marked ...rol that topography exerts on runoff production, and serves to emphasise the necessity of including convergent flow in hollows in any distributed model of hillslope soil water movement at this scale. The topographic effect has been shown to relate not only to the generation of a throughflow peak immediately after the storm, but since the hollow remained saturated throughout the month long recession, topography also has an effect on the maintenance of base flow.

CONCLUSIONS

The automatic monitoring of soil moisture tension has been accomplished by the use of a scanivalve fluid switch and an associated tensiometer grid This system was employed to monitor conditions on a hillslope over a storm and subsequent recession. The resulting spatial pattern of tension values was shown to accord with stream discharge over time. In addition the effect of topography in controlling convergent flow in hollows was exhibited, and it is therefore suggested that the spatial variation of soil moisture is a more significant factor in runoff production than has hitherto been acknowledged.

ACKNOWLEDGEMENTS

The authors are grateful to Simon Godden for drawing the figures, to David Tuff for constructing the electronic circuitry, and to Lee Williams for useful discussions concerning the scanivalve construction.

REFERENCES

Bear, J., Zaslavsky, D. and Irmay, S., 1968. Physical Principles of Water Percolation and Seepage. UNESCO, Paris, 465 pp.
Dunne, T. and Black, R.D., 1970. An experimental investigation of runoff production in permeable soils. Water Resour. Res., 6: 478–490.
Fitzsimmons, D.W. and Young, N.C., 1972. Tensiometer — pressure transducer system for studying unsteady flow through soils. Trans. Am. Soc. Agric. Eng., 15: 272–275.
Hewlett, J.D. and Hibbert, A.R., 1963. Moisture and energy conditions within a sloping soil mass during drainage. J. Geophys. Res., 64: 1081–1087.
Kirkby, M.J. and Chorley, R.J., 1967. Throughflow, overland flow and erosion. Bull. Int. Assoc. Sci. Hydrol., 12: 5–21.
Weyman, D.R., 1973. Measurements of the downslope flow of water in a soil. J. Hydrol. 3: 267–288.

(#1/2)

Journal of Hydrology, 33 (1977) 37–58

WATER FLUX IN SOIL AND SUBSOIL ON A STEEP FORESTED SLOPE

R.D. HARR

Forestry Sciences Laboratory, Pacific Northwest Forest and Range Experiment Station, Forest Service, U.S. Department of Agriculture, Corvallis, Oreg. 97331 (U.S.A.)

(Received December 11, 1975; revised and accepted May 6, 1976)

ABSTRACT

Harr, R.D., 1977. Water flux in soil and subsoil on a steep forested slope. J. Hydrol., 33: 37–58.

Soil analyses, tensiometer, piezometer and precipitation data were used to evaluate water flux in soil and subsoil on a steep forested slope. Although unsaturated flow predominated in the soil during fourteen moderate-size winter storms, discontinuous saturation of upslope subsoil did occur at depths of 110–150 cm after at least 30 mm of rain fell in 12 h or less. This saturation persisted less than 20 h. Very slight changes in soil-water pressure head during storms effected relatively large changes in both magnitude and direction of water flux. Maximum water fluxes in the top meter of soil were 3–4.5 mm/h and maximum flux in the subsoil was 0.5 mm/h, slightly less than the minimum flux in the soil. Between storms, the vertical component of flux at the 10-cm depth was much less than the downslope component but equaled the downslope component during storms. Conversely, vertical components of flux at 70- and 130-cm depths were much less than downslope components during storms but equaled downslope components between storms. Analysis of piezometric levels, discharge from a seep, and streamflow revealed an abrupt decrease in rate of water flux to the lower part of the slope about 10 h after the end of rainfall. This decrease corresponded with nearly complete draining of larger pores that had filled with water during storms. Subsurface flow and channel interception averaged 97 and 3% of storm flow which, in turn, averaged 38% of total storm precipitation. No overland flow was observed.

INTRODUCTION

Overland flow rarely occurs on undisturbed forest soils in the western Cascade Range in Oregon, but streams respond quickly to precipitation. Steep slopes and high permeabilities of surface soils (Dyrness, 1969) are conducive to rapid, shallow subsurface flow which would account for quick response of streams. Additional information about this important hydrologic process has been lacking for this region.

That overland flow is a rare occurrence in forested watersheds in humid regions is well documented, and the importance of subsurface flow to streamflow has been stressed repeatedly (Whipkey, 1965, 1969; Hewlett and Hibbert, 1967; Weyman, 1970; Hewlett, 1974). Only a small part of a watershed produces storm runoff. This part, which appears to expand and contract according to changes in rainfall intensity and soil conductivity, was the basis for the variable source area concept (U.S. Forest Service, 1961). Reports of field studies of the interaction between unsaturated and saturated flow and streamflow have generally supported this concept (Ragan, 1968; Betson and Marius, 1969; Dunne and Black, 1970a, b; Rawitz et al., 1970), although these reports have tended to replace the word "variable" with "partial" and imply source area as being relatively fixed (Hewlett, 1974). Recently, mathematical models have been developed to couple transient unsaturated—saturated flow to streamflow (Freeze, 1972a, b; Stephenson and Freeze, 1974; Hewlett and Troendle, 1975) in attempts to supplement these field studies. Other field investigations have shown that subsurface water may move rapidly through piping channels (Jones, 1971) or through biologically created non-capillary channels (Whipkey, 1965, 1969; Patric and Swanston, 1968; Chamberlin, 1972) in otherwise unsaturated soils. Other studies have sought to describe subsurface flow on sloping ground in other areas (van 't Woudt, 1954; Minshall and Jamison, 1965; Takeda and Ishii, 1968; Megahan, 1972; Wilson and Ligon, 1973; Rogowski et al., 1974).

This study was undertaken to obtain additional information about the lateral subsurface flow phenomenon in western Oregon. Specific questions to be answered were:

(1) What is the magnitude and direction of water flux in soil during and after typical rainstorms?

(2) What soil properties and other site factors influence magnitude and direction of water flux in the soil?

(3) What are the relative importances of water fluxes under saturated and unsaturated soil conditions?

(4) How is the timing of subsurface flow related to rainfall and to streamflow?

STUDY AREA AND METHODS

The study was conducted in watershed *10*, a 10.23-ha watershed in the U.S. Forest Service's H.J. Andrews Experimental Forest about 72 km east of Eugene, Oregon. Watershed *10* is the location of one of the intensive study sites of the Coniferous Forest Biome's analysis of ecosystems and lies within the western hemlock zone (Franklin and Dyrness, 1969). Overstory vegetation is primarily 450-year-old Douglas-fir with younger western hemlock and Douglas-fir. Annual precipitation averages 235 cm, about 80% of which falls between October and April during frequent, long duration, low to moderate intensity frontal storms. Although occasional accumulations of snow are generally short lived, snowmelt concurrent with large amounts of rain has produced the maximum runoff events of record.

The study area is located on a stream-to-ridge portion of the south aspect of watershed *10* (Fig. 1). The slope is slightly convex and its gradient averages

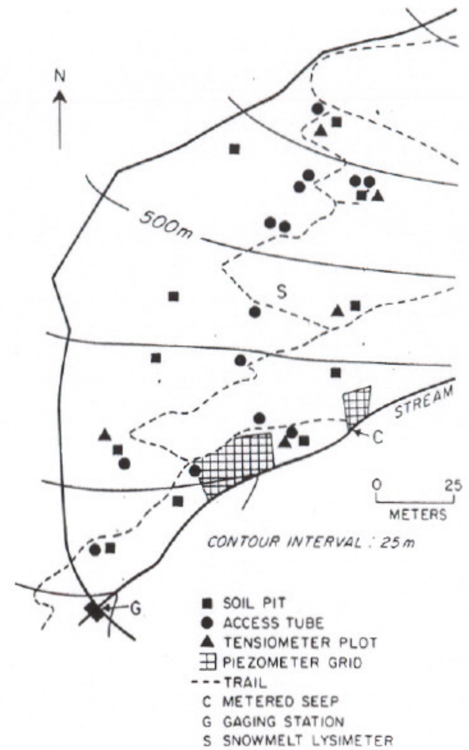

Fig. 1. Map of the study area.

