MEASUREMENTS OF THE DOWNSLOPE FLOW OF WATER IN A SOIL

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ABSTRACT


Measurements have been made of unsaturated and saturated lateral soil water flow on a convex hillslope with a good soil cover and impermeable bedrock during natural rainstorms. The hydraulics of flow are examined in detail with particular reference to the role of breaks in vertical permeability, the change from saturated to unsaturated flow and the velocity of flow. In this instance, after rainfall slope flow is dominated by vertical unsaturated movement towards the profile base. Preceding upslope moisture gradients result in the growth of a zone of soil saturation upwards from the slope base: Slope discharge through the B and B/C horizons, is related to the form of the saturation zone, within which flow is lateral, according to Darcy's law. The time required for vertical percolation and the low hydraulic conductivity of the lower soil horizons result in a hillslope hydrograph which is delayed and attenuated and cannot be regarded as stormflow. During drainage the saturation zone contracts and is replaced by a lateral unsaturated flow system at the profile base which supplies discharge from the B/C horizon for up to 42 days without further recharge. It is concluded that, in general, either distinct soil horizons or impermeable bedrock are essential for the initiation of lateral flow. Saturated flow is likely to dominate hillslope hydrographs through non-capillary pore spaces but these may be integrated to the point where Darcy's law still holds. Although lateral soil water flow must be a widespread phenomenon, it is unlikely to provide storm runoff to the stream unless saturated conditions are generated within the organic horizons or flow within the lower soil horizons is dominated by non-Darcian flow through non-capillary spaces in the soil.

INTRODUCTION

Although there is now an extensive literature on the topic of lateral soil water movement in a downslope direction, only a few experiments have yet been conducted under natural conditions. Moreover, experimental results have indicated a confusing range of soil requirements, hydraulic conditions and flow velocities for lateral soil water movement. The experiment described here is a further attempt to resolve some of the existing contradictions and particularly the following:

1) Does lateral flow occur within any soil, or are distinct soil horizons or impermeable bedrock necessary for its initiation?

2) Does lateral flow occur within both saturated and unsaturated soil and, if so, does one soil state dominate the hillslope hydrograph?

3) What effect do non-capillary pore spaces have upon flow and is part of the flow system hydraulically distinct from the soil matrix?
(4) Where lateral flow occurs, does it contribute to the storm hydrograph of the drainage basin, or does it merely supply or augment stream baseflow?

The early literature on the problem tended to view lateral soil water flow as either exclusively a baseflow or a stormflow phenomenon. Roessel (1951) found a poor correlation between groundwater levels and stream baseflow in headwater areas and attributed the maintenance of long-term flows to "soil water". Hewlett (1961) and Hewlett and Hibbert (1963) built an artificial slope of homogeneous soil and recorded discharge from the slope base for a period of 71 days after initial saturation. They concluded that unsaturated "soil water seepage" was able to sustain baseflows in steep upland basins with a good soil cover. In contrast, rapid soil water movement was first described by Van't Woudt (1954) who used open-sided lysimeters to measure "stormflow" in volcanic ash soils. Whipkey (1965) recorded complete storm hydrographs for subsurface flow from a natural forested slope following simulated rainstorms. Part of the flow was taking place in a saturated state above horizons of decreased soil permeability. Similar situations have been described for highland Britain by Knapp (1970) and others.

Recognition of these subsurface processes led to a re-evaluation of the Horton (1945) model of overland flow generated by rainfall intensities in excess of the soil infiltration capacity. Under some conditions lateral soil water flow might provide basin storm runoff. Alternatively, slow soil water seepage might maintain saturated soil conditions at the slope base (Betson, 1964; Tennessee Valley Authority, 1965) or in areas of thin A horizons (Betson and Marius, 1969) giving rise to "saturated overland flow" during a storm. Subsequent work revealed a range of surface and subsurface processes generating stormflow. Ragan (1968) measured stormflow from a saturated litter horizon but found no substantial lateral water movement within the mineral layers of the soil. Dunne and Black (1970a and b) found that hillslope storm runoff was dominated by saturated overland flow in concave sectors although subsurface flow was also recorded. Moreover, they stressed the importance of channel precipitation and related processes, as against hillslope runoff, in the generation of the basin storm hydrograph. Both Ragan and Dunne and Black were working in areas where horizontated soils overlay glacial deposits: the presence of breaks in vertical permeability may therefore be an important element in both the subsurface stormflow and saturated overland flow models. For areas where lateral flow does occur, there remains the possibility that water may move rapidly through the profile along waterworn pipes (Jones, 1971) or through organically-created non-capillary routes (Chamberlin, 1972) without reference to the hydraulic system of the main soil mass.

