from destructive to nondestructive and noninvasive. Soil samples can be removed from the field and moisture content measured destructively. Although accurate, the major disadvantages include repeatability, the need for drilling equipment to sample at significant depths, and the long time it takes to get the actual data.

A variety of nondestructive techniques operate on the electrical or magnetic properties of the soil-water system. Each method develops moisture content from the sphere of measurement of the instrument. These methods include time domain reflectometry (propagating an electromagnetic wave between electrodes to measure the dielectric properties — see ASTM D 6565 [ASTM, 2004e]), nuclear magnetic resonance (generating magnetic fields to measure induction decays — discriminates between bound and free water in the soil), soil capacitance (measuring the capacitance between two buried electrodes), and fiber optics (measuring light attenuation from a known source).

Remote sensing techniques have the ability to rapidly cover large areas, but possess much less sensitivity. These techniques include thermal infrared imagery and radar.

Monitoring Vadose Zone Transmission Properties

Vadose zone transmission properties are generally of greater interest in ground-water monitoring studies than are storage properties. Field methods for measuring the unsaturated hydraulic conductivity, which are described in detail in ASTM Standard D 5126 (ASTM, 2004f), are complicated due to its dependence on moisture content. However, one can measure flux and use this to calculate fluid transmission rates at known levels of saturation or matric potential. In a manner similar to ground-water investigations, boundary and initial conditions are prescribed (flow rate), variables are recorded (tension, moisture content), and then, from hydraulic theory (Equation 3.16), hydraulic conductivity is estimated. The rate of moisture movement is determined indirectly from infiltration rates or measurements of unsaturated flow. When using field data to estimate transmission properties in the vadose zone, it is important to remember that large variations in these parameters can easily result from soil heterogeneities.

Field Measurements of Infiltration Rates

Field measurements of infiltration rates are appropriate for estimating downward fluid transmission during the wetting cycle. Infiltration rates are affected by soil texture and structure (including soil layers), initial moisture content, entrapped air, and water salinity. Waste disposal options for which the principal component of flux is downward include surface spreading or ponding of wastes and installation of landfill liners composed of earthen materials.

Infiltration is determined using infiltrometers; infiltrometers do not directly measure hydraulic conductivity. Infiltration is the process by which water enters a permeable material. When infiltration begins, the infiltration rate is relatively high and is dominated by matric potential gradient. As the matric potential gradient decreases, the infiltration rate asymptotically decreases with time until the gravity-induced infiltration rate, called the steady-state infiltrability, is approached (Hillel, 1980a). This relationship is shown in Figure 3.13. Steady-state infiltrability is directly proportional to saturated hydraulic conductivity and hydraulic gradient. Therefore, in order to calculate saturated hydraulic conductivity from infiltration data, the hydraulic gradient and the extent of lateral flow must be known. Gradient data can be obtained using many of the instruments described in the previous section. Saturated hydraulic conductivity is of interest even in the vadose zone because it is the upper boundary for unsaturated hydraulic conductivity; use of this value provides a conservative estimate of fluid transmission time. With very long times, the gradient approaches unity and hence, from Darcy's law, the infiltration rate approaches the value of the saturated hydraulic conductivity.

While infiltrometers can be designed with either a single ring or a double ring, the double-ring method is preferred because its design minimizes lateral flow, simplifying the calculation of saturated hydraulic conductivity. The method is described in ASTM Standard D 3385 (ASTM, 2004g). The principle of operation is based on maintaining a constant head in the inner and outer rings of the infiltrometer. Both rings are sealed in the soil to prevent leakage under the rings. Water is added to the rings to maintain the constant head; if the inner ring is covered to prevent evaporation, the volume of water added to the inner ring is equal to the water infiltrating into the soil. In the design of Daniel and Trautwein (1986) water is added to the inner ring through an intravenous (IV) bag (Figure 3.14). As water from both rings enters the soil, water exits the IV bag and moves into the inner ring to maintain a constant head. The IV bag design is well suited to soils with low infiltration rates because the small amount of added water can be measured accurately by weight. For more permeable soils, the water level can be maintained by adding measured volumes of water to the inner ring. Measurements of infiltration are taken until the system reaches steady-state infiltrability. If the test is

