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Rapid Water-Level Fluctuations in a Thin Colluvium Landslide West of Cincinnati, Ohio

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Rapid Water-Level Fluctuations in a Thin Colluvium Landslide West of Cincinnati, Ohio

By WILLIAM C. HANEBERG *and* A. ÖNDER GÖKCE

LANDSLIDES OF THE CINCINNATI, OHIO, AREA

U.S. GEOLOGICAL SURVEY BULLETIN 2059–C

The fluctuations are analyzed using analytical and numerical models of modified Dupuit hillside flow

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LIST OF SYMBOLS

A_0	Amplitude of periodic perturbation to the phreatic surface
c	Celerity of a traveling waveform (= sin β)
D	Thickness of a porous slab
H	Normalized height of phreatic surface (= h/D)
h	Height of phreatic surface measured perpendicular to the substrate
i	Spatial grid index
j	Temporal grid index
K	Hydraulic conductivity in saturated soil
L	Length of a porous slab
O	Order
p	Effective porosity
R	Net recharge to the water table
T	Normalized time (= t/t_r)
t	Time
t_r	Characteristic time (= pD/K)
X	Normalized slope-parallel coordinate (= ξ/L)
β	Slope angle in degrees
λ	Wavelength of periodic perturbation to the phreatic surface
Φ	Total hydraulic head ($h \cos \beta - \xi \sin \beta$)
σ	Aspect ratio of porous slab (= D/L)
ξ	Slope-parallel coordinate on a sloping, porous slab, giving distance from upper end of slab to point of measurement
ζ	Coordinate normal to slope on a sloping, porous slab, giving distance from base of slab to point of measurement

RAPID WATER-LEVEL FLUCTUATIONS IN A THIN COLLUVIUM LANDSLIDE WEST OF CINCINNATI, OHIO

By William C. Haneberg[1, 2] and A. Önder Gökce[1, 3]

ABSTRACT

Data from 14 rainstorms between March and May 1980 show that water levels in colluvium at the Delhi Pike landslide complex, which lies along the Ohio River valley about 15 km west of downtown Cincinnati, can rise and fall quickly during and immediately after storms. Piezometer water-level records are characteristically asymmetric, with steep rising limbs and flatter, sigmoidal falling limbs. Rainfall intensity, rainfall duration, and effective porosity all appear to control the shapes and amplitudes of these water-level records. A modified Dupuit hillside flow model shows that water levels will fluctuate uniformly along the entire slope in response to either an increase or a decrease in net recharge (due to either rainfall or evapotranspiration) and that wavelike downhill flow can occur only when net recharge is negligible. For each storm observed, we divided total rainfall by total change in ground water level to derive an estimate of pre-storm maximum effective porosity. These figures ranged from 1 to 7 percent for all but one of the storms. We found, however, that effective porosity values in the range of 1 to 3 percent gave a much better simulation of actual water-level records when used in a finite-difference approximation of modified Dupuit flow, suggesting that the hillside must have been at or very near tension saturation before the storms of record. Our model is very sensitive to a combination of effective porosity, hydraulic conductivity, rainfall intensity, and rainfall duration (all of which are difficult to characterize in the field), and suggests that caution should be exercised when attempting to predict the hydrologic response of hillsides to rainfall.

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INTRODUCTION

Several years ago, while examining dozens of landslides along the Ohio River and its tributaries near Cincinnati, Ohio, we began to realize that in some landslides destabilizing pore water pressure is artesian, whereas in others it is hydrostatic. We selected two study areas, one in glacial deposits and another in colluvium, for detailed observation and began to monitor movement and water levels in both places.

We began exploring the colluvium landslide complex in March 1979 by drilling two types of boreholes: on gentle slopes, we used a trailer-mounted auger rig to make deep boreholes; and on steeper slopes, we used a two-person portable auger to make shallow boreholes. We installed porous-stone piezometers in all holes and measured piezometric levels by hand at 2-week intervals. After realizing that piezometric levels in the shallow colluvium could change very rapidly in direct response to rainfall (Fleming and others, 1981; Gökce, 1989), we installed instruments to record both rainfall and piezometric level in one piezometer at 15-minute intervals from late March through mid-May 1980. In this paper we describe details of these rainfall and piezometric-level records and then use a simple mathematical model of hillside ground-water flow to explain some important characteristics of piezometer water-level records.

ACKNOWLEDGMENTS

Most of the work described in this chapter was done while the authors were in residence at the University of Cincinnati Department of Geology. Fieldwork and data analysis by A. Önder Gökce were supported by the U.S. Geological Survey (USGS) and the University of Cincinnati. Data analysis and modeling by William Haneberg were supported by The Hillside Trust, the USGS, and the University of Cincinnati. Geotechnical testing was performed by personnel of the USGS soil mechanics laboratory in Golden, Colo., under the direction of Roger Nichols. Reviews by Rex Baum, Robert Fleming, Arvid Johnson, and David Nash helped to improve many parts of this paper.

PREVIOUS WORK

Iverson and Major (1987) describe both annual and short-term water-level fluctuations in a long, thin, northern California landslide, and note both that high-frequency fluctuations due to individual storms attenuate with depth and that the events of previous water years control annual water-level increases. These results are in line with those of two separate field studies by Gillham (1984) and Novakowski and Gillham (1988), who concluded that high antecedent moisture content resulted in more rapid water-table response to rainfall. Pyles and others (1987) observed asymmetric water-level fluctuations similar to those described in this paper during their studies of thin landslides in western Oregon. Harr (1977) monitored both saturated and unsaturated flow in a steeply sloping, forested hillside in the same western Oregon experimental forest used by Pyles and others (1987), and showed that flow was almost always unsaturated except near the foot of the hillside. He used measurements of stream and seepage-face discharge to show that the lower reaches of the hillside remained saturated due to drainage of soil pores farther upslope that had been filled during rainstorms. Upper reaches of the hillside became saturated at depth from time to time, but only after at least 3 mm of rain fell within 12 hours, and even then the saturation persisted less than 20 hours. Harr attributed these intermittent zones of saturation to the accumulation of infiltrating water along the contact between soil and underlying bedrock, where hydraulic conductivity decreased by an order of magnitude. Longitudinal water-table profiles from a natural hillside described by Anderson and Burt (1977a) show that water accumulates in the lower portion of the hillside while water level falls in the upper portions. In a related laboratory study, Anderson and Burt (1977b) show that a sloping, sand-filled gutter drains from the top down and a small saturated wedge remains at the base throughout the experiment. Measurements of discharge from the laboratory setup agree closely with those predicted using the slope of the water table as the hydraulic gradient, but do not agree with predictions that factor in the release of water from the unsaturated zone, suggesting that gravity drainage of unsaturated sand was negligible under the conditions investigated by Anderson and Burt (1977b).

SUBSEQUENT WORK

Since the revised manuscript for this chapter was returned to the U.S. Geological Survey for publication, additional work has almost completely superseded the mathematical model it describes.

Jackson and Cundy (1992), for example, developed a more complicated two-dimensional finite-difference model of modified Dupuit flow incorporating topographically convergent and divergent flow, which they used to successfully reproduce ground-water-level fluctuations observed in a steep hillslope hollow in the Cascade Mountains of Washington. Haneberg (1991a) (1) described short-term, predominantly unsaturated pressure-head fluctuations that were observed during the spring of 1988 at the Delhi Pike landslide complex, (2) presented a dimensional analysis of the partial differential equation governing flow through a variably saturated sloping layer in order to establish the conditions under which slope-normal and slope-parallel flow would predominate, (3) used a finite-difference model of slope-normal pore-pressure diffusion in an attempt to reproduce field measurements collected during the spring of 1988, and (4) concluded that storm frequency, storm intensity, and pre-storm fillable porosity influenced the hydrologic response of thin landslides to rainfall. Haneberg (1991b) applied the pore-pressure-diffusion model to published data from several landslides in order to calculate characteristic response times, and discussed how the diffusion model could be used to explain the nonlinear storm intensity-duration curves that are common products of empirical studies of rainfall thresholds associated with slope instability. This chapter was optimistically cited as Haneberg and Gökce (1991) in all three of the papers described above.

Finally, Haneberg (1992) presented a simple mass-balance model of subsurface flow through hillside soils, and compared the results of this model to field data from the Delhi Pike landslide. This work showed that the steady-state ratio of recharge to discharge at Delhi Pike, estimated using spring 1988 precipitation values, was near unity, and that simulations of unsteady flow could reproduce average pore pressure values observed at Delhi Pike.

GEOLOGIC SETTING AND SOIL PROPERTIES

Landslides in the Delhi Pike area (fig. 1) occur in illitic, bouldery-silty-clayey colluvium formed along the northern side of the Ohio River valley about 15 km west of downtown Cincinnati. Colluvium thickness ranges from less than a meter near hilltops to more than 10 m at river level, and the basal slip surfaces of the slides are generally along the contact between colluvium and weathered bedrock. Flat-lying, interstratified shale and limestone of the Upper Ordovician Kope Formation (approximately 80 percent shale and 20 percent limestone) and the overlying Upper Ordovician Fairview Formation (approximately 60 percent shale and 40 percent limestone) compose the local bedrock and are parent material for the overlying colluvium.