The results of the present experiment refer, inevitably, to another particular site. In this instance, however, lateral soil water flow, operating across a range
of velocities, formed an important element of basin response to rainfall during the hydrograph “recession-limb” and at baseflow. Moreover, although the process did not generate stormflow under the observed conditions, an extrapolation to subsurface stormflow and saturated overland flow can be envisaged. The term “throughflow”, after Kirkby and Chorley (1967), has been applied to the lateral downslope movement of water to escape the “stormflow/baseflow” connotation of earlier work. It was also found possible to distinguish between dominantly saturated and unsaturated flow and between non-capillary and capillary flow within the hillslope hydrograph whilst maintaining a continuity of process. Consideration of the pattern of water movement within the soil mantle has also made it possible to offer some indication of the relative importance of soil horizons and the nature of bedrock in the initiation of throughflow.

All empirical regression equations quoted in this paper have been tested with Pearson’s product-moment correlation coefficient and are significant at \( \alpha = 0.05 \).

MEASUREMENT SITE AND METHODS

The experimental hillslope is located in the basin of the East Twin Brook, Burrington, Somerset. The slope is 670 m long, convex in profile and covers an elevation range of 50 m. Slope angle varies from 23° at the base, through a 15° section (7–50 m upslope) and then declines rapidly to a long 2° slope towards the divide. The slope is underlain by shales and sandstones of Old Red Sandstone (Devonian) age, rendered virtually impermeable by secondary silification (Whittard, 1949). The soil developed over bedrock is a free-draining brown earth mapped by Findlay (1965) as Maesbury Series. Four soil horizons were identified by Findlay:

- **A horizon** 0–10 cm below surface: an acid mull
- **A/B horizon** 10–25 cm below surface: a silty loam with more than 10% organic matter
- **B horizon** 25–40 cm below surface: a stony loam with sub-angular blocky structure
- **B/C horizon** 40–60 cm below surface: a very stony loam merging into fractured bedrock

Over the bottom 3 m of slope the B and B/C horizons thicken rapidly reaching 25–50 and 50–75 cm below surface, respectively, at the natural soil face at the slope base. The brown earth supports a dense growth of bracken (*Pteridium aquilinum*) with subordinate bramble (*Rubus* sp.) and a thin mat of grasses (mainly *Festuca ovina*).
The experimental slope was arbitrarily defined as 1.0 m in width at the slope base. The hillslope is broadly convex in plan and the width of the experimental plot consequently decreases upslope. Beyond about 350 m upslope contour convexity increases and the plot becomes very narrow.

Hillslope discharge was measured at the slope base using the system described by Whipkey (1965). Four lateral troughs, each 1 m long, were inserted into the natural exposed soil face at the slope base. Each trough took flow exclusively from one of the soil horizons described above. Discharge measurement was made by hand using a stop-watch and measuring cylinder.

An assemblage of instruments was located upslope of the measuring troughs to ascertain soil water tension. Vacuum-gauge tensiometers, mercury-manometer tensiometers (Webster, 1966), and Bouyoucos gypsum blocks were employed to give tension readings for each horizon at a number of points upslope. In practice, a high rate of instrument failure meant that consistent results were obtained only for the mercury-manometer tensiometers, representing tension readings for the A/B and B horizons only and four upslope locations (1.0, 3.5, 6.0 and 12.0 m). The results discussed here are based largely upon those instruments but it is felt that the dominant patterns of flow on the slope are realistically represented by them. In particular, it was found that the bottom few metres of slope completely dominated the runoff situation. When the importance of saturated, rather than unsaturated, flow became clear a number of piezometers were installed in the B and B/C horizons over the same length of slope.