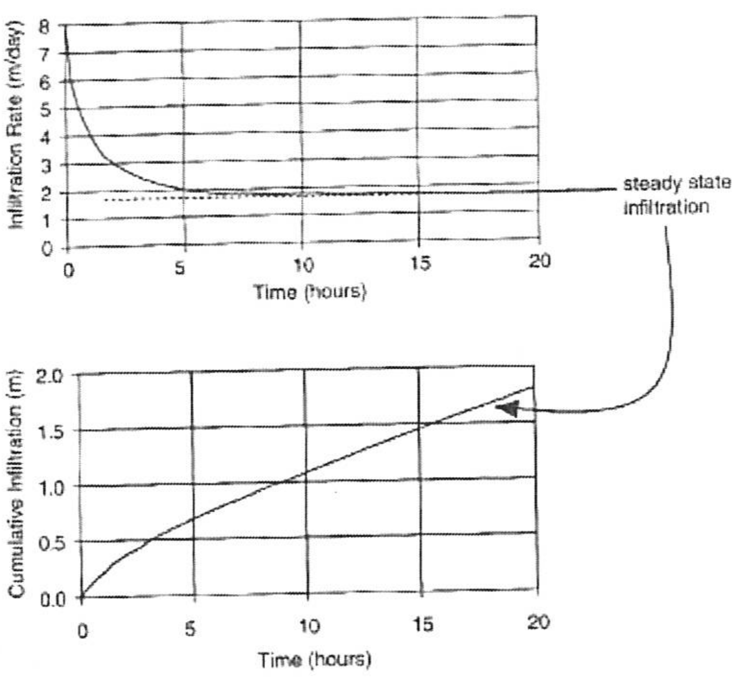

FIGURE 3.13
The relation between infiltration rate and cumulative infiltration.

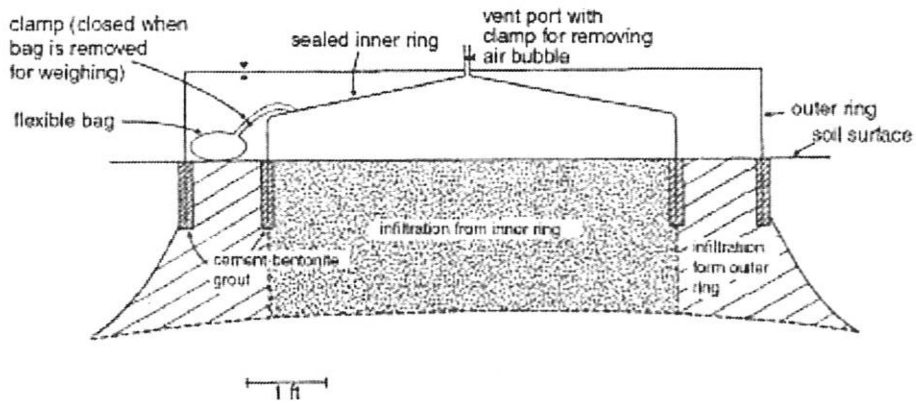

FIGURE 3.14
Schematic diagram of a sealed double-ring infiltrometer.

performed to prove that a soil meets some regulatory requirement, such as the requirement that earthen liners have a saturated hydraulic conductivity of 1×10^{-7} cm/sec or less (U.S. EPA, 1988), the test may end when this infiltration rate is achieved because the infiltration rate decreases with time and saturated hydraulic conductivity will be no more than the infiltration rate.

Infiltrometers generally range in size from less than one square foot to about 25 square feet. Large infiltrometers with IV bags were designed for soils with low infiltration rates, generally in the range of 1×10^{-5} to 1×10^{-8} cm/sec (Daniel and Trautwein, 1986). The large size is necessary to include macrostructures and to obtain measurable amounts of water loss.

Hydraulic conductivity can be calculated from infiltration rate using either Darcy's law or the Green–Ampt (1911) approximation. If Darcy's law is used, the hydraulic conductivity (K) is equal to the discharge of water flowing out of the infiltrometer (Q) divided by the product of the infiltrometer area (A) times the vertical head gradient (I): $K = Q/(AI)$. Here, Q/A is the measured, steady-state infiltration rate (infiltrability) per unit area. The head gradient, I, is measured with tensiometers at various depths. This method assumes that the flow is occurring under saturated conditions.