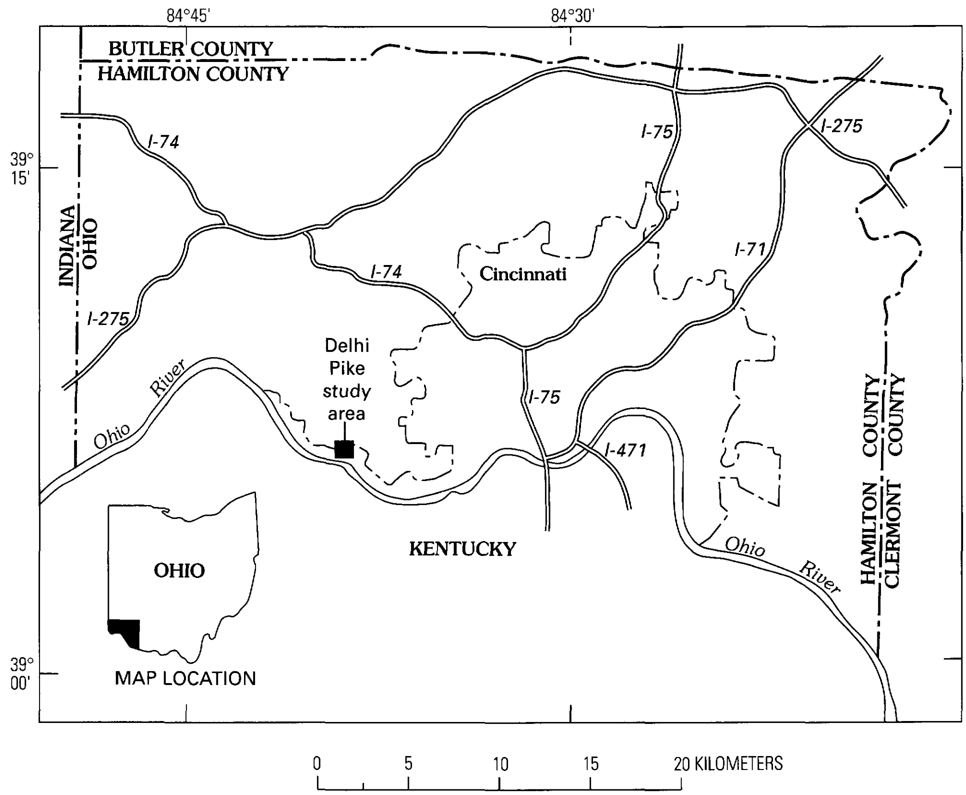
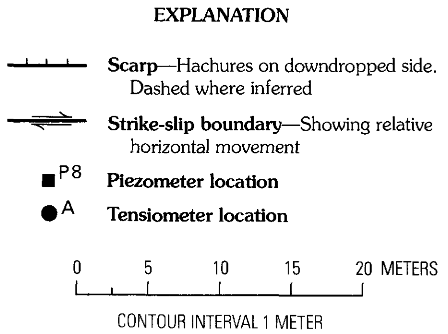


Figure 1. Location of Cincinnati, Hamilton County, the Ohio River, and the Delhi Pike study area.

For the most part, landslides in the Delhi Pike area are elongate parallel to topographic contours but, in some cases, this pattern is broken by narrow U-shaped landslides reaching farther uphill. The landslide that we describe is one of these U-shaped fingers, as shown in a detailed plane-table map (fig. 2). Its topography is generally irregular, with many small hummocks and scarps about a meter high, but the overall slope of the ground surface range from 20° to 30°, and the bedrock surface slopes approximately 20° across most of the hillside. The main slide mass is bounded by a left-lateral strike-slip zone along its entire eastern flank, a head scarp to the north, and a right-lateral strike-slip zone along the northern half of its western flank. All of these boundaries coincide with topographic depressions, in some places as much as 1 m deep. The southern half of the western flank grades into several east-west scarps and, hence, has no distinct boundary. The northern half of the slide mass is broken into a number of smaller slides, each of which has a subsidiary, U-shaped head scarp reflecting the shape of the slide as a whole. The vertical offset along these scarps is generally 1 m or less. Subsidiary scarps in the southern half of the slide, where the western boundary becomes indistinct, tend to be nearly linear and parallel to topographic contours. Locally, however, these parallel scarps branch and join to form an anastomosing network. The subtle increase in slope above the +30 m contour (all elevations are relative to an arbitrary control point on Delhi Pike) reflects the transition from Kope Formation to Fairview Formation bedrock.

The colluvium at Delhi Pike is mapped as Eden silty clay loam, a residual soil formed on moderate to steep slopes, in the Hamilton County Soil Survey (Lerch and others, 1982). It is described as well drained, with slow permeability and very rapid runoff (although our field observations showed very little runoff during storms). Lerch and others further state that excavations in the lower parts of slopes commonly have seepage problems and recommend that drains be installed around foundations. We had five test trenches excavated, using backhoes and larger tracked excavators, and found that within about 0.1 to 0.2 m of the surface the colluvium is grayish-brown and commonly breaks into peds, whereas colluvium 0.2 m or more below the surface is yellowish and massive. In places, however, brown colluvium is found in isolated lenses within the yellow colluvium. Both types of colluvium are composed of about 70 percent silty-clayey matrix and 30 percent limestone slabs as much as 0.3 m thick and 3 m long, with no apparent preferred orientation. The distribution of brown colluvium, yellow colluvium, large limestone slabs, and slip surfaces exposed in a longitudinal trench is shown in figure 3.



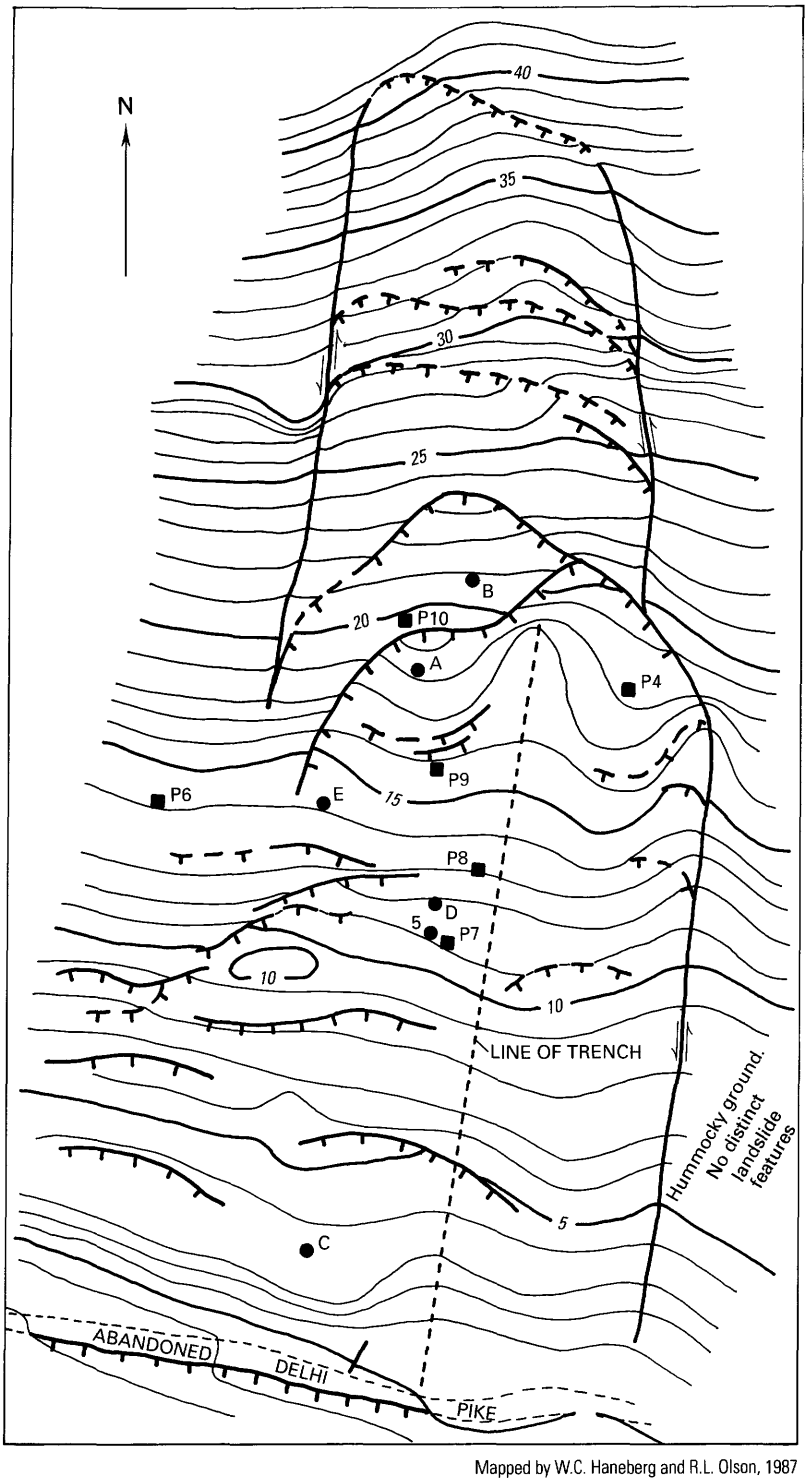


Figure 2. Plane-table geologic map of the portion of the Delhi Pike landslide complex studied in this paper. All elevations are relative to an arbitrary station on Delhi Pike. Data obtained from piezometer P–4 during the spring of 1980 are described in this paper. See Haneberg (1991a) for a discussion of data from other piezometers and tensiometers shown here.