Measurement of slope discharge and soil water tension were made at regular intervals for 18 months during 1969 and 1970. As far as was practicable continuous observation was made during a number of storms. Results from permanent instrumentation were supplemented with simple infiltration tests, bulk density determinations and laboratory calibration of tensiometers.

RESULTS

Results are presented here in roughly the order of actual discovery. Starting with the observable expression of throughflow — hillslope discharge — an attempt is made to examine the hydraulics of saturated flow which governs instantaneous discharge and then the hydraulics of unsaturated supply from further upslope is examined. Only after the major features of flow and discharge have been described is an attempt made to combine observed patterns into a general statement of the sequence of events during the course of a storm.

Hillslope discharge

During the period of investigation no overland flow was observed on the ex-
the experimental hillslope. Infiltration tests suggested that final infiltration capacity of the soil was still an order of magnitude greater than the maximum rainfall intensity recorded during the experiment (20 mm/h) and several times higher than the maximum intensity ever recorded in the area (70 mm/h) (Hanwell and Newson, 1970). Similarly, the soil water table (see below) never came close enough to the ground surface to generate saturated overland flow. All runoff from the hillslope was in the form of throughflow within the mineral layers of the soil. Moreover, discharge was entirely confined to the B and B/C horizons of the soil, the higher horizons producing no discharge.

Discharge from the B/C horizon was continuous throughout the entire experiment which included several prolonged periods with no rainfall. Maximum B/C horizon discharge recorded was only 10 cm³/min maxima occurring between 12 and 80 h after rainfall. Discharge from the B horizon was intermittent, occurring only after substantial rainstorms and forming a very attenuated hydrograph. Peak B horizon discharge (up to 430 cm³/min) occurred a minimum of 12 h after peak rainfall (B/C horizon peak is coincidental with or later than B horizon peak). Discharge from the B horizon ceased within 100 h of peak discharge. The general pattern of hillslope runoff, in response to a complex winter storm of 70.6 mm, is shown in Fig. 1. By analogy with the stream hydrograph (Weyman, 1971; 1973) the B/C horizon was shown to provide basin baseflow and B horizon discharge represented the main basin response to rainfall (but cannot, of course, be regarded as stormflow since the time to peak discharge is so long).

The relationships between rainfall, antecedent moisture conditions and runoff volume, peak discharge magnitude and timing are discussed in detail elsewhere (Weyman, 1973). This paper will therefore concentrate upon the relationships of system components within a storm rather than between storms.

Saturated flow and slope discharge

Water leaves the soil mantle through a zone of permanent soil saturation at the slope base. During the course of a storm this zone of saturation was observed to grow upslope and up the soil profile in the form of an expanding wedge. Water may move into this saturated zone from surrounding unsaturated soil across a continuous pressure gradient. At any moment, however, discharge from the slope base should be related to the form of the saturated wedge within which water moves according to Darcy’s law (Childs, 1969):

\[ v = -K \cdot \text{grad} \phi \]

where \( v \) is the velocity of flow in the direction of the maximum pressure gradient; \( K \) is the saturated hydraulic conductivity of the medium; and \( \text{grad} \phi \) is the po-
potential gradient with movement of water from high to low potential. The hydraulic head ($\phi$) for water at any point in the system can be defined as:

$$\phi = z + H$$

where $z$ is the height of the point above an arbitrary datum (in this case the slope base); and $H$ is the height above $z$ to which water will stand in a piezometer tube.

In the case of discharge from beneath a free water table, the hydrostatic pressure ($H$) at the water table will be zero (atmospheric pressure) and the potential gradient of flow will be equal to the slope of the water surface:

$$-\text{grad } \phi = -\frac{dz}{dl} = -\sin \theta$$

where $l$ is the distance along the water table; and $\theta$ is the angle of the watertable.
For an experimental slope situation, with width $w$ and a saturated zone of thickness $h$, total discharge ($Q$) in the direction of maximum potential may be expressed as:

$$Q = K \cdot w \cdot h \cdot \sin \theta$$

assuming the soil to be isotropic and of uniform hydraulic conductivity.