75%, ranging from 50% near the ridge to 110% adjacent to the stream. Elevation ranges from 425 to 530 m. Generally, slightly more than 1 m of relatively poorly developed soil is underlain by 2—7 m of subsoil (saprolite) consisting of highly-weathered coarse volcanic breccias which formed as a result of mudflows and pyroclastic flows (Swanson and James, 1975). In the upper half of the main stream channel and the entire length of the tributary stream, water flows over this saprolite. The lower half of the main channel was scoured to bedrock by a debris avalanche that originated in the upper portion of the south aspect of the watershed. Thus, water in the lower half of the main channel flows over relatively unweathered bedrock and could be collected and its rate of flow measured. This ease of flow measurement was one criterion for selection of the study slope.

Imped. water infilt.

Soil and subsoil properties

During initial phases of study, hydrologic properties of the soil and subsoil were determined. A total of 450 undisturbed soil samples were taken from eleven soil pits throughout the study area (Fig. 1). A bulk-density sampler (Blake, 1965) was used to remove at least six samples from depths of 10, 30, 70 and 110 cm in the soil, and 130 and 150 cm in the subsoil as soil pits were dug as deeply as possible. Generally, visible discontinuities in physical properties of the soil at the time of soil sampling were located between these depths. In several pits, samples were also taken at depths of 200 and 250 cm. Soil samples were analyzed for saturated hydraulic conductivity with a constant-head permeameter that accommodated undisturbed samples (Ranken, 1974); pore-size distribution and drainage characteristics by the tension-table method (Vomocil, 1965); and particle-size distribution by the hydrometer method (Day, 1965). Total porosity and bulk density were also obtained from the tension-table data. Tensions applied equaled 10, 20, 30, 40, 60 and 100 cm of water. Additional details of soil analyses are given elsewhere (Ranken, 1974).

In soil and subsoil in the study area, structural characteristics affect hydrologic properties more than do textural characteristics. Texture changes little with depth (Table I). Most surface soils are gravelly clay loams, lower soil layers are gravelly silty clay loams or clay loams, and subsoils are gravelly clays or clay loams. Surface soils are well aggregated, but lower depths (70–110 cm) show subangular blocky structure with less aggregation than surface soils. Subsoil exhibits a massive structure reflecting many characteristics of the weathered pyroclastic parent material.

TABLE I

Mean values of selected soil and subsoil properties for the soil pit uppermost on the study slopes

Depth (cm)	Bulk density (g/cm^3)	Particle-size distribution (%)			Porosity (%)	Amount of porosity (%) in pores with diameter		Saturated hydraulic conductivity (cm/h)
		sand	silt	clay		> 0.30 mm	< 0.030 mm	
10	0.807	22.5	38.8	38.7	60.8	38.4	46.0	352
30	0.897	29.8	34.1	36.1	63.8	40.7	47.2	412
70	1.015	28.5	33.8	37.7	60.3	23.4	65.2	163
110	0.981	28.8	33.0	38.2	63.1	21.4	69.3	175
130	1.080	30.2	31.4	38.4	55.4	11.3	80.6	16
150	1.053	34.0	31.3	34.7	57.6	7.9	83.2	22

Differences in soil aggregation from depth to depth most likely account for differences in pore-size distribution and hydraulic conductivity. Although total porosity changes only slightly with depth, pore-size distribution and hydraulic conductivity change markedly (Table I). Significant shifts occur between 30 and 70 cm and between 110 and 130 cm involving primarily pores 0.03 and 0.3 mm in diameter. Lower depths have less porosity in the 0.3-mm diameter class and more porosity in the 0.03-mm diameter class than do upper depths. Differences in pore-size distribution among depths are further reflected in the drainage curves for soil (Fig. 2). At the bottom of the slope the more massive structure characteristic of the saprolite is located closer to the ground surface. Most likely, past landslides in this area of over-steepened slopes adjacent to the stream removed the more developed surface soils.

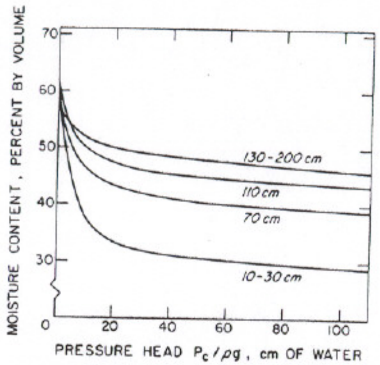

Fig. 2. Soil-water characteristics for soil and subsoil.

Instrumentation

Magnitude and direction of water flux at several depths in the soil and subsoil were estimated by 72 tensiometers in five plots within 5 m of soil pits (Fig. 1). A tensiometer plot consisted of two or four groups of tensiometers. Half of a four-group plot is shown in Fig. 3. Individual tensiometers of a group were spaced 20 cm apart and were located at depths of 10, 30, 70, 110, 130 and 150 cm, the soil sampling depths at nearby soil pits.

Water input to the study slope was measured continuously with a modification of the snowmelt lysimeter described by Haupt (1969). Although the lysimeter could measure both rainfall and snowmelt, water input consisted of only rainfall for the period described in this paper.

Water was observed flowing from the base of the study slope and from other slopes in the lower half of watershed *10*, in relatively well-defined seeps of various sizes and shapes and spaced various distances apart (Brown et al., 1972). These seeps are highly localized zones of saturated soil rather than pipelike channels described by Jones (1971) and Chamberlin (1972) and are apparently related to the microrelief of the relatively unweathered bedrock near the stream or to the presence of vertical, andesitic dikes (Swanson and James, 1975). Some saturated zones have been traced 12 m upslope.

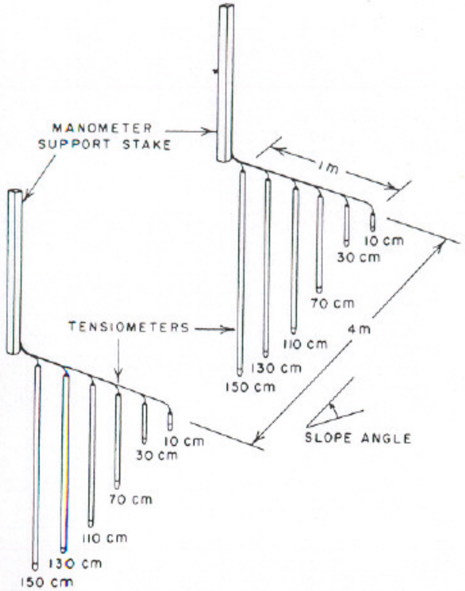

Fig. 3. Relative location of tensiometers within a tensiometer plot.

During and between major storm events, virtually no water was observed entering the stream from soil located between outlets of the seeps.