HYDRAULIC CONDUCTIVITY

A falling-head permeability test was run on each of nine samples taken from different depths in a trench at Delhi Pike, and the results were averaged to obtain a single value for each sample. These average values range from a high of 2.63×10^{-5} m/s to a low of 7.70×10^{-8} m/s and have an arithmetic mean of 1.31×10^{-5} m/s and a geometric mean of 6.67×10^{-6} m/s (fig. 4); if the solitary value in the 10^{-8} range is excluded, the lowest average value becomes 3.26×10^{-6} m/s and the arithmetic mean becomes 1.47×10^{-5} m/s. With the exception of two errant values, hydraulic conductivity appears to be very nearly constant with depth. Both horizontal and vertical components of hydraulic conductivity were measured for one of the samples from a 1.2 m depth, and the average vertical conductivity is 5.11×10^{-6} m/s, whereas the average horizontal conductivity is 1.03×10^{-5} m/s. It is difficult to assess the significance of measurements on only one sample, but, if accurate, these results suggest that the colluvium at Delhi Pike is weakly anisotropic.

GRAIN SIZE, ATTERBERG LIMITS, AND CLAY MINERALOGY

Grain size, Atterberg limits, and clay mineralogy were determined using four samples taken from different depths in an auger hole, approximately 60 m southeast of piezometer P–4 but not shown in figure 2. There is very little variation in grain size distribution, Atterberg limits, or clay mineralogy with depth in the borehole (table 1). Both the colluvium and the weathered bedrock samples were composed primarily of clay-sized grains with lesser amounts of silt-sized grains. Gravel- and sand-sized grains composed less than 15 percent of sample weight, in all cases. Liquid limits average 46 percent and plastic limits average 24 percent; both decrease slightly with depth.

MEASUREMENT OF WATER-LEVEL FLUCTUATIONS

Our piezometers were cylindrical porous stones, 5 cm in diameter and 30 cm long, attached to a 2.5-cm-diameter (1-inch O.D.) PVC pipe. Each piezometer hole was backfilled with quartz sand to a level approximately 20 cm above the top of the porous stone and sealed with approximately 40 cm of bentonite (fig. 5). We measured water levels by hand in six piezometers at Delhi Pike at least twice a month between March 1979 and August 1980. In addition, we installed an automatic recorder in piezometer P–4 (see fig. 2 for location), approximately halfway up the landslide, to monitor piezometric levels at 15-minute intervals between late March and mid-April 1980. Our original intent was to use these short-term data to estimate piezometer lag time, but they also provided the first evidence of rapid water-level fluctuations in the Delhi Pike landslide complex. Subsequent field work during 1987 and 1988 has shown that P–4 often contains water when other piezometers are dry, and we have limited data to suggest that this was also true during 1980.

Table 1. Grain-size distribution, Atterberg limits, and clay type of colluvium in samples from a borehole at the Delhi Pike landslide complex.

Depth (m)	Grain size (weight percent)				Atterberg limits (percent)		Clay type
	Gravel	Sand	Silt	Clay	Liquid	Plastic	
1.2	1	6	26	67	50	25	Mixed layer, illite.
2.1	4	7	29	60	48	24	Mixed layer, illite.
4.8	1	4	31	64	45	24	Mixed layer, illite.
6.8	5	9	31	55	41	21	Illite, kaolinite.

GENERAL CHARACTERISTICS OF P–4 WATER-LEVEL RECORDS

We see several common characteristics in the water-level records that we describe below, which are typical of all our data from P–4. During the summer and early autumn, the bottom of P–4 is dry (water level at least 1.20 m below the surface), whereas during the late winter and spring of 1980, the background water level stood from 0.8 to 0.9 m below the surface. This background water level is punctuated by distinct increases of several decimeters that correspond to individual rainstorms. Short, intense storms generally produced sharp peaks, whereas gentle and prolonged rainfall generally produced flat-topped peaks. Regardless of rainfall intensity and duration, the water-level record peaks are asymmetric. In some cases, piezometric level rose virtually instantaneously in response to rainfall, and in other cases the rise in piezometric level was considerably slower, but piezometric levels always fell more slowly than they rose. After rainfall had stopped, piezometric levels characteristically fell slowly at first, then more rapidly, and then slowly again as the background piezometric level was approached; this pattern formed a sigmoidal falling water-level record showing that the water level continued to fall very slowly for many tens of hours, in the absence of additional rain. When more rain fell several hours after a storm, the fall of the water table slowed substantially. In some cases a small hump appears to be superimposed upon the overall slope of the falling water-level record, but in other cases there is no sign of a hump. We use specific examples below, in chronological order, to illustrate these generalities.

INDIVIDUAL STORMS

March 20 through March 22, 1980.—The first P–4 water-level record that we describe, illustrated in figure 6*A*, is flat-topped and results from a period of intermittent, relatively gentle rainfall distributed over about 17 hours. Base level was about 0.83 m beneath ground level both before and after the rainfall. Significant features of this water-level record are a compound rising limb reflecting two distinct storms on the evening of March 20, and a broad, gently sloping crest associated with gentle rainfall on the morning

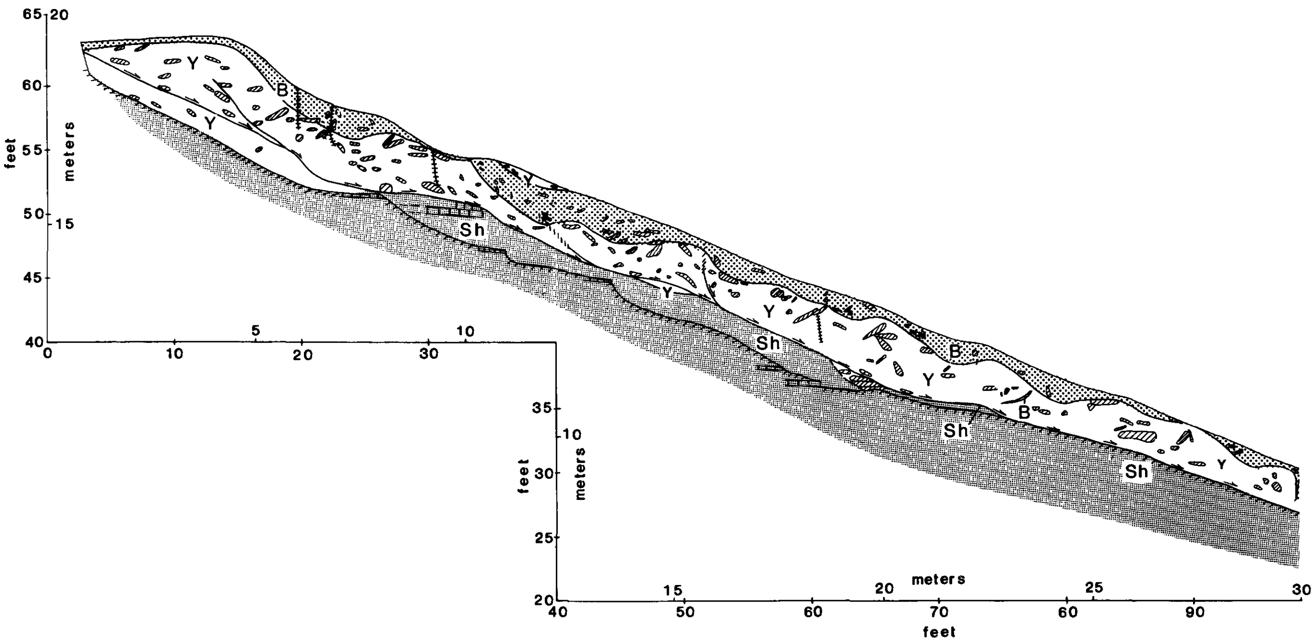


Figure 3 (above and facing page). North-south longitudinal cross section along the trench shown in figure 2. Section shows distribution of brown colluvium, yellow colluvium, limestone slabs, slip surfaces, and the contact between colluvium and weathered bedrock. Figure reduced from Fleming and Johnson (1994, pl.1).

of March 21. As with all of our records, the falling limb of the water-level record is sigmoidal in shape and is, on the whole, much flatter than the rising limb.