In the present instance, $h$ can be taken as the height of the top of the saturated zone above the profile base at the discharging face. Empirically it was found that:

$$q = 0.0003 h_c^{3.73}$$

where $q$ is total plot discharge in cm$^3$/min; and $h_c$ is the height of the water table above the profile base at piezometer c, 3 m upslope (see Fig. 2). (Saturation depth at 3 m upslope is probably a reasonable approximation of discharging face saturation depth – see below.)
The upslope extent of saturation was more difficult to determine than saturation depth due to piezometer spacing. Nevertheless, enough information was gathered to suggest that saturation depth and upslope extent are linearly related. In other words, the saturation zone tends to be of constant shape (Fig. 3). It is therefore possible to suggest that $\theta$ is constant for one point on the slope for a range of discharge. Plot width ($w$) can also be assumed constant for the section of slope under consideration (maximum upslope saturation observed was only 15 m). The simplest discharge model applicable to this situation therefore predicts that discharge should be a linear function of water table height. In fact, the relationship was shown to be non-linear. One source of this non-linearity is thought to be the expected increase in lateral hydraulic conductivity with height up the soil profile. Theoretically, the discharge equation could then be written as:

$$Q = w \cdot \sin \theta \int_{y=0}^{y=h} k \cdot dy$$

where $y$ is height above the profile base.
Now, if:

\[ K \propto y^m \]

\[ Q \propto h^{m+1} \]

Since \( Q \propto h^{3.73} \) empirically, then \( K \propto y^{2.73} \).

A very crude check can be made upon this conclusion. Direct measurement of bulk density through the profile indicates that bulk density increases with depth to 2.0 gm/cm³ at the profile base. Assuming that porosity (\( e \)) is close to zero at the profile base (suggested by the rapid fall-off in discharge as saturation depth decreases), it is possible to assume that porosity is inversely related to bulk density and increases up the soil profile according to:

\[ e \propto y^{-1.135} \] (derived from the bulk density profile)

Terzaghi and Peck (1948) indicate that hydraulic conductivity is a function of \( e^2 \), in which case:

\[ K \propto y^{2.26} \]

which might be compared with \( K \propto y^{3.73} \) derived above.

The unresolved non-linearity indicated by this comparison may be entirely spurious or may point to inaccuracies in the original derivation (for example, the use of piezometer \( c \) to represent discharging face water table height) or may be the result of true non-linearities elsewhere in the system. In particular, the isotropic assumption may not be met in as far as part of the saturated flow may be taking place within the capillary pore system and part within non-capillary structural voids. The short-term answer in this instance, however, is that the proven relationship between discharge and water-table height could only exist if any non-capillary system was integrated with the capillary system to give a true water table.

Using the relationship between discharge and saturation height at the discharging face, the mean velocity of saturated throughflow (capillary and non-capillary flow across the whole face) was found to vary from 0.01 cm/min (at \( q = 20 \text{ cm}^3/\text{min} \)) to 0.075 cm/min (at \( q = 300 \text{ cm}^3/\text{min} \)). Maximum velocity within the soil profile (non-capillary flow towards the top of the B horizon) will be considerably in excess of these mean figures. T.C. Atkinson (personal communication, 1971) ran a tracer test over the same section of slope using the fluorescent dye “Pyranine Conc.” and reported an average downslope velocity (over 6.5 m) of 0.3 cm/min on the basis of rainfall to dye recovery time. Because the slope is strongly convex in profile, saturation depth does not increase markedly downslope for a given discharge but water table slope (\( \theta \)) increases leading to downslope increase in velocity. Final saturated throughflow velocity is therefore probably in excess of 0.3 cm/min.
Unsaturated flow during drainage

The saturated zone is fed by water moving out of the unsaturated soil lying above and upslope of the saturated zone. Within that unsaturated soil water is held at less than atmospheric pressure (tension). Flow will still take place from areas of high potential to areas of low potential but hydraulic head ($\phi$) must now be re-defined as:

$$\phi = z - \psi$$

where $\psi$ is a suction, or tension, exerted by gravitational, capillary and osmotic forces (Remson and Randolph, 1962).