The Green–Ampt approximation assumes that the wetting front is sharp, the matric potential at the front is constant, and the wetted zone is uniformly wetted and of constant hydraulic conductivity. The assumption of a sharp wetting front may be reasonable for fine-grained soils, as shown by dye studies in an experimental earthen liner (Albrecht et al., 1989). The Green–Ampt approximation differs from the Darcy's law calculation in that knowledge of the depth of the wetting front is required instead of a measured hydraulic gradient. Under these assumptions, the analytical solution to vertical infiltration produces an equation that resembles the Darcy's equation:

$$K = i\left[1 + \frac{h + \psi_T}{L_f}\right]^{-1} \tag{3.18}$$

where $i\ (=Q/A)$ is the steady-state infiltration rate and the bracketed term is the hydraulic gradient. Within the bracketed term, h is the height of the water in the infiltrometer, ψ_T is the total potential at the wetting front (also known as the wetting front suction), and L_f is the depth of the wetting front below the bottom of the infiltrometer. In many instances (especially after long times), ψ_T is assumed to be 0. This assumption results in a lower estimate of K, which may or may not be conservative, depending on the nature of the objective (irrigation, waste disposal, etc.).

In a similar fashion, a disc permeameter or tension infiltrometer also measures *in situ* sorptivity (S) and hydraulic conductivity for a prescribed potential. These devices allow potential and water content to be controlled accurately over a range of negative and positive heads and therefore have the ability to conveniently measure sorptivity (S) at selected tensions. Sorptivity is a combination of hydraulic conductivity, potential, and moisture content (Tindall and Kunkel, 1999) and basically represents the proportionality constant between infiltration rate and the inverse square root of time ($I = S/2t^{1/2}$). The disc permeameter (Perroux and White, 1988) is uncomplicated and does not greatly disturb the soil surface being measured. Methods and calculations for the disc devices may be found in references such as Ankeney et al. (1988).

Determination of Water Flux Characteristics

Hydraulic conductivity, flow velocity, and flux are important transmission parameters for the vadose zone. Measurement of these parameters has become more commonplace due to the need to understand how fluids move from the land surface to the groundwater system. This information is particularly necessary for waste disposal applications. A comparison of methods available for quantifying soil-water flux is presented in ASTM Standard D6642 (ASTM, 2004h).

Theoretical Perspective

Steady-state infiltration, discussed in the last section, is an appropriate base for determining flux during the wetting cycle. During the drying cycle, three major approaches to evaluating flux are possible (Everett, 1980). These include (1) calculating flux from mathematical formulae and empirical relationships between soil suction, soil-water content, and hydraulic conductivity; (2) measuring changes in the water content of the soil profile over time; and (3) direct measurements using flow meters.

Darcy's Law

The easiest method available for calculating saturated flow from infiltrometer data is the use of Darcy's law. This method is conservative because it assumes the soil is saturated; it is appropriate for the wetting cycle when steady-state flux is determined from an infiltration test. In simple terms, solving for average linear velocity (V_x), Darcy's law can be written as $V_x = Q(n_e A)$, where Q is the discharge, n_e is the effective porosity, and A is the cross-sectional area of flow. Q/A is the measured steady-state infiltration rate per unit area; for use in Darcy's law, Q/A is negative because flow is downward.

Green–Ampt Wetting Front Model

The Green–Ampt wetting front model is used with infiltration data and assumes unsaturated conditions below a wetting front. Travel time (velocity times distance) is predicted from:

$$t = \left\{\frac{\theta_s - \theta_i}{K_{sat}}\right\}\left[L_f - (h + \psi_f)\ln\left\{1 + \frac{L_f}{h + \psi_T}\right\}\right] \tag{3.19}$$

where θ_s and θ_i are initial and saturated moisture contents, L_f is the depth to the wetting front, h is the pond depth, and ψ_T is the total moisture potential just below the wetting front.

Internal Drainage Method

The internal drainage method can be used to determine the unsaturated hydraulic conductivity in the field by monitoring the transient internal drainage of a near-surface soil profile. The method is described in detail in Hillel (1980b) and was extended to layered profiles by Hillel et al. (1972). It requires simultaneous measurement of moisture content and suction under conditions of internal drainage alone; evapotranspiration must be prevented. The method also assumes that flow is vertical and that the water table is deep enough so that it does not affect the drainage process.