The nature of water flow in the region immediately adjacent to the stream was determined by two grids of piezometers installed in the vicinity of two seeps (Fig. 1). Each piezometer consisted of 1.9-cm PVC pipe with a perforated lower end covered with fine-mesh nylon screen. After piezometers were placed in holes power-augered to bedrock, holes were backfilled with sand and bentonite in a manner similar to that described by Swanston (1967). The smaller grid contained ten piezometers placed 40—90 cm deep and spaced 1—2 m apart, and the larger grid contained 34 piezometers at a depth of from 48 cm to over 4 m and spaced 2—4 m apart.

Preliminary observations indicated considerable fluctuation in piezometric levels over relatively short periods of time and among piezometers. Because of this variation, the difficulty in visiting each piezometer frequently enough, and the damage that would be done to the fragile, highly porous surface soils during frequent visits, an electronic water-level measuring and recording system was devised (Holbo et al., 1975). This system, which operates under battery power, was capable of measuring and recording water level in sixteen piezometers simultaneously every half hour.

In addition, flow from one large seep downslope from the small grid of piezometers (Fig. 1) was caught in a metal collector and routed to a large

tipping-bucket measuring device. This seep is located in or upslope from an andesitic dike that trends in a north-northwesterly direction from the stream. Attempts to trace this dike upslope and to locate other similar dikes were unsuccessful. Thus the size of the source area of this seep is unknown.

RESULTS AND DISCUSSION

Soil-water pressure head

Measurements of soil- and subsoil-water pressure head were made at 4-h intervals during fourteen winter storms in the 1973/1974 winter rainy season, and at 8-h intervals between these storms. Storm events were defined as periods of major rainfall separated by at least 6 h of rainfall intensities averaging less than 0.1 mm/h. Storm sizes ranged from 16 to 127 mm. Average intensity of storms was about 3 mm/h, ranging from less than 0.1 mm/h to 22 mm/h.

An example of fluctuation in soil- and subsoil-water pressure head is shown in Fig. 4. These data are from one pair of tensiometer groups in the tensiometer plot uppermost on the study slope (Fig. 1) and represent only two storm events. Fluctuations of similar magnitude were observed at other locations on the study slope during other storms, although relative differences in pressure head with depth varied somewhat from place to place. Minimum pressure head observed at any depth during the study period was —28

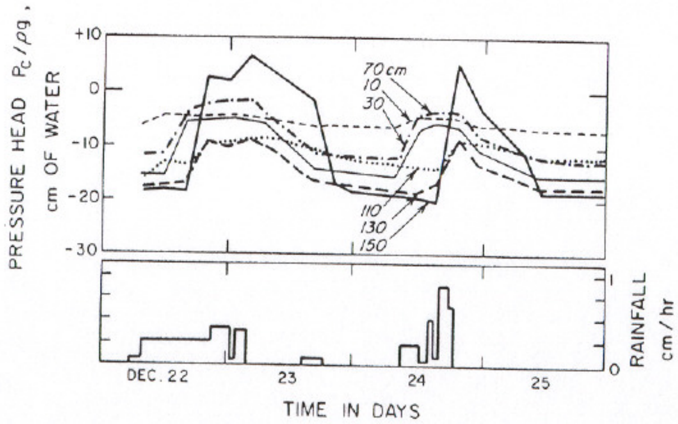

Fig. 4. Fluctuation in soil-water pressure head at highest elevation tensiometer plot for two storms.

cm of [illegible]. Subsequent discussion will center largely on two groups of tensiometers, data from one of which are shown in Fig. 4, but it also applies to most other groups of tensiometers located elsewhere on the study slope.

An important feature of Fig. 4 is the positive pressure head detected at the 150-cm depth near the end of the storm event on December 22 and again after the storm event on December 24. Results of the soil analyses had suggested that saturation would most likely occur at 130 cm, the level at which saturated hydraulic conductivity shows the greatest decrease (Table I) because of an abrupt change in pore-size distribution. The positive pressure head shown in Fig. 4 indicates a saturated zone at 150 cm with a maximum depth of 6.6 cm above the tensiometer level. A saturated layer was not detected at any other depth at this location nor at any depth in the downslope group of tensiometers. However, saturated subsoil was detected frequently at the 110-cm level in the other pair of tensiometer groups in this plot and at the 110- or 130-cm levels in other plots. Local variation in the location of the soil—subsoil boundary and the critical change in pore-size distribution probably account for the variation in the depth at which saturation occurred. Non-uniform saturation occurred only four times during the 3-month study period, after at least 30 mm of rain fell in less than 12 h. Saturation did not persist longer than 20 h at any location.

Tensiometer data were used in a two-dimensional analysis of water flux. Total flux consisted of a vertical component, e.g., between the 10- and 30-cm depths in one tensiometer group, and a downslope (parallel to slope or lateral) component, e.g., between the 10-cm depths of a pair of tensiometer groups. In either case, flux was determined by the general equation:

$$q = K(P_c)i \tag{1}$$

where q = water flux, K = hydraulic conductivity as a function of capillary pressure P_c, and i = hydraulic gradient. Hydraulic gradient is described by:

$$i = \Delta H/L$$

where H = hydraulic head = $Z + P$ at each tensiometer, and L = distance between tensiometers in the direction of the flux. Z is gravitational head (elevation of a tensiometer cup), and P is the pressure head of water.

Unsaturated hydraulic conductivity

Conductivity of soil and subsoil were determined for transient unsaturated conditions following an empirical method described by Brooks and Corey (1966). This method uses saturated hydraulic conductivity and the shape of the moisture release curve of an undisturbed soil sample to estimate unsaturated conductivity as a function of capillary pressure. Unsaturated hydraulic conductivity was estimated for each tensiometer depth by:

$$K(P_c) = K_s(P_b/P_c)^{\eta} \tag{2}$$

where $K(P_c)$ = unsaturated conductivity, K_s = saturated hydrau[illegible] conductivity, P_b = the bubbling pressure of the soil or subsoil (approximately the minimum P_c on the drainage cycle at which air is continuous in the soil), P_c = capillary pressure, and $\eta = 2 + 3\lambda$. Both P_b and λ are derived from soil and subsoil desorption data. Brooks and Corey (1966) refer to λ as an index of pore-size distribution: λ is large for porous media having a narrow range of pore sizes and small for media having a wide range of pore sizes. Sands, for example typically exhibit λ values around 3.5 indicating a narrow range of pore size. Mean values of λ and $P_b/\rho g$ are given in Table II (ρ = density of water; g = acceleration due to gravity). The relatively low values of λ shown here indicate a wide range of pore sizes at all depths.

TABLE II

Mean values of minimum and maximum unsaturated conductivity at various depths for the soil pit uppermost on the study slope as determined by Brooks and Corey's (1966) method

Depth (cm)	λ	η	$P_b/\rho g$	Saturated hydraulic conductivity, K_s (cm/h)	Unsaturated hydraulic conductivity, $K(P_c)$ (cm/h)	
					minimum	maximum
10	0.386	3.16	0.38	352	0.05	0.46
30	0.396	3.19	0.22	412	0.01	0.36
70	0.199	2.60	0.57	163	0.06	11.4
110	0.171	2.51	0.36	165	0.01	0.06
130	0.097	2.29	0.93	16	0.02	0.09
150	0.104	2.31	1.2	22	0.03	22*

*Saturation occurred at this depth. At $P_c/\rho g = 0$, $K(P_c) = K_s$.

Water flux

Water fluxes in the vertical and downslope directions were summed to obtain resultant flux. The magnitude of the resultant flux is shown graphically in Fig. 5 and can be written:

$$q_R = [(q_D + q_V \sin\alpha)^2 + (q_V \cos\alpha)^2]^{0.5} \tag{3}$$

where q = water flux; R, D, V denote resultant downslope and vertical directions; and α = slope angle. Flux angle γ becomes:

$$\gamma = \sin^{-1}(q_D \cos\alpha/q_R) \tag{4}$$

Theoretically, γ can very from 0° (vertical, downward) through 180° (vertical, upward) to 360° (vertical, downward).