April 8 through April 18, 1980.—The P–4 water-level record for this 10-day period (fig. 6*B*) contains a closely spaced series of four storms and attendant water-level increases. All of the water-level-record peaks during these ten days were much sharper than the water-level record from March 20 through March 22 (fig. 6*A*), but the rates (slopes of the water-level-record limbs) of water-level changes are similar. What appears to be missing is the prolonged, intermittent rainfall that occurred during March 20 through March 22. Base level before and after this series of storms was from 1.12 to 1.15 m below ground level. During the first storm, approximately 11 mm of rain caused a water-level increase of only 3 cm, but in subsequent storms the water-level increase is much greater. Rainfall amounts totaling 4 mm on the 9th, 15 mm on the 11th, and 14 mm on the 13th resulted in water-level increases of 23, 51, and 54 cm, respectively. All four of these water-level records are distinctly asymmetric, as is typical of our data.

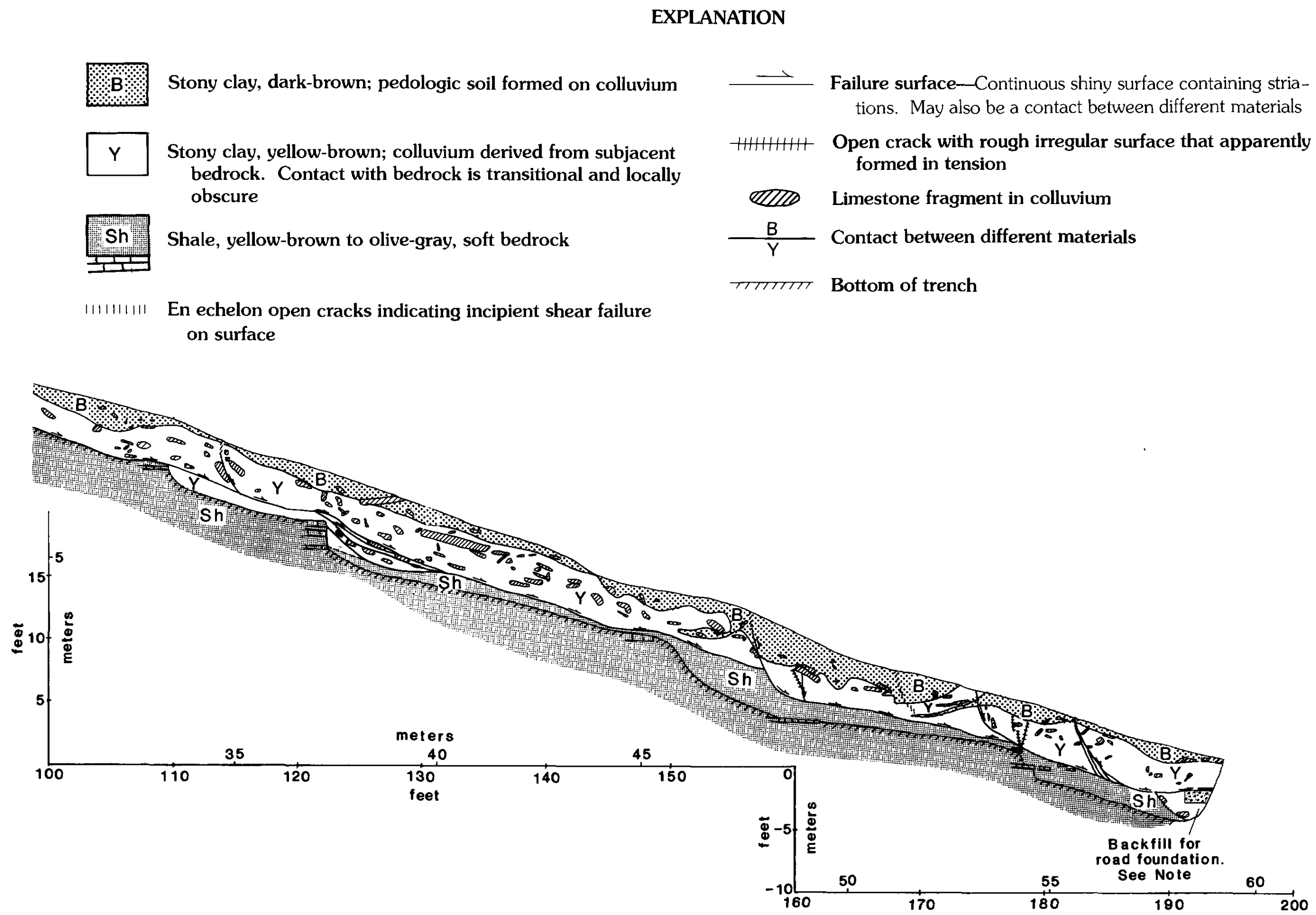
May 12 through 13, 1980.—We include this water-level record (fig. 6*C*) in our descriptions because it contains the most rapid piezometric-level increase (0.54 m/hr), the most rainfall in a single storm (40 mm), and the highest piezometric levels (0.30 m below ground level) we recorded at Delhi Pike. As shown in figure 6*C*, the rising water-level record limbs for both storms are very nearly vertical, suggesting a nearly instantaneous water-level increase.

Effective porosity.—We estimated maximum effective porosity for each storm as total rainfall divided by total change in water level. In doing this, we assume that all rainfall entered the soil and contributed to the water-table rise, so our estimates are maximum possible values. In all but one case, we found maximum effective porosity ranged between 1 percent and 7 percent for all of the storms during our period of record (fig. 7) and, in general, that the magnitude of water-level increase is proportional to the amount of rainfall. The lone exception occurred on April 8, 1980, when rainfall of 11 mm produced a piezometric level increase of 25 mm, for a maximum effective porosity of 44 percent. This storm will be discussed later in the chapter.

A MATHEMATICAL MODEL FOR HILLSIDE FLOW

MODIFIED DUPUIT FLOW

The well-known Dupuit approximation of ground-water flow through a phreatic aquifer, drawn from the observation that in most phreatic aquifers the water-table slope is very small (Bear, 1972, p. 361), incorporates the assumptions that flow is horizontal and equipotential lines are vertical. Boussinesq (1904) extended the Dupuit model to account for unconfined flow above a sloping impervious substrate but retained the assumption of horizontal flow, so the model works well only for very gentle slopes. Schmid



and Luthin (1964) derived an analytic solution for the Boussinesq model for steady-state flow and fixed-head boundaries, and Guitjens and Luthin (1965) used Hele-Shaw models to show that the solution is accurate for slopes as steep as 30 percent. Departing from the Boussinesq assumption of horizontal flow, Klute and others (1965) solved the problem of steady-state, two-dimensional, slope-parallel flow along an incline under fully saturated conditions. They found that flow through an inclined soil slab that has a length-to-thickness ratio greater than 10:1 could be accurately represented by a one-dimensional, slope-parallel flow model. Childs (1971) and Towner (1975) give analytic solutions for steady-state, unconfined, slope-parallel flow both with and without recharge. Chauan and others (1968) and Ram and Chauan (1987) present analytic solutions for the problem of transient slope-parallel flow between two trenches on a slope with constant recharge. Karadi and others (1968) solve the related problem of transient flow between drains resting on a shallow, impermeable substrate, and Wong (1977) has published nomographs for the problem of flow to drains in sandy leachate collection layers overlying clay landfill liners. Beven (1977) used a two-dimensional finite-element model of saturated and unsaturated flow, and predicted that saturated zones, controlled by pre-existing pressure heads, would form by accumulation of water at the toes of hillsides. More recently, Beven (1981, 1982) and Hurley and Pantelis (1985) have used linearized, one-dimensional approximations to successfully model some aspects of saturated and unsaturated hillside discharge hydrographs. Hurley and Pantelis show that flow through a thin sloping aquifer is indeed one-dimensional to a first approximation, but they solve the problem of flow through unsaturated soil in terms of total soil moisture flux.

We approximate hillside flow, at least as we have observed it, as unconfined flow in a fully saturated aquifer. Bear (1979, p. 74–75) discusses limitations of the "fully saturated" approach, particularly the error introduced by

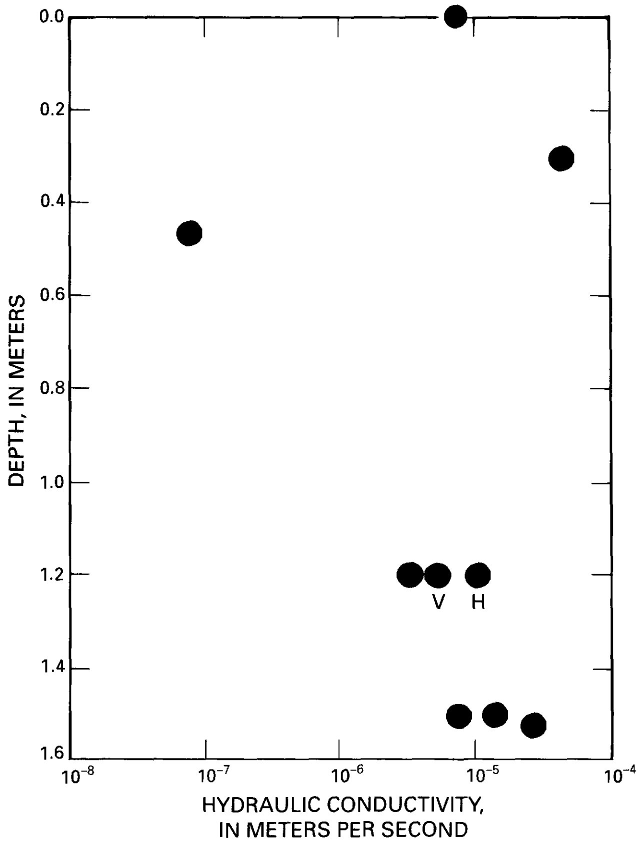


Figure 4. Variation of colluvium hydraulic conductivity with depth. Points labeled *H* and *V* are values for horizontal and vertical conductivity of closely spaced samples from a depth of 1.2 m.