For practical purposes and omitting the osmotic forces, $\psi$ may be interpreted directly as the reading on a tensiometer cell located at point $z$. With this adaptation, Darcy's law may now be used for unsaturated flow if the isotropic and uniform hydraulic conductivity assumptions are still met and if $K$ is interpreted as unsaturated hydraulic conductivity.

The potential flow net for saturated flow can be constructed fairly simply once the free water table is located. For unsaturated flow, however, the potential field will be created in response to a number of interacting forces. In particular, gravitational forces tend to pull water downwards whereas capillary forces initiated by the evapotranspiration process are tending to drain the soil by movement towards the ground surface.

Considering gravity alone, for a constant moisture content ($\psi$ is a constant in a homogeneous soil if hysteresis is ignored), hydraulic head ($\phi$) will increase directly with height ($z$) above the outflow point leading to a downwards movement of water. This movement will result in a decrease in water content ($\psi$ increases) with height. The limit of the gravity drainage process should occur at the point where:

$$\psi = z$$

If this state were attained, in a downslope direction $\nabla \phi = 0$ and consequently $\nu = 0$. Water would cease to move under gravity. In practice, tension is likely to increase by gravity only to the point where the field capacity is reached (Buckman and Brady, 1969). If field capacity is reached before $\psi = z$, gravity drainage will cease.

Within these limits it has been suggested that an upslope tension gradient will develop (Hewlett and Hibbert, 1963), approximated by:

$$\psi = b z$$

where $b$ is a coefficient which increases over time as the slope drains to the gravity limit $b = 1$.

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Fig. 4. Theoretical considerations of unsaturated flow on the hillslope: A. Unattainable state where soil water tension increases uniformly with height leading to horizontal equipotential lines (\(\phi\)) and entirely vertical water movement. B. Lateral unsaturated flow throughout the profile can only be attained when equipotential lines are orthogonal to the slope. C. During drainage by gravity, one equipotential line (\(\phi_{eq}\)) may be expected to rotate from the near horizontal (at time \(t_1\)) towards the orthogonal (at time \(t_n\)). D. In the later stages of drainage by evapotranspiration water is removed from near-surface horizons until the potential gradient is reversed over the whole profile (at time \(t_n\)).

ormal soil where tension (\(\psi\)) is a function of variations in both porosity and moisture content with height (\(z\)). Even in a homogeneous soil, however, the tension/moisture relationship tends to be non-linear at the upper and lower limits. If a linear approximation is applied, it will be seen that the relationship will still not hold for a homogeneous soil overlying impermeable bedrock since the resultant horizontal equipotentials direct flow vertically throughout the whole soil profile (Fig. 4A). Alternatively, a situation in which only lateral unsaturated flow was taking place (Fig. 4B) would require equipotential lines orthogonal to the slope. The tension/height relationship (\(d\psi/dz\)) cannot then be constant for both upslope and up-profile directions. Since tension gradients will be low at the start of drainage (and equipotential lines consequently nearly horizontal) it is possible to envisage a continuum from near-vertical to near-lateral flow as drainage progresses (Fig. 4C). Equipotential lines should rotate, until nearly orthogonal to the slope, and simultaneously move upslope as moisture content decreases.

In contrast to gravity drainage, upwards capillary drainage of the soil to an evaporating surface will be marked by a reversal of the normal potential gradient so that total potential decreases with height up the profile (that is: \(d\psi/dy > 1.0\)
with $y$ the height above profile base). Assuming that plant roots decrease in density with depth below the surface, transpiration suction processes should reinforce this pattern. During drainage, capillary suction will tend to draw water from an increasing depth over time as the potential reversal extends downwards (Fig. 4D). Capillary action is not limited by the field capacity of the soil but may exert tensions of up to 10,000 cm/water (Buckman and Brady, 1969). In other words, water may eventually be drawn up the entire profile length in a soil where tension at the profile base plus soil depth is less than 10,000 cm.

These theoretical considerations of the impact of gravity and capillary drainage on unsaturated flow patterns were used as a starting point for the sorting of empirical data from the experimental slope. Observations were made during a number of prolonged drainage cycles. Of those cycles, two are used here for discussion: cycle 1: 18 September—17 October, 1969; and cycle 2: 17 May—16 June, 1970.