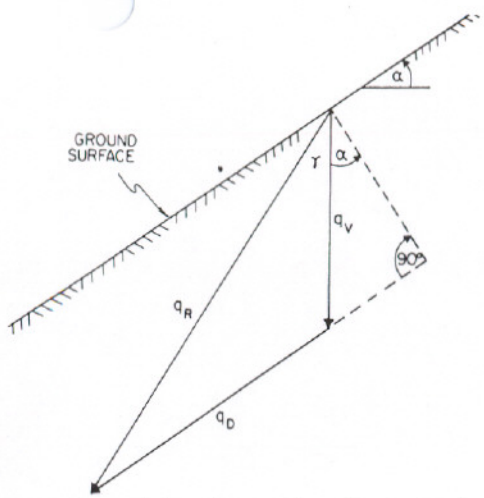

Fig. 5. Vertical, downslope, and resultant water fluxes.

The magnitude and direction of resultant water fluxes during and between two storm events are shown in Fig. 6 for three depths. At this uppermost slope position, slope angle is 28° (53% slope). Flux parallel to the slope would therefore have a flux angle of 62° (90°—28° = 62°), and downward vertical flux would have a flux angle of 0°. Fig. 6 is based partly on the data shown in Fig. 4.

Fig. 6A shows fluctuations in magnitude and direction of water flux between 10 and 30 cm. Here, flux varied from about 0.9 mm/h between storms to 2—4 mm/h during storms. As expected, response to rainfall was most rapid at this shallow depth. Flow direction at 10 cm showed a fairly consistent pattern of fluctuation during the experimental period (Fig. 6A). Between storms, when flux was minimum, flow was directed mostly downslope. Conversely, during storms, when water flux was considerably greater than between storms, the vertical and downslope components of flow were about equal. Thus, water moving from 10 to 30 cm during or shortly after a storm acquired a substantial downslope component.

Water flux between 70 and 110 cm is shown in Fig. 6B. Note that the inverse relationship between flux and flux angle did not occur at this depth. Generally when flux was low, flow was more downward. As flux increased during storms, it also became directed more downslope. Flux at this depth varied between about 0.5 and about 4.5 mm/h.

Water flux in the subsoil is shown in Fig. 6C. This figure differs substantially from either Figs. 6A or 6B in two ways. First, the magnitude of water flux was considerably less in the subsoil than in the top meter of soil. Although magnitude of flux in the subsoil more than doubled during storms, maximum fluxes were still less than minimum fluxes in the top 110 cm of soil. Thus, the top 110 cm of soil at this slope position appears to be the most active zone of downslope movement of water. The second difference in flux in the subsoil concerns flux angle. The flux angle at 130 cm at this location frequently exceeded 62°, the angle of flow parallel to the slope. This large flux angle was caused by the saturated zone at 150 cm in the upslope group of tensiometers. Because of this saturation, hydraulic head was greater at 150 cm than at 130 cm, where subsoil was still unsaturated, causing q_V in Fig.5 to be vertical upward and the resulting flux angle γ to exceed 62°.

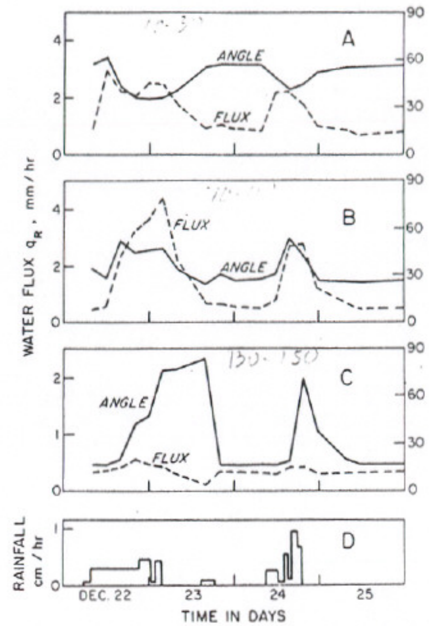

Fig. 6. Water flux at (A) 10 cm, (B) 70 cm and (C) 130 cm, in relation to rainfall (D).

That flux angles were always greater than 0°, indicating a substantial lateral component of flow, agrees with Zaslavsky and Rogowski's (1969) theoretical analysis of lateral flow. In this analysis, a lateral component was predicted to occur in sloping soil wherever a vertical decrease in permeability exists.

Malfunctions of one or more tensiometers in other tensiometer plots precluded flux analyses for other slope positions; thus, comparisons of water flux at different slope positions are not possible. In subsequent discussion, soil-water fluxes described above will be assumed to be reasonably applicable to [illegible]out the lower 10–15 m of the study slope. Similarity of properties exhibited by soil and subsoil, from midslope to upslope positions (Ranken, 1974) tends to support the validity of this assumption. Additionally, intermittent tensiometer readings indicate that capillary pressures were similar to those measured at the plot uppermost on the slope. Further support is added by an analysis of η, P_b and K_s for the entire study slope. These parameters vary with depth just as do those of the soil pit uppermost on the slope (Table II) and for which similar analyses were made.

Although computed fluxes between tensiometers were very small if one or more tensiometers were located in unsaturated soil, computed water fluxes in the subsoil would have been much greater if both upslope and downslope tensiometers at that depth had been in saturated subsoil. If a saturated zone were parallel to the slope of the ground surface, the hydraulic gradient in this saturated zone could be approximated by the sine of the slope angle, about 0.47 at this plot. Laboratory measurements show saturated hydraulic conductivities for the subsoil here could be 20–50 cm/h. Thus, fluxes of 10–25 cm/h directed entirely downslope (no component vertically downward) could have been possible if the mid- to upslope saturated zones were continuous and connected with the seeps at the bottom of the slope. Such a network of saturated zones could comprise a subsurface drainage system similar to those described for other areas (Whipkey, 1969; Chamberlin, 1972) and would partly account for the rapid hydrologic response of watershed *10* to precipitation. But if the midslope to upslope saturated zones were isolated as measurements of soil-water pressure head suggest, then water movement through these zones would be limited by the relative hydraulic conductivity of the unsaturated subsoil immediately downslope from the saturated zones, a situation that is approximated in Fig. 6C.

Saturated zones do occur elsewhere on watershed *10*. Access tubes installed throughout the watershed for soil-moisture measurements with a neutron soil-moisture meter consist of 3.66-m sections of tubing connected by a non-watertight joint. During many storm events, water has entered through this joint in some access tubes located on mid- to upslope positions on the north aspect of the watershed. Water entry could occur only if the subsoil surrounding these joints became saturated and soil-water pressures were positive. The 1.70–3.0-m depth of these joints suggests considerably more variation in the vertical location of saturated zones than was observed on the study slope. No estimates of water fluxes in these saturated zones are available.

Slope–stream linkage

At some point in a sloping soil which supplies water to a stream, saturation must occur so that a hydraulic gradient exists from the soil to the stream. Earlier, transient saturated zones at some midslope to upslope locations on the study slope were described. But saturated zones, because of their spatial and temporal limits, probably had little effect on the overall hydraulic behavior of the slope. Additional saturated zones did occ[illegible] certain locations along the base of the slope. The conductivity of saturated zones and the supply of water to them by unsaturated drainage from above determined their upslope extent. Measurements of water level in piezometers indicate the saturated zones extended both upslope and laterally during runoff events. However, because of discontinuity of the saturated zones and problems associated with installing piezometers in grids of small enough spacing, exact upslope and lateral extents are unknown.