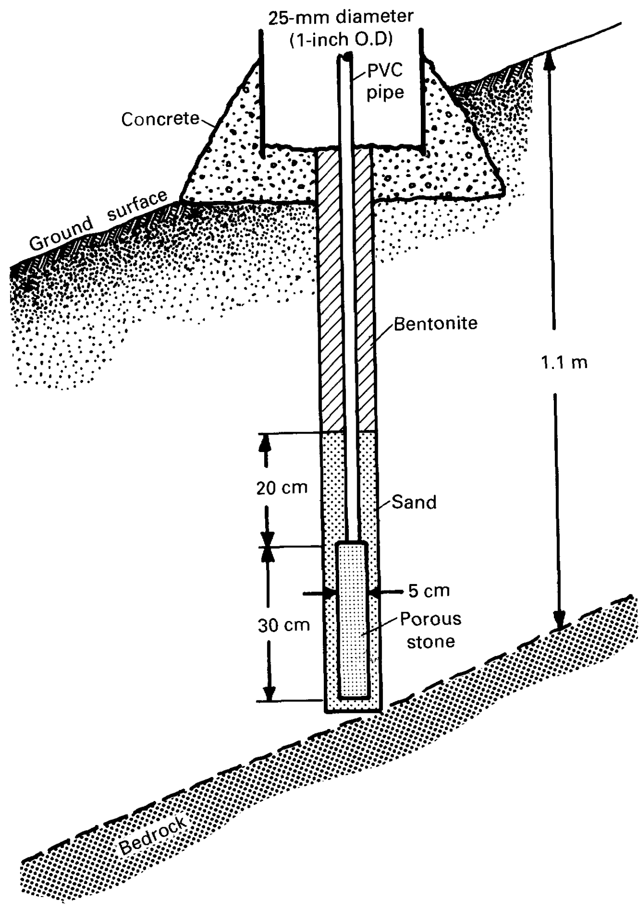


Figure 5. Schematic drawing of piezometer P–4 at Delhi Pike. See figure 2 for location.

ignoring the capillary fringe, or zone of tension saturation, above the water table. As we will show, however, our model has the advantage of predicting water-table height rather than total soil moisture flux, as would a more rigorous model that considered both saturated and unsaturated zones, and it can account for many important characteristics of our field data.

THE GOVERNING EQUATION

For a thin, porous slab underlain by a sloping, impervious substrate, we define total hydraulic head, Φ, as:

$$\Phi = h \cos \beta - \xi \sin \beta \qquad (1)$$

(compare with Beven, 1981; Hurley and Pantelis, 1985) where ξ is the slope-parallel coordinate, giving distance from the upper end of the slab to the point of measurement [in units of length], h is height of the phreatic surface [length] measured perpendicular to the substrate (fig. 8), and β is slope angle [degrees]. Combining equation 1 with Darcy's law, the governing equation for incompressible one-dimensional flow through a homogeneous, fully saturated, unconfined aquifer is:

$$\frac{\partial h}{\partial t} = \frac{K}{p}\frac{\partial}{\partial \xi}\left(h\frac{\partial \Phi}{\partial \xi}\right) + \frac{R}{p} \qquad (2)$$

where p is effective porosity, K is saturated hydraulic conductivity [length/time], R is net recharge (R is positive for precipitation and negative for evapotranspiration) to the water table [length/time], and t is time. Positive recharge is not necessarily the same as rainfall, so the two should not be used synonymously unless there is evidence to suggest that all rainfall reaches the water table. Because we have not observed any runoff at Delhi Pike except in areas where we have removed the organic soil and exposed colluvium at the surface during fieldwork, we will assume that a large proportion of rain finds its way to the water table and that recharge is approximately equal to rainfall when weather is cloudy and cool during winter and spring storms. We also use effective porosity, p, in place of the more common specific yield on the assumption that both the water and the soil are incompressible, so that there is no elastic storage (Bear, 1972, p. 376). For the limiting case of p=0, corresponding to a state of tension saturation, the temporal derivative in equation 2

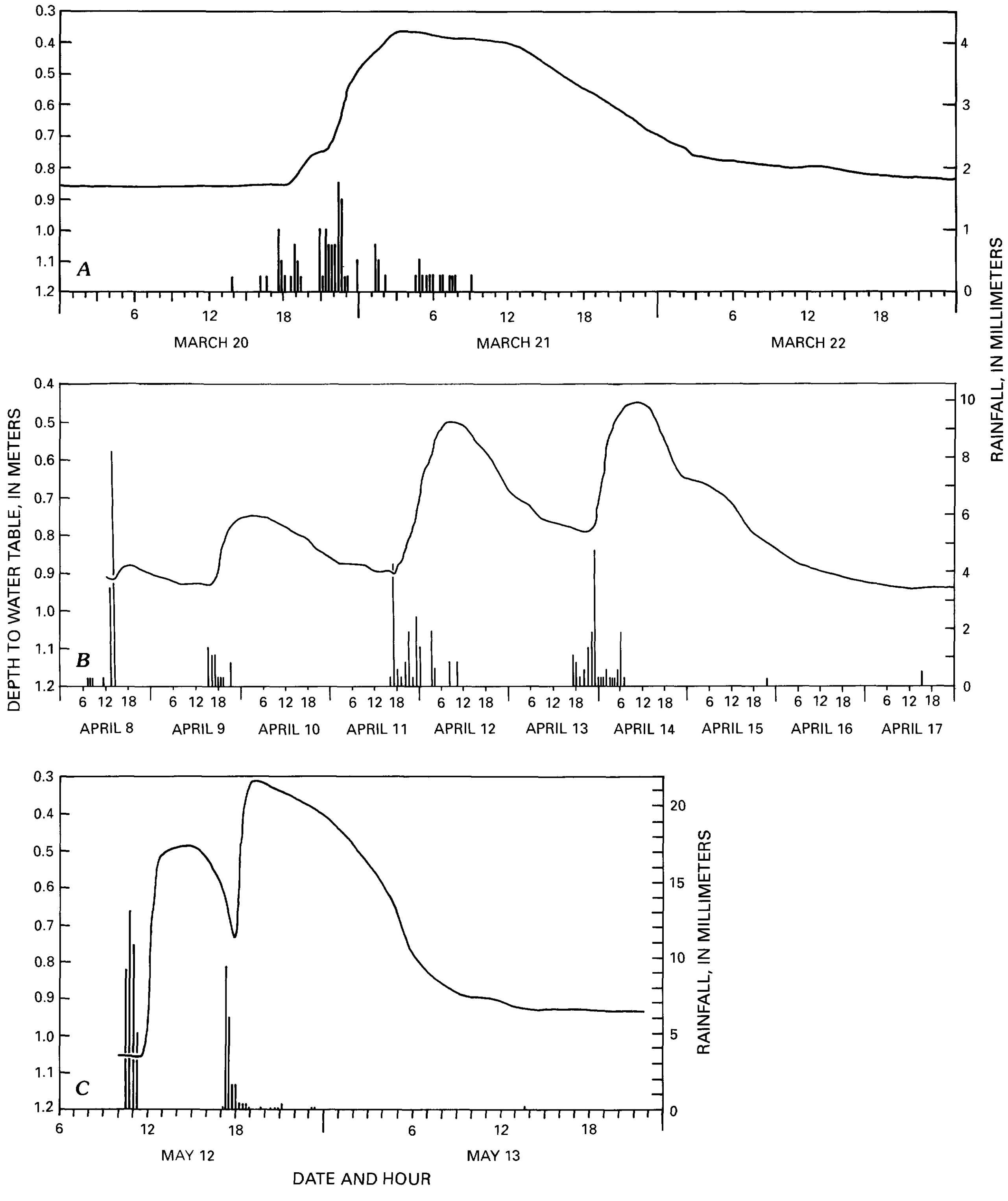


Figure 6. Rainfall and piezometric level in piezometer P–4 for three rainy periods in the spring of 1980. Vertical bars represent cumulative rainfall totals during 15-minute intervals (graphs *A* and *C*) or during 1-hour intervals (graph *B*). *A*, March 19–22, an example of a flat-topped record in which water levels remain high due to prolonged gentle rain following an initial storm or storms. *B*, April 8–18, an example of a sharp-peaked record in which water levels rise and fall rapidly following a short-lived intense storm. In this record, the slopes of the rising limbs have nearly the same absolute values as the maximum slopes of the falling limbs. *C*, May 12–13, another example of a sharp-peaked record in which water levels rise and fall rapidly following a short-lived intense storm. In this record, the rising limbs are nearly vertical and are much steeper than the falling limbs.