Tensiometers and neutron access tubes or gypsum blocks are installed near the center of the test area. Depth intervals for the instruments should not exceed 30 cm, with a desirable total depth of up to 2 m. The test area (at least 5 m by 5 m in plan view to avoid lateral disturbances on the monitoring devices) is then ponded or irrigated until the soil profile is as wet as practical (at or near saturation), then covered with plastic to avoid future fluxes across this boundary. Simultaneous measurements of soil suction and moisture content are collected until soil suction exceeds 0.5 bar; at greater suction, the drainage process may be so slow that changes become imperceptible. The measurement period for this test can be several weeks for slowly draining soils. Data are graphed as moisture content (Figure 3.15a) and suction versus time (Figure 3.15b) for each measured depth within the soil profile. Also, the snapshots of matric suction (Figure 3.15c) and total hydraulic head (suction plus depth, Figure 3.15d) assemble the data for subsequent analysis. The plots help to visually determine possible effects of nonhomogeneities. From these measurements, instantaneous values of potential gradient and flux can be obtained, allowing the calculation of hydraulic conductivity and, hence, flow velocity.

Soil moisture flux is calculated at each time and depth from:

$$q_z = \Delta z \frac{\partial \theta}{\partial t} \tag{3.20}$$

where $\partial\theta/\partial t$ is the slope of the wetness curve at the time of interest (calculated from the data that generated Figure 3.15a) and Δz is the depth increment over which the measurements are made (each curve in Figure 3.15a). Equation 3.20 is the flux through the bottom of the uppermost increment and is due to the loss in moisture in the first zone. Flux through the bottom of each succeeding layer is obtained by summing these incremental fluxes for all layers overlying the depth of interest. The flux out of the bottom of the subsequent zone is due to the dewatering in that zone, plus the flux from above, hence the summation. Flow velocity can be calculated from Darcy's law, as discussed earlier.

Hydraulic head profiles (total potentials) are obtained using the suction (matric potential) versus time data, adding depth (gravitational potential) to suction to obtain the total hydraulic head for each time (Figure 3.15d). Hydraulic conductivity, K_z, is calculated from:

$$K_z = \frac{q_z}{(\partial H/\partial z)} \tag{3.21}$$

where $\partial H/\partial z$ is the slope of the hydraulic head versus depth curve for the time of interest. K_z is calculated for several depths and times, each of which has a corresponding moisture content. As the final step, moisture content or soil suction is plotted against hydraulic conductivity so that flux and velocity can be calculated using field data at actual monitoring points. This yields a plot similar to Figure 3.11.

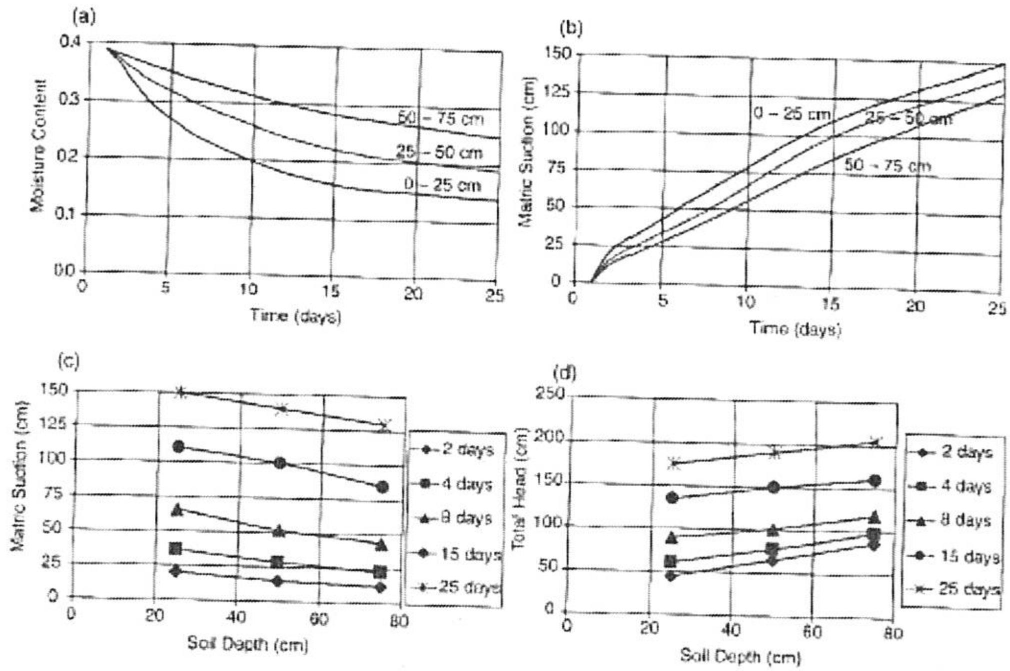

FIGURE 3.15
(a) Volumetric wetness as a function of time for different depth layers in a draining profile. (b) Matric suction variation with time for different depth layers in a draining profile. (c) Matric suction variation with time and depth during drainage. (d) Total hydraulic head variation with time and depth during drainage. Head values are suction: the higher the suction, the lower the energy.