Some insight into the hydraulic linkage between the slope and the stream can be obtained, however, by analysis of streamflow and saturated flow conditions in soil at the bottom of the slope. Streamflow, discharge from the metered seep, and piezometric level upslope from the seep all responded relatively quickly to precipitation (Fig. 7). Because of channel interception, the effects of rainfall were first evident in streamflow. For example, prior to the storm on December 22, streamflow had been decreasing after the peak of the December 20 storm. At 05^h30^m, rainfall began, and channel interception noticeably slowed the rate of decrease in streamflow. However, streamflow did not increase until about 11^h00^m when flow rate increased from the seep. Rainfall essentially ceased at 03^h30^m on December 23, but stream-

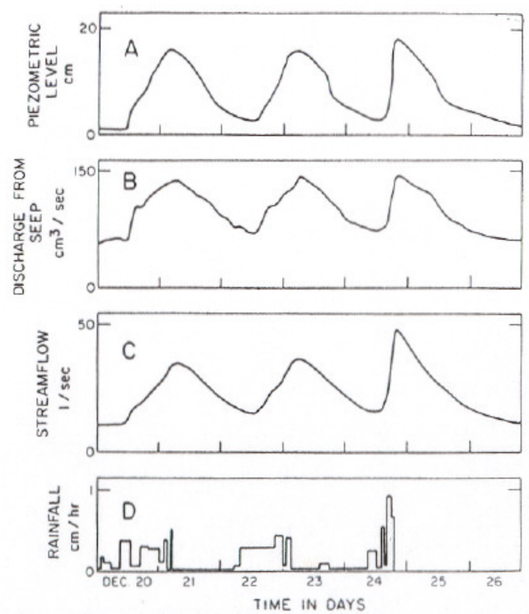

Fig. 7. Fluctuation of (A) piezometric level at a piezometer 1 m upslope from the metered seep, (B) discharge from the seep, (C) streamflow, and (D) rainfall from three storms.

flow didn't peak until $05^{h}00^{m}$; discharge from the seep didn't peak until about $07^{h}00^{m}$, reflecting continued drainage of the slope above this seep.

The water level in piezometer *35* (Fig. 7A) peaked some 4 h before flow from the seep. Analysis of data from piezometer *35* and other piezometers in the same grid suggests at least two sources of water for the seep downslope from this grid. One source appears to be shallow and close to the piezometer grid so that flow from this source peaks almost immediately after the end of rainfall as indicated by piezometric level in Fig. 7A. Water from the larger and probably more distant source continuous to enter the saturated zone upslope from the seep after rainfall has ceased, causing the discharge from the seep to peak several hours after the end of a storm.

Analysis of recession rates of the plots in Fig. 7 revealed additional similarities among piezometric level, discharge from the seep, and streamflow. In Fig. 8, values of each of these are plotted over time after the maximum value of each. Logarithmic scales are used for both axes of all plots because the logarithm of unsaturated drainage tends to be linearly related to the logarithm of time (cf. Wilcox, 1959; Nixon and Lawless, 1960; Hewlett and Hibbert, 1963; Weyman, 1970). Overland flow was never observed in the

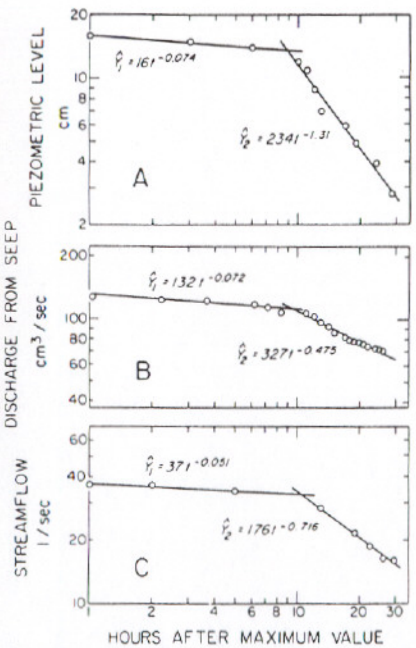

Fig. 8 Relationship of (A) piezometric level at a piezometer 1 m upslope from the metered seep, (B) discharge from the seep and (C) streamflow with time after maximum value of each for the storm of December 22, 1973.

watershed; subsurface flow predominated and ultimately controlled the rate of delivery of water to the saturated zone at the bottom of the study slope and other slopes in the watershed. Consequently, streamflow was largely a function of the rate of unsaturated drainage of sloping soil masses in the watershed, so that linear relationships between log of time and log of streamflow were expected.

In each plot in Fig. 8 there are two separate periods during which linear relationships exist between log of the Y variable and log of time. During each period, the Y variable is related to time by:

$$Y = at^{-b} \tag{5}$$

where Y = piezometric level, discharge from the seep, or streamflow; t = time after maximum value of piezometric level, discharge from the seep,or streamflow; and a and b are positive constants peculiar to each drainage period for each Y variable. The two periods are separated by a change in slope at t = 10 h. Differentiating eq. 5 for the two drainage periods in Fig. 8 shows the rate of decrease in piezometric level, discharge from the subsurface channel, and that streamflow increased about 10 h after the peak value of each. Streamflow plots are similar to plots of discharge from the seep; this suggests that the remainder of watershed *10* responded to storms in much the same way as did the study slope. Most likely, streamflow is the sum of a series of components of flow similar to that from the seep (Fig. 7B).

The forms of the curves in Fig. 8 are similar in some respects to those reported from two other studies of discharge of subsurface water from sloping soil masses. Hewlett and Hibbert (1963) described a two-phase process of drainage from a repacked, sloping soil mass 1 m deep. During the initial 1.5 days, drainage of large pores occurred; and after a 5-day transition period, drainage was from small pores. Weyman (1970, 1973) found that a 1-day first stage of drainage from a natural slope was dominated by saturated flow through primarily noncapillary pores. After a 4-day transition period, drainage from the soil was dominated by unsaturated lateral flow in small pores to a narrow saturated zone at the base of the slope. In both these studies, log-log plots of discharge from the slope over elapsed time showed that slopes of the discharge curves during the transition periods were steeper than for the first phase of drainage. The curves of the second phase of drainage were less steep than for the transition period but steeper than for the first phase of drainage.

In the study area, moderate-size winter storms associated with fronts moving in a wavelike fashion from the Pacific Ocean were spaced 24—48 h apart during the initial part of the study period. Therefore, in the examples shown in Fig. 8, piezometric level, discharge from the seep, and streamflow could recede for only about 30 h before beginning to respond to the next storm (Fig. 7). A third phase of drainage was observed, however, during the latter part of a virtually rainless 18-day period following the 7-cm storm on December 28—30. At $t \geqslant 60$ h, plots of both piezometric level and discharge from the seep showed another increasing rate of decrease. At $t = 62$ h, piezometer *35* emptied although the seep discharged water continuously at a very low rate. Other piezometers showed that a shallow saturated zone still existed, presumably because of drainage of water from the more extensive and remote source described earlier. At $t = 90$ h, streamflow decrease finally showed a similar increase in rate. Adding this third phase of drainage described above made the pattern of drainage of the study slope similar to that of the B horizon described in Weyman's (1970) study. Most of the discharge from Weyman's slope occurred from the B horizon while unsaturated percolation supplied water to a rapidly expanding zone of saturation.