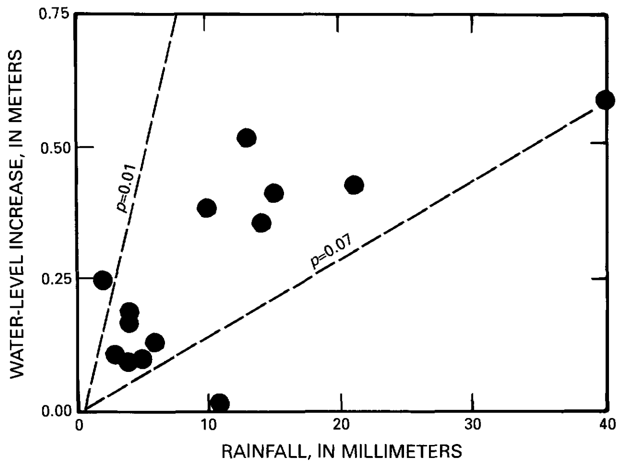


Figure 7. Change in water level versus rainfall. With one exception, maximum effective porosity (p) estimates range from 0.01 to 0.07, suggesting that the colluvium was nearly saturated before rapid water-level increases.

goes to infinity and the piezometric level should rise instantaneously in response to rainfall.

Differentiating equation 1 with respect to ξ and substituting the result into equation 2, we arrive at:

$$\frac{\partial h}{\partial t} = \frac{K}{p}\frac{\partial}{\partial \xi}\left(\frac{\partial h}{\partial \xi}h\cos\beta - h\sin\beta\right) + \frac{R}{p} \tag{3}$$

or, equivalently:

$$\frac{\partial h}{\partial t} = \frac{K}{p}\left[\frac{\partial^2 h}{\partial \xi^2}h\cos\beta + \left(\frac{\partial h}{\partial \xi}\right)^2\cos\beta - \frac{\partial h}{\partial \xi}\sin\beta\right] + \frac{R}{p} \tag{4}$$

both of which are nonlinear in h and have no known analytic solutions for time-variant recharge.

DIMENSIONAL ANALYSIS OF THE GOVERNING EQUATION

For a porous slab of length L and thickness D, we introduce the dimensionless variables:

$$H=h/D$$
$$X=\xi/L$$
$$T=t/t_r$$
$$t_r=pD/K$$

as well as the scaling factor:

$$\sigma=D/L$$

and then we normalize equation 4 to arrive at:

$$\frac{\partial h}{\partial T}= \left[\frac{\partial^2 H}{\partial X^2}H + \left(\frac{\partial H}{\partial X}\right)^2\right]\sigma^2\cos\beta - \frac{\partial H}{\partial X}\sigma\sin\beta + \frac{R}{K} \tag{5}$$

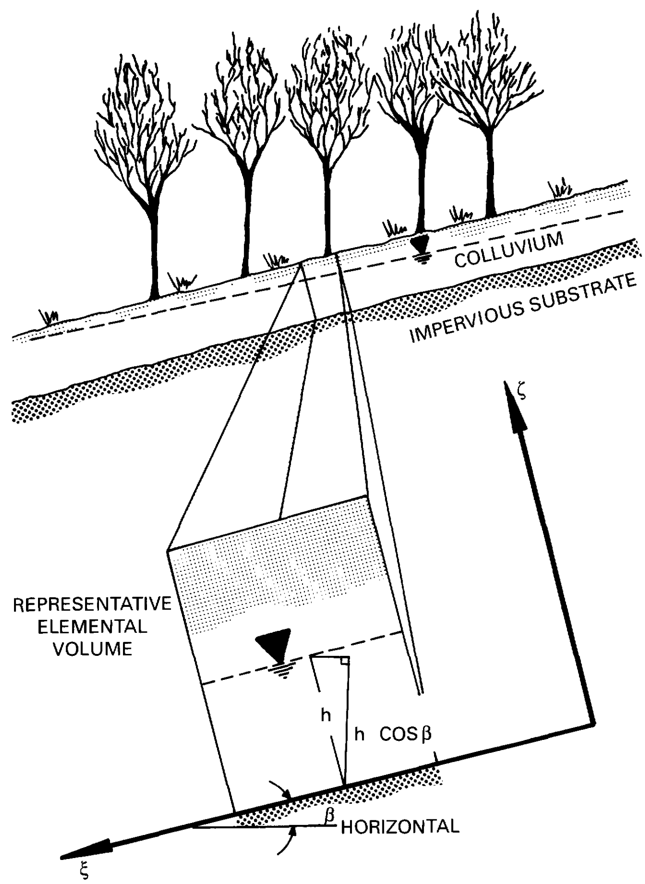


Figure 8. Idealization of a thin slab of uniform thickness resting atop a sloping, impervious substrate and the representative element used to derive the governing equation for hillside flow through a homogeneous, isotropic, phreatic aquifer.

We chose to use a characteristic time of $t_r=pD/K$ because we are interested principally in short-term water-level changes rather than the long-term fluctuations that would occur on a scale of $t_r=pL/K$. Thus for a slab on the order of $D=10^0$ m, $10^{-2}<p<10^{-1}$, and $K=10^{-5}$ m/s, typical of the hillsides at Delhi Pike, we will be looking at changes that occur on a time scale of $10^3<t_r<10^4$ s, or hours to tens of hours. For long, thin slabs all terms containing σ will be negligible; hence, integration to solve equation 5 for H yields:

$$H = \int_{T_1}^{T_2}\frac{R}{K}dT + H_0 \tag{6}$$

where H_0 is uniform water level along the slope at T_1. Equation 6 shows that water levels will rise and fall uniformly along the entire slope, with no component of slope-parallel flow over the short term. Pre-storm values of p, which affect the magnitude of t_r, will change with time, thereby changing the magnitude and duration of rainfall necessary to produce a given water-level increase. Thus prediction of rainfall-induced water-level changes, even using this simple model, requires detailed knowledge of pre-storm effective porosity, hydraulic conductivity, rainfall intensity, and rainfall duration.

One way to produce slope-parallel flow is to redefine the characteristic time of the system as $t_r=pL/K$, so that typical values of $L=10^2$ m, $10^{-2}<p<10^{-1}$, and $K=10^{-5}$ m/s give a characteristic time of $10^5<t_r<10^6$ s, or about 1 to 10 days. Using this new definition, equation 4 can be written as the nonhomogeneous kinematic-wave equation (Lighthill and Whitham, 1955; Miller, 1984):

$$\frac{\partial H}{\partial T}+\frac{\partial H}{\partial X}\sin\beta = \frac{R}{\sigma K} \qquad (7)$$

which will produce slope-parallel, wavelike flow on a time scale $1/\sigma$ times as long as the uniform changes predicted by equation 6. Another way to produce slope-parallel, wavelike flow with the model is to return to our original definition of $t_r=pD/K$, set R=0, and assume that for a nearly saturated hillside $\sigma\approx p$. To first order in σ, then, equation 4 can be rewritten as the homogeneous kinematic-wave equation:

$$\frac{\partial H}{\partial T} = -c\frac{\partial H}{\partial Z} \qquad (8)$$

where

$$c=\sin\beta$$

is the celerity of the traveling waveform. For a phreatic surface with periodic perturbations of arbitrary wavelength λ and amplitude A_0, a general solution to equation 8 is:

$$H = A_0\sin\left[\frac{2\pi}{\lambda}(X-cT)\right] \qquad (9)$$

which can be verified by differentiating with respect to X and T, and then substituting the results into equation 8. Thus net recharge, due to either infiltration or evapotranspiration, must be insignificant in order for slope-parallel flow to be significant over the short term. It is also possible for water-level increases to occur much more quickly than water-level decreases (1) if the pre-storm effective porosity is much less than the drainable porosity of the colluvium, which is essentially a constant for a given soil (Bear, 1972, p. 484), or (2) if recharge rates are much higher than evapotranspiration rates.