Borehole Permeameters

Some methods for determining saturated hydraulic conductivity and sorptivity evolved from borehole methods. Here, a borehole of constant radius (r) has a constant depth of water (H) maintained in it. The result is a bulb-shaped wetting front that moves away from the borehole. Theoretical treatments (Nasberg, 1951; Glover, 1953; Reynolds et al., 1983) describe the relations between the borehole water head (H) and the saturated hydraulic conductivity for a given borehole radius. In general, for these theories to be valid, the depth to the water table must be greater than three times the water depth in the borehole. In addition, the steady-state solutions are most valid for $H/r > 10$. A good comparison of the methods appears in Stephens (1996). Of note, this method has been commercialized (i.e., the Guelph Permeameter), and methods developed to also identify the hydraulic conductivity versus pressure (suction) relationship (Reynolds and Elrick, 1986).

Measurement of Tracer Movement

Tracers are matter or energy carried by ground water, which can provide information on the rate and direction of ground-water movement. Tracers can be natural, such as heat carried by hot springs; intentionally added, such as dyes; or accidentally introduced, such as oil from an underground storage tank (Davis et al., 1985). Use of tracers in the saturated zone for determining aquifer parameters is discussed in Chapter 14, so only a brief overview is presented here. The main difference between use of tracers under unsaturated versus saturated conditions is the practical problem of sampling a tracer at increasing depths under unsaturated conditions (Everett, 1980).

Davis et al. (1985) presents a thorough discussion of tracers. The most important property of any selected tracer is that its behavior in the subsurface should be well understood. Ideally, it should move at the same rate as the ground water, should not interact with the soil matrix, and should not modify the hydraulic conductivity or other properties of the medium being monitored. Concentration of the tracer should be much greater than the background concentration of the same constituent in the natural system. The tracer should be relatively inexpensive and easily detectable with widely available technology. For most applications, the tracer should also be nontoxic.

A variety of tracers have been successfully used to monitor moisture movement in the unsaturated zone. Fluorescein and rhodamine WT dyes have been successfully used to track the wetting front beneath double-ring infiltrometers and to indicate preferential flow paths in an experimental earthen liner (Albrecht et al., 1989). Tritium from a low-level radioactive waste disposal site was successfully used to determine the rate of water (and tritium) movement in the unsaturated zone at the waste disposal site (Healy et al., 1986).

Monitoring Water Quality in the Vadose Zone

The goal of most vadose-zone monitoring programs is to measure the spatial and temporal changes in water quality. Monitoring the vadose zone can provide an early warning system to detect contaminant movement so that corrective action can begin before an underlying aquifer is contaminated. Wilson (1980) presents a thorough discussion of the chemical reactions affecting contaminant migration in the vadose zone.

Three types of methods are available for monitoring water quality in the vadose zone. These include (1) indirect methods, including measurements of electrical and thermal properties; (2) direct measurement of pore water from soil cores; and (3) direct soil-water sampling.

Electrical Properties Measurements

Electrical conductivity (EC) or its inverse, resistivity, is used extensively to characterize soil salinity and to map shallow contaminant plumes. For shallow soils, electrical conductivity is primarily a function of soil solution (Wilson, 1980). The success of using electrical properties to delineate plumes is dependent upon the contrast between the conductivity of the plume and the natural water, the depth and thickness of the plume, and lateral variations in geology.

Electrical resistivity can be measured using surface geophysical techniques, as discussed in Chapter 4, or by direct-push-deployed sensors, as described in Chapter 6. It can also be measured by using electrical resistance blocks (salinity sensors) to evaluate soil salinity. Electrical resistance blocks were discussed earlier as a means of measuring *in situ* moisture content. They can be installed beneath a waste disposal site before the site becomes operational and monitored remotely. Salinity sensors must be calibrated to provide a curve of soil salinity versus electrical conductance. Electrical conductance is highly temperature dependent, so accurate measurement of soil solution temperature is a necessary companion to this device.

Soil Sampling and Water Sampling

Pore Water Extraction

Collection of soil cores is discussed in Chapter 5 and Chapter 6. Soil cores can provide pore water for water-quality analysis. Because of the difficulty and expense of obtaining soil