In any event, drainage reported here appears to be much faster than that described by either Hewlett and Hibbert (1963) or Weyman (1970, 1973). Presumably, differences in pore-size distributions and in slope account for these differences in drainage rates.

Because unsaturated flow controls rate of water supply to the base of the slope, dynamics of water flux in the vicinity of the saturated zones as well as elsewhere on the slope should account for the change in rate of decreasing drainage at $t = 10$ h. This does appear to be the case. During the first stage of drainage, i.e., up to about 10 h, water moved out of the larger pores. This drainage can be illustrated by inserting transient values of $P_c/\rho g$ from Fig. 4 into the equation for capillary rise which is used to relate size of pores filled with water to $P_c/\rho g$. This analysis, ignoring the effects of hysteresis, indicates that pores with diameters <0.25 mm in the lower soil (70-cm depth) were water filled when $P_c/\rho g = -12$ cm at the start of the storm on December 22. At this time and depth, water flux was about 0.5 mm/h (Fig. 6B). As rainfall continued, larger pores began filling with water until pores with diameters up to about 1.5 mm (as indicated by $P_c/\rho g = -2$ cm) became filled. During this period, which lasted 10–12 hr, water flux at 70 cm was at a maximum of 3–4.5 mm/h. When rainfall ceased at 03^h30^m on December 23, these larger pores began draining as water was no longer being supplied from above by rainfall. As soil water flux decreased, flux to the lower slope decreased but initially at a relatively low rate. Within about 10 h after the end of rainfall, $P_c/\rho g = -11$ cm, which was nearly equal to pre-storm $P_c/\rho g$, and nearly all larger pores that had filled during the storm had drained. At this point, water flux had decreased to about 0.7 mm/h. Decreasing water flux to the lower slope was reflected by an increasing rate of decrease in discharge from the metered seep and in streamflow as the saturated zones contracted. After this 10-h period, soil continued to drain slowly for an additional 20 h during which time water flux gradually decreased to less than 0.5 mm/h. Then the soil began to respond to a new storm, and the cycle was repeated. Similar sequences of pores filling and draining can be observed for other soil depths although sizes of pores differed somewhat among depths.

pore filling & drainage related to Fig. 8

During the 18-day rainless period described earlier, the soil undoubtedly drained considerably more than that described above. But, because temperatures during this period were subfreezing, tensiometers could not be used to measure capillary pressures. Also, because of the severe freezing weather, precipitation and soil-water fluxes could not be determined for the study period's maximum storm event which occurred immediately after this period of freezing.

Contributions to storm flow

As stated before, overland flow has not been observed in watershed *10*. Subsequently, streamflow has only two components: (1) channel interception, i.e., rain falling directly onto the water and rock surfaces between streambanks; and (2) subsurface flow.

The absence of overland flow can be explained by the values of unsaturated hydraulic conductivity $K(P_c)$ for the top 70 cm of the soil mantle (Table II) and low rainfall intensities. The highly porous, well-aggregated surface soils can readily accept all precipitation and easily transmit water to lower soil depths. When pressure heads were least prior to a storm, values of $K(P_c)$ throughout the soil were 0.01–0.06 cm/h. As water was added to any depth, $K(P_c)$ at that depth increased as pressure head increased and as water moved into larger pores. During storms, soil at the 70-cm depth apparently came closest to saturation. Before a storm, saturation here was 75%, increasing to 90% during storms. Similarly, before storms, saturations at 10- and 30-cm depths, respectively, were 67 and 55%, both increasing to 75% during storms. Considerable storage was available during storms for additional water had it occurred. However, the wetter the soil became, the better able it would have been to transmit water, and up to a point, overland flow perhaps would have been no more likely to have occurred than it was during storms in this study.

Hewlett and Hibbert (1967) have described a concept of rapid watershed response to precipitation without appreciable overland flow. They discriminate between two types of streamflow: quick flow and delayed flow. An arbitrary separation line with a constant slope of 0.55 l sec^{-1} km^{-2} h^{-1} separates quick flow from delayed flow. Each unit of rainfall on a slope contributions to both temporary storage, which eventually becomes delayed flow during some later period, and quick flow. Upslope from a stream, a unit of rainfall contributes less to quick flow and more to retention–detention storage. Hewlett and Hibbert further theorize that only part of the water contributed to quick flow is actually new water, i.e., water from the current storm; the remainder is called "translatory flow" or flow by displacement. This is water already stored in the soil prior to the storm and will be released in large quantities only where the soil is extremely wet. They regard this movement as the result of thickening of water films around soil particles and a resulting pulse of water as the saturated zone is approached. This translatory process has been supported by laboratory studies (Horton and Hawkins, 1965) in which 2.5 cm of tritiated water was added to 1.2-m soil drained to equilibrium. Then 2.5 cm of pure water was added each day, and effluent was analyzed daily for tritium. Only after 87% of the original water flowed from the columns did the tritiated water emerge. Horton and Hawkins concluded that in deep soil profiles rainwater likewise tends to displace water as it moves primarily through capillary pores. More recently, Martinec (1975), using environmental tritium to trace the movement of snowmelt water, concluded that infiltration of meltwater causes a corresponding increase in outflow of subsurface water to streams. Thus, infiltrated water, although moving slowly, appears "in substitution" in the hydrograph, represented by water previously stored in the soil mantle.

Table III shows relative contributions of channel interception and subsurface storm flow to total storm flow for seven storm events in the study reported here. Quick flow here is synonymous with that defined by Hewlett and Hibbert (1967) and is composed of channel interception and subsurface storm flow. This subsurface storm flow is roughly synonymous with that described by van 't Woudt (1954), Whipkey (1965, 1969) and Freeze (1972b). Strict comparisons with subsurface storm flow as described by these authors or with "interflow" (Chamberlin, 1972) and "throughflow" (Kirkby and Chorley, 1967; Weyman, 1973) would, or course, require uniformity of methods used to determine duration of storm flow and to separate quick flow from delayed flow.

A relatively high proportion of gross precipitation appeared as quick flow. Quick flow averaged 38% of gross precipitation, ranging from 23 to 51% for the storms analyzed. Proportion of quick flow contributed by channel interception was estimated from measurements of stream channel length and width during the rainy season combined with storm precipitation after the time streamflow began to increase. As can be seen in Table III, channel interception was only 2–4% of quick flow. (The channel interception which served to slow streamflow recession as described earlier is not included here.) The remaining 96–98% was contributed by subsurface storm flow. Thus, subsurface streamflow is by far the major contributor to storm runoff in this watershed. Most likely this is also true for most other headwater basins in the western Cascade Range of Oregon.

If the ratio of storm flow to gross precipitation can be taken as a rough estimate of the percentage of a watershed contributing to storm flow (Hewlett, 1974), then, according to Table III, on the average about 38% of watershed *10* contributed to storm flow during the study period. If watershed *10* were idealized somewhat (total channel length = 550 m and slope gradient for the lower half of all slopes = 75%), a storm flow-contributing area equal to 38% of the watershed corresponds to an area extending about 42 m upslope on either side of stream channels. Clearly this area would have to be 100% saturated for a given volume of rainfall to completely displace water stored in it. Judged by pressure and piezometric levels, however, these zones were at best only about 65% saturated. Thus, the area contributing to storm flow must have extended, on the average, more than 42 m upslope to allow for incomplete displacement of soil water stored in the contributing area of the watershed. The exact upslope extent of this contributing area remains unknown, however. Most likely it is highly irregular owing to the presence of the discontinuous saturated zones on lower slopes.