FINITE DIFFERENCE FORMULATION

We assume that spatial grid increments are uniform, solve for the second derivative, and linearize by letting coefficient h values lag one time increment behind (Anderson and others, 1984, p. 337). In order to minimize the error introduced by letting coefficients lag over large temporal increments, we use the alternate method of Douglas and Jones (1963, p. 199). This method uses a predictor to estimate values of h at time $t+(\delta t/2)$, which are then substituted into a Crank-Nicholson corrector to solve for values of h at time $t+\delta t$. For interior nodes, the predictor form of equation 4 is:

$$h_{i-1}^{j+1/2}-2\left[1+\frac{p(\delta\xi)^2}{K\delta t\, h_i^j\cos\beta}\right]h_i^{j+1/2}+h_{i+1}^{j+1/2}=\frac{-2p(\delta\xi)^2}{K\delta t\cos\beta}$$

$$-\frac{1}{4h_i^j}(h_{i+1}^j-h_{i-1}^j)^2+\frac{\xi\tan\beta}{2h_i^j}(h_{i+1}^j-h_{i-1}^j)+\frac{R(\delta\xi)^2}{Kh_i^j\cos\beta} \qquad (10)$$

where i is the spatial grid index and j is the temporal grid index. The corrector form of equation 4 for interior nodes is

$$h_{i-1}^{j+1}-2\left[1+\frac{p(\delta\xi)^2}{K\delta t\, h_i^{j+1/2}\cos\beta}\right]h_i^{j+1}+h_{i+1}^{j+1}$$

$$=\frac{-2p(\delta\xi)^2}{K\delta t\, h_i^{j+1/2}\cos\beta}-\frac{1}{4h_i^{j+1/2}}(h_{i+1}^j-h_{i-1}^j)^2$$

$$+\frac{\xi\tan\beta}{2h_i^{j+1/2}}(h_{i+1}^j-h_{i-1}^j)+\frac{R(\delta\xi)^2}{Kh_i^{j+1/2}\cos\beta}$$

$$-h_{i-1}^j+2h_i^j-h_{i+1}^j \qquad (11)$$

The predictor, equation 10, uses a forward-difference temporal derivative and is only first-order accurate in time. The Crank-Nicholson corrector, equation 11, uses a central-difference temporal derivative and is second-order accurate in time. Both the predictor and the corrector are second-order accurate in space, so the combined accuracy over one entire time increment is $O[(\delta\xi)^2+(\delta t)^{3/2}]$ (Douglas and Jones, 1963).

BOUNDARY AND INITIAL CONDITIONS

We chose to fix water-table height at both the uphill and downhill ends of our numerical model to account for the steady base flow, represented by relatively constant water levels in P–4 between storms that we observed at Delhi Pike, and also to ensure that water levels near the lower boundary did not exceed soil thickness. As shown in the perturbation analysis above (equations 5 and 6), however, boundary effects act only near the ends of the problem domain, so the choice of a specific set of boundary conditions will not greatly affect the behavior of the model along most of the slope. For a fixed piezometric level at spatial node 1, the boundary form of the predictor, equation 10, expanded around neighboring spatial node 2 is:

$$-2\left[1+\frac{p\,(\delta\xi)^2}{K\,\delta t\, h_2^j\cos\beta}\right]h_2^{j+1/2} + h_3^{j+1/2} = \frac{-2p\,(\delta\xi)^2}{K\,\delta t\cos\beta}$$

$$-\frac{1}{4h_2^j}\,(h_3^j-h_1^j)^2 + \frac{\xi\tan\beta}{2h_2^j}\,(h_3^j-h_1^j) + \frac{R\,(\delta\xi)^2}{Kh_2^j\cos\beta} - h_1^{fixed} \qquad (12)$$

and the accompanying form of the corrector, equation 11, is:

$$-2\left[1+\frac{p\,(\delta\xi)^2}{K\,\delta t\, h_2^{j+1/2}\cos\beta}\right]h_2^{j+1} + h_3^{j+1}$$

$$= \frac{-2p\,h_2^j\,(\delta\xi)^2}{K\,\delta t\, h_2^{j+1/2}\cos\beta} - \frac{1}{4h_2^{j+1/2}}\,(h_3^j-h_1^j)^2$$

$$+\frac{\xi\tan\beta}{2h_2^{j+1/2}}\,(h_3^j-h_1^j) + \frac{R\,(\delta\xi)^2}{Kh_2^{j+1/2}\cos\beta}$$

$$-2h_1^{fixed} + 2h_2^j - h_3^j \qquad (13)$$

Both the predictor and the corrector for the downhill boundary expanded around the (n–1)th node are similar in form to equations 12 and 13.

To solve the model, we used the standard Thomas algorithm (Press and others, 1986; Anderson and others, 1984) for tridiagonal linear systems to solve finite-difference approximations of the forms given in equations 10 and 11 with boundary approximations of the forms given by equations 12 and 13, and we verified our algorithm using an analytical steady-state perturbation solution of equation 4. The computer program and the analytic solution are given in appendixes C and E of Haneberg (1989).

NUMERICAL RESULTS

On the basis of our theoretical analyses and field observations, we chose representative values for all of the constants in equation 2 and then experimented with the numerical model in order to produce synthetic water-level records similar to those we recorded at Delhi Pike. First, we attempted to match actual water-level records with synthetic water-level records from our model. Then, after we were satisfied that the model could account for our field observations, we experimented with different combinations of rainfall intensity and duration to see what effect changes in these variables would have on water-level record shapes.

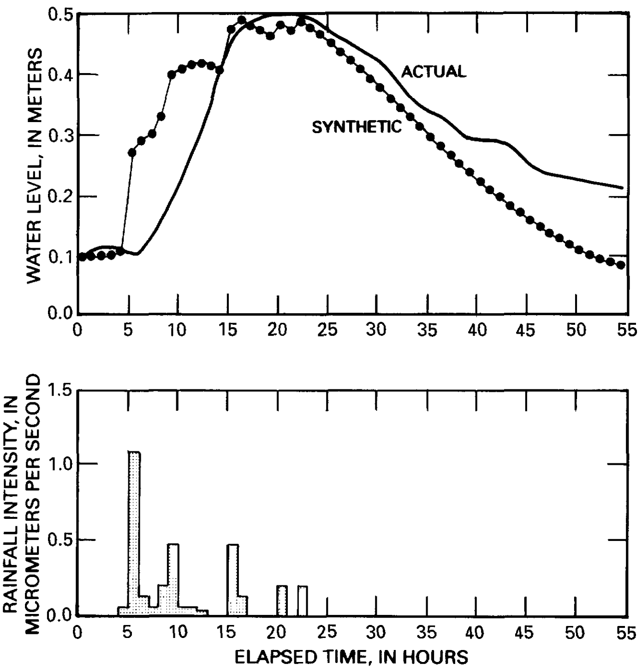


Figure 9. Comparison of actual and synthetic water-level records from P–4 on April 11–12, 1980. Effective porosity in the numerical model was lowered from the estimated maximum effective porosity of 0.035 to 0.025, and recharge to the model water table was set equal to actual rainfall.

In order to better understand the effects of rainfall patterns on water-level record shape, we decided to use relatively simple rainfall distributions rather than the more complex patterns we recorded at Delhi Pike.

SIMULATION OF P–4 WATER-LEVEL RECORDS

By trial and error, we found it difficult to model field water-level records that show very abrupt changes or cover a period including multiple storms, such as the one we have described for May 12 through May 13, 1980 (fig. 6*C*). However, we were able to obtain good results for simple, single storms in which water-level records changed gradually.

In general, we found that the model is very sensitive to changes in effective porosity, p, hydraulic conductivity, K, and recharge, R. For example, figure 9 compares actual and calculated water-level records for the storm of April 11 and 12, 1980. We used values of p=0.025, K=1.3×10^{-5} m/s, δt=3600 s, $\delta\xi$=5 m, β=22°, upslope head fixed at 0.1 m, downslope head fixed at 1.2 m, and recharge, R, equal to measured rainfall. Except for effective porosity, p, we selected these representative values from our field observations. Calculated maximum effective porosity for

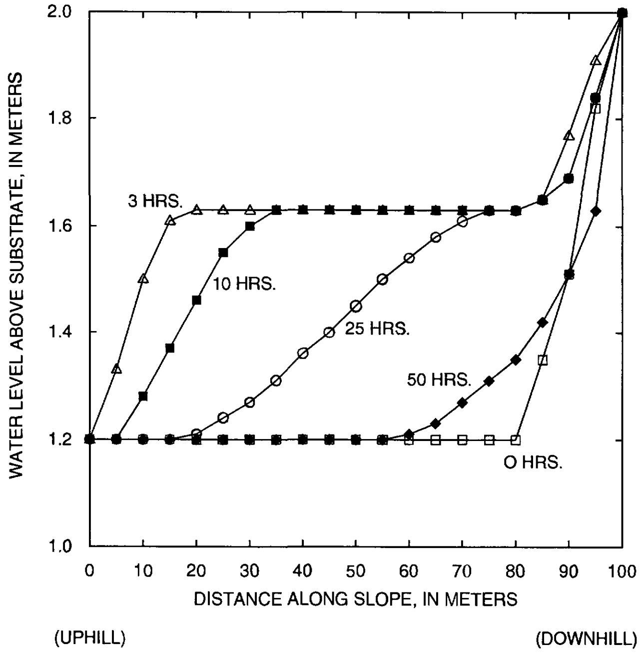


Figure 10. Model results for drainage of a fully saturated, thin, homogeneous aquifer of uniform thickness resting atop a sloping, impervious substrate. Water level rises almost uniformly along the entire slope, but drains from head to the toe of the hillside. Variables used were p=0.01, K=1.3x10^{-5} m/s, β=22°, h=2 m, and L=100 m. Rainfall intensity was 0.1 µm/s for the first hour, 0.5 µm/s for the second and third hours, 0.1 µm/s for the fourth hour, and zero thereafter.

this storm, which assumes that all of the rain contributes to the measured water-level increase, was 0.035, but our trials showed that p=0.025 gave better results.