Quick flow appears to be a larger percentage of storm precipitation in headwater basins in the western Cascade Range in Oregon than in other areas for which similar analyses have been reported. In the eastern United States, only the exceptional storm produces quick flow in excess of 25% of gross precipiation and the largest flows have rarely exceeded 50% of the rainfall that produced them (Hewlett and Hibbert, 1967). In contrast, quick flow from watershed *10* in Oregon averaged 38% for seven winter storms that are low to moderate size at best; quick flow from a 17.1-cm storm of 5-day duration amounted to 51% of rainfall. Presumably, during the largest storm events during which 24-h rainfall has exceeded 16 cm (Rothacher et al., 1967) and more than 32 cm of rain has fallen in 3 days, even a greater proportion of the watershed can contribute to storm flow.

Results of storm flow analyses are in agreement with the concept of variable source area of runoff production (cf. Hewlett, 1974). Piezometer data show that saturated zones whose outlets are visible at the toe of the study slope expand gradually upslope and laterally as rainfall continues. However, overland flow on wet zones near streams, which can be an important factor in runoff production in other areas (Ragan, 1968; Betson and Marius, 1969; Dunne and Black, 1970a, b), was notably absent and has not appeared in similar small watersheds in this region. Stream channels apparently do not expand or contract appreciably over the course of frequent

TABLE III

Relative magnitude of storm flow, channel interception, and subsurface storm flow for watershed *10* during seven storm events

Date	Gross precipitation (cm)	Quick flow as percent of gross precipitation	Channel interception as percent of quick flow	Subsurface storm flow as percent of quick flow
Dec. 19–21	7.0	34	3	97
Dec. 22/23	6.7	23	3	97
Dec. 24	4.2	40	2	98
Dec. 26–30*	18.3	38	3	97
Jan. 30–Feb. 1	10.7	42	2	97
Feb. 3/4	4.3	38	4	96
Feb. 15–20	17.8	51	2	98
	Mean	38	3	97

*Complex storm and hydrograph occurred during this period.

winter st...s. That soils have high conductivities, slopes are steep, and slopes are slightly convex is very conducive to subsurface storm flow and precludes overland flow on any segment of a slope.

SUMMARY

(1) Both magnitude and direction of water flux in soil and subsoil varied temporally. Maximum fluxes of 3–4 mm/h in the top meter of the soil mantle were 6–9 times greater than those in the subsoil. Water flux below 30 cm was directed mostly downslope during storms and more vertically downward between storms. Water flux in surface soils was directed downslope between storms and more vertically downward during storms.

(2) A gradual change in pore-size distribution throughout the soil and an abrupt change at the soil—subsoil boundary influenced values of relative hydraulic conductivity estimated from measurements of capillary pressures. Because of steep slopes, the gravitational component of hydraulic head overshadowed the small changes in pressure head measured during the course of winter storms. However, these slight changes in pressure head effected large changes in both magnitude and direction of water flux at various depths.

(3) Unsaturated flow dominated over all but the bottom 12–15 m of the study slope. Temporary saturated zones frequently occurred at the soil–subsoil boundary at mid- to upslope locations. Potential fluxes in these saturated zones were high if these zones were connected to the permanently saturated zones at the base of the slope. However, these upslope saturated zones appeared to be isolated, and flux in them was limited by the unsaturated conductivity of subsoil immediately downslope.

(4) Saturated flow conditions at the bottom of the slopes in the watershed appeared to be related to the drainage of soil pores which filled with water during a storm. About 10 h after rainfall had ceased, these pores had essentially drained. Discharge from a seep and streamflow exhibited a marked increase in their rates of decrease at this time.

(5) Subsurface storm flow dominated storm flow from the study watershed. Overland flow did not occur. Total storm flow averaged 38% of storm precipitation. Subsurface storm flow and channel interception, respectively, averaged 97 and 3% of total storm flow.

ACKNOWLEDGMENTS

Winter fieldwork by John S. Fekete, computer programing by Tom C. Corbett, and helpful comments by Royal H. Brooks are gratefully acknowledged. This study was partially funded by National Science Foundation Grant No. GB-38610X to the Coniferous Forest Biome, U.S. Analysis of Ecosystems, International Biological Program while the author was Assistant Professor of Forest Hydrology, School of Forestry, Oregon State University, Corvallis. This is Contribution No. 178 from the Coniferous Forest Biome.

REFERENCES

Betson, R.P. and Marius, J.B., 1969. Source areas of storm runoff. Water Resour. Res., 5: 574--582.

Blake, G.R., 1965. Bulk density. In: C.A. Black (Editor), Methods of Soil Analysis, Am. Soc. Agron. Madison, Wisc., pp. 374–390.

Brooks, R.H. and Corey, A.T., 1966. Properties of porous media affecting fluid flow. J. Irrig. Drain. Proc. ASCE, 4855 (IR2): 61–88.

Brown, G.W., Burgy, R.H., Harr, R.D. and Riley, J.P., 1972. Hydrologic modeling in the Coniferous Forest Biome. In: J.F. Franklin, L.J. Dempster and R.H. Waring (Editors), Proceedings—Research on Coniferous Forest Ecosystems—A Symposium, Pac. Northwest For. Range Exp. Stn., For. Serv. U.S. Dep. Agric., Portland, Oreg., pp. 49–70.

Chamberlin, T.W., 1972. Interflow in the mountainous forest soils of coastal British Columbia. In: O. Slaymaker and H.J. McPherson (Editors), Mountain Geomorphology, Tantalus Res., Vancouver, B.C., pp. 121–127.

Day, P.R., 1965. Particle fractionation and particle-size analysis. In: C.A. Black (Editor), Methods of Soil Analysis, Am. Soc. Agron., Madison, Wisc., pp. 545–567.

Dunne, T. and Black, R.D., 1970a. An experimental investigation of runoff production in permeable soils. Water Resour. Res., 6: 478–490.

Dunne, T. and Black, R.D., 1970b. Partial area contributions to storm runoff in a small New England watershed. Water Resour. Res., 6: 1296–1311.

Dyrness, C.T., 1969. Hydrologic properties of soils on three small watersheds in the western Cascades of Oregon. USDA For. Serv., Res. Note PNW-111, Pac. Northwest For. Range Exp. Stn., Portland, Oreg., 17 pp.

Franklin, J.F. and Dyrness, C.T., 1969. Vegetation of Oregon and Washington. USDA For Serv. Res. Pap. PNW-80, Pac. Northwest For. Range Exp. Stn., Portland, Oreg., 216 pp.

Freeze, R.A., 1972a. Role of subsurface flow in generating surface runoff, 1. Base flow contributions to channel flow. Water Resour. Res., 8: 609–623.

Freeze, R.A., 1972b. Role of subsurface flow in generating surface runoff, 2. Upstream source areas. Water Resour. Res., 8: 1272–1283.

Haupt, H.F., 1969. A simple snowmelt lysimeter. Water Resour. Res., 5: 714–718.

Hewlett, J.D., 1974. Comments on letters relating to *Role of subsurface flow in generating surface runoff, 2. Upstream source areas* by R. Allan Freeze. Water Resour. Res., 10: 605–607.

Hewlett, J.D. and Hibbert, A.R., 1963. Moisture and energy conditions within a sloping soil mass during drainage. J. Geophys. Res., 68: 1081–1087.

Hewlett, J.D. and Hibbert, A.R., 1967. Factors affecting the response of small watersheds to precipitation in humid areas. In: W.E. Sopper and H.W. Lull (Editors), Forest Hydrology, Pergamon, New York, N.Y., pp. 275–290.