Another variable is location along the slope and, again by trial and error, we found that synthetic water-level records from the finite difference node at ξ=15 m (measured from the top of the slope) gave the best match with field data from P–4. The synthetic water-level record in figure 9 begins to rise steeply as soon as rainfall begins, whereas the actual field water-level record begins to rise slowly 2 hours after rainfall begins. Furthermore, the synthetic water-level record is very sensitive to changes in rainfall intensity, producing an irregular curve, whereas the actual water-level record is smooth and does not reflect variations in rainfall intensity. Both water-level records peak at about 0.50 m and 22 hours elapsed time. (Note that we give model results in water-table height above the impermeable substrate and elapsed time, as compared to water depth beneath the surface and time of day for our field data). The synthetic water-level record falls more rapidly to its fixed-head boundary base level of 0.10 m, whereas the actual water-level record falls to about 0.20 m before leveling off and shows several small perturbations on its falling limb.

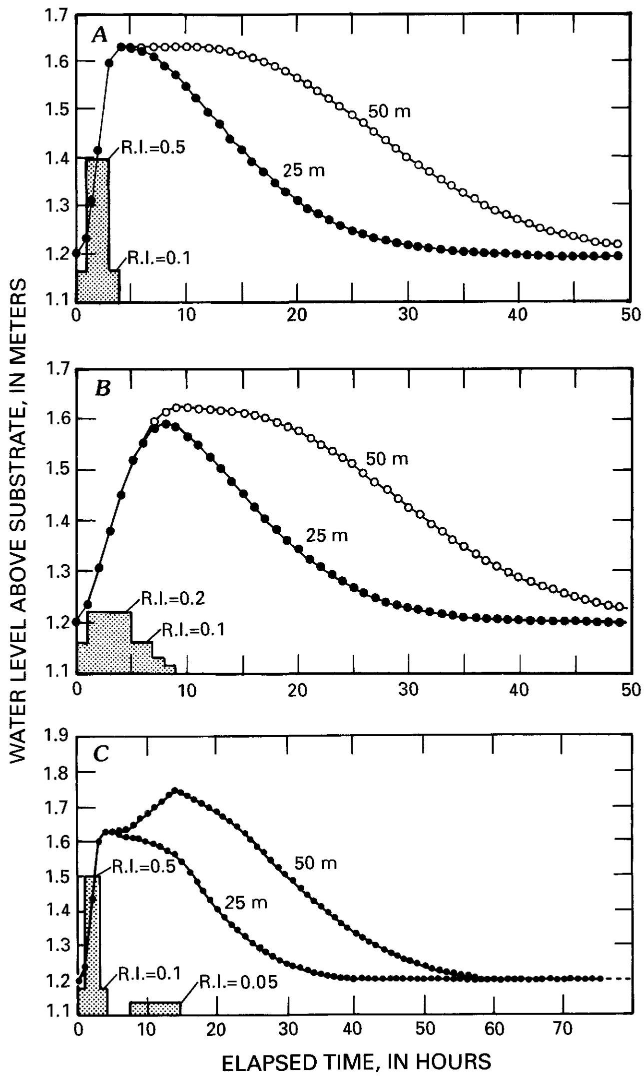


Figure 11. Synthetic water-level records showing projected response to three different patterns of rainfall. Curves are for finite difference nodes located 25 and 50 m from the top of the hillside. Shaded bars show pattern of rainfall intensity. R.I. = rainfall intensity in micrometers per second. *A*, Response to a short, intense rainstorm providing 4 mm of net recharge to the water table. *B*, Response to gentle rainfall providing 4 mm of net recharge. Note the differences between *A* and *B* produced by variations in the intensity and duration of rainfall, even though net recharge was the same. *C*, Response to a short, intense storm followed by several hours of gentle rainfall. Initial storm provided 4 mm of net recharge, as in *A* and *B*, but the additional recharge significantly decreased the rate of decline following the storm.

Simulation of short, intense rainfall.—Our numerical experiments with short, intense rainfall produced distinctly asymmetric water-level records with very steep rising limbs. Figures 10 and 11*A* show that for one simulation of this type, water level began to rise almost immediately after rainfall began, and peaked shortly after rainfall stopped. Although water level rose uniformly at all of the interior finite-difference nodes, it fell much more rapidly near the top of the hillside than at its midpoint, and the water-level record from ξ=50 m is much flatter than the water-level record from ξ=25 m. Total recharge for the simulation shown in figure 11*A* was 4 mm.

Simulation of prolonged, gentle rainfall.—We simulated the effects of a prolonged, gentle rainfall providing the same amount of recharge as the short, intense rainfall simulation that we describe above (4 mm). Figure 11*B* shows that the ξ=25 m water-level record peaked slightly below and before the ξ=50 m water-level record. Except for this offset of peaks, the falling limbs are virtually identical to those in figure 11*A*. The rising limb, however, is much flatter and is no steeper than the steepest portion of the falling limb.

Simulation of intense rainfall followed by gentle rainfall.—We observed that in some cases—for example, the water-level record of March 19–22, 1980—a small amount of rain distributed over several hours seemed to maintain the water table at a high level until the rain stopped. So, we modeled such a situation using an initial short, intense storm identical to that in figure 11*A* followed by 7 hours of very light but continuous rain (0.05 μm/s). The results of our simulation (fig. 11*C*) show that water level first peaked at the same level and time as in figure 11*A* and then began to fall slowly when the rain stopped. Before the water level could fall very much, however, the light rain began and the slope of the falling limb for ξ=25 m remained constant until the rain stopped, and then fell. The ξ=50 m water-level record, however, fell only until the second round of rain began, and then rose to peak at about 1.75 m while the ξ=25 m water-level record was falling. Thus, the effect of a small, secondary storm differs along the length of the hillside.

DISCUSSION

Our simple, theoretical model accounts for the shapes and amplitudes of asymmetric, short-term piezometer hydrographs from the Delhi Pike landslide complex surprisingly well if we assume that pre-storm effective porosity was on the order of a few percent when our field data were collected. Most importantly, we have shown that rapid water-level fluctuations should occur uniformly along the entire slope over the short term if either rainfall or evapotranspiration is significant. Slope-parallel, wavelike flow is controlled by higher order terms in the governing equation that are significant only over the long term. Slope-parallel flow will be significant over the short term only if both rainfall and evapotranspiration are negligible, and then only because slope-normal flow vanishes.

Although we have developed a conceptually simple model, we stress that it is also extremely sensitive to variations in all four of the values needed to predict rapid water-level fluctuations: rainfall intensity, rainfall duration, hydraulic conductivity, and pre-storm effective porosity. Because these critical values are extremely difficult to assess in complex natural hillsides, we are wary of attempts to predict sudden hillside failures during rainstorms by correlating so-called threshold rainfall with slope instability, particularly when such predictions may have a bearing on human life and limb (for example, Keefer and others, 1987). We recommend that instead of relying solely on statistical correlations, any future slope instability warning system should include monitoring of prestorm soil moisture conditions in areas where sudden failure would have significant economic or social consequences.

Numerical solutions of our model, using reasonable values for the hillside at Delhi Pike, agree in both shape and magnitude with our field data but, like the first-order analytical model, the solutions are also sensitive to variations in rainfall, conductivity, and porosity estimates. Differences between actual and synthetic water-level records, produced using the model (fig. 9), give us some insight into the hillside hydrologic system. The first major difference, a noticeable time lag between the onset of rainfall and piezometer response, can be attributed largely to the time required for rainwater to infiltrate through the unsaturated zone. For example, if we assume that R was constant with time and p=0.03 before the storm of April 11, 1980, we can solve equation 7 for $R=5\times10^{-6}$ m/s, using the observed 7200-s time lag, which suggests that the water table is being recharged at a rate of about $K/2$. Examination of Murdoch's (1987) laboratory data for colluvium from Delhi Pike shows that unsaturated hydraulic conductivity for p=0.03 is about two-thirds to one-half of saturated hydraulic conductivity, so a 7200-s time lag for a piezometer tip about 1 m deep in slightly unsaturated colluvium is reasonable. Because the hydraulic conductivity of unsaturated soils is a highly nonlinear function of pressure head, both the time lag and the slope of the rising hydrograph will vary exponentially with pre-storm effective porosity, which is also a function of pressure head. Once the soil does become saturated, however, water can flow into the piezometer tube, and Hvorslev's (1951) method predicts a response time for P–4 of about one minute, so the effect of the piezometer tube itself on response time should be negligible.

The differences between the falling limbs of the actual and synthetic water-level records in figure 9 suggest to us that water is being introduced from a source not considered in the model, perhaps bedrock seeps. Localized seepage from fractured limestone layers in the bedrock would also explain the steady base flow observed in P–4 between storms and is consistent with observations of weakly artesian bedrock springs throughout the Cincinnati area (Baum, 1983).

There are not enough high-frequency water-level data uphill and downhill from P–4 to tell us how well our model, or any other model, applies to the entire hillside. The data do tell us, however, that sampling frequencies of several days or weeks can be too low to resolve the effects of single storms at Delhi Pike. The Hamilton County Soil Survey (Lerch and others, 1982) does describe common seepage problems along only the lower portions of Eden silty clay loam slopes, so there is some evidence for slope-parallel drainage in hillsides such as those at Delhi Pike, and similar results have been reported elsewhere.

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