Hewlett, J.D. and Troendle, C.A., 1975. Non-point and diffused water sources: a variable source area problem. Watershed Management, Am. Soc. Civ. Eng., New York, N.Y., pp. 21–46.

Holbo, H.R., Harr, R.D. and Hyde, J.D., 1975. A multiple-well, water-level measuring and recording system. J. Hydrol., 27: 199–206.

Horton, J.H. and Hawkins, R.H., 1965. Flow path of rain from the soil surface to the water table. Soil Sci., 100: 377–383.

Jones, A., 1971. Soil piping and stream channel initiation. Water Resour. Res., 7: 602–610.

Kirkby, M.J. and Chorley, R.J., 1967. Throughflow, overland flow, and erosion. Int. Assoc. Sci. Hydrol. Bull., 12: 5–21.

Martinec, J., 1975. Subsurface flow from snowmelt traced by tritium. Water Resour. Res., 11: 496–498.

Megahan, W.F., 1972. Subsurface flow interception by a logging road in mountains of central Idaho. In: Proc. Nat. Symp. on Watersheds in Transition. Am. Water Resour. Assoc., Urbana, Ill., pp. 350—356.
Minshall, N.E. and Jamison, V.C., 1965. Interflow in claypan soils. Water Resour. Res. 1: 381—390.
Nixon, P.R. and Lawless, G.P., 1960. Translocation of moisture with time in unsaturated soil profiles. J. Geophys. Res., 65: 665—661.
Patric, J.H. and Swanston, D.N., 1968. Hydrology of a slide-prone glacial till soil in southeastern Alaska. J. For., 66: 62—66.
Ragan, R.M., 1968. An experimental investigation of partial-area contributions. Int. Assoc. Sci. Hydrol. Publ., 76: 241—249.
Ranken, D.W., 1974. Hydrologic properties of soil on a steep forested slope. M. Sc. Thesis, Oregon State University, Corvallis, Oreg. (unpublished).
Rawitz, E., Engman, E.T. and Cline, G.D., 1970. Use of the mass balance method for examining the role of soils in controlling watershed performance. Water Resour. Res., 6: 1115—1123.
Rogowski, A.S., Engman, E.T., and Jacoby, Jr., E.L., 1974. Transient response of a layered, sloping soil to natural rainfall in the presence of a shallow water table: experimental results. Agric. Res. Serv. Res. Pap. ARS-NE-30.
Rothacher, J., Dyrness, C.T. and Fredriksen, R.L., 1967. Hydrologic and Related Characteristics of Three Small Watersheds in the Oregon Cascades. Pac. Northwest For. Range Exp. Stn., For. Serv., U.S. Dep. Agric., Portland, Oreg., 54 pp.
Stephenson, G.R. and Freeze, R.A., 1974. Mathematical simulation of subsurface flow contributions to snowmelt runoff, Reynolds Creek watershed, Idaho. Water Resour. Res., 10: 284—294.
Swanson, F.J. and James, M.E., 1975. Geology and geomorphology of the H.J. Andrews Experimental Forest, western Cascades, Oregon. USDA For. Serv. Res. Pap. PNW-188, Pac. Northwest For. Range Exp. Stn. Portland, Oreg., 14 pp.
Swanston, D.N., 1967. Soil-water piezometry in a southeast Alaska landslide area. USDA For. Serv. Res. Note, PNW-68, Pac. Northwest For. Range Exp. Stn., Portland, Oreg., 17 pp.
Takeda, S. and Ishii, M., 1968. Movement of subsurface flow from forest area and mechanism that affects soil erosion. Mt. Land Res. Lab., Bull., No. 1, Iwate Univ. Morioka.
U.S. Forest Service, 1961. Watershed management: Some ideas about storm runoff and base flow. Annu. Rep., Southeastern For. Exp. Stn., Asheville, N.C., pp. 61—66.
Van 't Woudt, B.D., 1954. On factors governing the subsurface storm flow in volcanic ash soils, New Zealand. Trans., Amer. Geophys. Union, 35: 136—144.
Vomocil, J.A., 1965. Porosity. In: C.A. Black (Editor), Methods of Soil Analysis, Am. Soc. Agron., Madison, Wisc., pp. 299—314.
Weyman, D.R., 1970. Throughflow on hillslopes and its relation to the stream hydrograph. Int. Assoc. Sci. Hydrol. Bull., 15: 25—33.
Weyman, D.R., 1973. Measurements of the downslope flow of water in a soil. J. Hydrol., 20: 267—288.
Whipkey, R.Z., 1965. Subsurface stormflow from forested slopes. Int. Assoc. Sci. Hydrol. Bull., 10: 74—85.
Whipkey, R.Z., 1969. Storm runoff from forested catchments by subsurface routes. Int. Assoc. Sci. Hydrol. Publ. 85: 773—779.
Wilcox, J.C., 1959. Rate of soil drainage following an irrigation,I. Nature of soil drainage curves. Can. J. Soil. Sci., 39: 107—119.
Wilson, T.V. and Ligon, J.T., 1973. The interflow process on sloping watershed areas. Water Resour. Res. Inst. Rep., No. 38, Clemson Univ., Clemson, S.C.
Zaslavsky, D. and Rogowski, A.S., 1969. Hydrologic and morphologic implications of anisotropy and infiltration in soil profile development. Soil Sci. Soc. Am. Proc., 33: 594—599.

Journal of Hydrology, 33 (1977) 59—72

FINITE-ELEMENT ANALYSIS OF GROUNDWATER FLOW IN MULTI-AQUIFER SYSTEMS, I. THE BEHAVIOR OF HYDROLOGICAL PROPERTIES IN AN AQUITARD WHILE BEING PUMPED

K. FUJINAWA

Department of Agricultural Engineering, Kyoto University, Kyoto (Japan)

(Received March 29, 1976; accepted for publication August 9, 1976)

ABSTRACT

Fujinawa, K., 1977. Finite-element analysis of groundwater flow in multiaquifer systems, I. The behavior of hydrological properties in an aquitard while being pumped. J. Hydrol., 33: 59—72.

In multiaquifer systems, aquifers are separated by aquitards through which water flows from one aquifer to another. The assumption of vertical flow in aquitards and horizontal flow in aquifers has been found reasonable in many field situations and has been utilized in the design of quasi three-dimensional flow models. It has also been revealed by earlier workers that the vertical leakage is affected by the storage in the aquitard as well as by its permeability and the properties of the adjacent aquifers. In this paper one-dimensional finite-element analysis is used to determine the effect of storage in an aquitard separating two confined aquifers on the distribution of hydraulic head in the aquitard and the flux from or into the adjacent aquifer. This paper also considers how many nodes are required to solve the problem numerically with adequate accuracy. The results of this study should be helpful in the design of numerical schemes for analysing multiaquifer systems.

INTRODUCTION

Since it has been recognized that flow through aquitards has a significant effect on the head decline of pumped or unpumped aquifers, multiaquifer groundwater flow has been analysed as one of the most important problems. In 1969, general analytical solutions to the flow in multiaquifer systems were obtained by Neuman and Witherspoon (1969a, b). Their theory incorporated for the first time both storage in aquitards and drawdown of the adjoining aquifers caused by leakage through the separating aquitards. If one wants to construct a quasi three-dimensional model of multiaquifer systems, the leakage through aquitards may be incorporated into the areal flow in one of two ways. One method is to use the analytical solutions introduced by Bredehoeft and Pinder (1970), and the other is to use numerical solutions. Bredehoeft and Pinder analysed the areal flow by coupling a finite-difference analysis of the horizontal flow in aquifers and an analytical analysis of the