Within each drainage cycle, relationships were sought between soil water tension, height above the slope base ($z$ in cm), height above the profile base ($y$ in cm) and time from start of drainage ($t$ in days).

**Height above slope base**

Using B horizon data alone, the general linear relationship:

$$
\psi = a' + b'z
$$

held approximately true for any point in time (Fig. 5). The data range probably does not include, therefore, the non-linear component of the moisture/tension relationship. Within each drainage cycle, $a'$ and $b'$ both increased over time. Thus for cycle 1:

$$
a' = 6.60 + 2.810 t \quad r = 0.991; \quad n = 12 \\
b' = 0.03 + 0.004 t \quad r = 0.947; \quad n = 12
$$

These relationships suggest that the upslope moisture gradient in the B horizon is low for any one point in time, and that the whole horizon drains at an almost constant rate. In other words, the rate of drainage does not increase rapidly with height. Consequently, there is little tendency for the tension/height gradient to rotate towards the expected optimum of $\psi = z$ (Fig. 5). For the lower part of the slope at least, B horizon drainage appears to be dominated by a general, presumably vertical, movement of water out of the horizon. The point $z$ on the slope at which drainage rate becomes dominated by upslope changes in moisture ($b'$) rather than by general drainage ($a'$) can be calculated as:

$$
\frac{db'}{dt} \cdot z = \frac{da'}{dt}
$$
Dots decrease in processes should tend to draw water extends downwards of the soil but (Brady, 1969). In profile length in an 10,000 cm. and capillary drain- for the sorting were made during t were used here for cycle 2: 17 May—
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Height above profile base

Treating each tensiometer bank as a separate unit (a single point upslope), the relationship between tension and height above profile base was calculated for each time of observation within each drainage cycle. The general linear relationship can be written as:

\[ \psi = a'' + b''y \]

The relationship in this case is a linear generalisation of a complex vertical pattern in which tension can be expected to be a non-linear function of both porosity and moisture variations with height. The number of measurement points was too small to allow the fitting of a more complex function but some generalisation was required for comparison purposes.

During cycle 1, tension consistently decreased with profile height indicating a rapid increase of potential up the profile giving downwards movement of water. During cycle 2 this pattern was initially repeated but was soon replaced by a phase during which tension increased up the profile, but not rapidly enough to compensate for increasing potential as y increased (i.e., \( b'' \) was positive but < 1.0). Water therefore continued to move vertically downwards through the profile. Finally, tension increased more rapidly than height for the rest of the cycle indicating potential reversal (Fig. 6), and upwards movement of water. Overall, therefore, the measured part of the profile (A/B and B hori- zons) at all measuring sites indicated that the early stages of drainage are dominated by downwards vertical water movements, the late stages by upward vertical water movements with possibly a short-lived phase of lateral movement between.
The variation of soil water tension with height above profile base at 1.0 m upslope during drainage ($t = \text{time since start of drainage in days}$).

**Time from start of drainage**

The change in tension over time was calculated for each tensiometer cell during each drainage cycle. The best-fit equations showed a semi-logarithmic relationship of the general form:

$$\log \psi = c + ft$$

(see Fig. 7)

No relationship was found between rate of drainage ($f$) and height of the tensiometer cell above slope base ($z$) which tends to confirm the observation made above that water movement in the measured profile is vertical rather than lateral. A relationship was found between rate of drainage and height of the cell above the profile base ($y$) for drainage cycle 2:

$$f = 0.110 + 0.120y$$

$r = 0.735; \quad n = 17$

No such relationship was found for cycle 1. The result probably reflects the earlier observation that a potential reversal occurred in the soil profile during cycle 2 but not during cycle 1. The difference between the cycles is probably due to differences in the rate of evapotranspiration (2.0 mm/day for cycle 2 and 1.0 mm/day for cycle 1; data from Bristol Avon River Authority records).

The pattern of unsaturated flow during drainage is therefore dominated in the early stages by vertical downwards movement through the soil profile. Since discharge from the slope base is continuous, lateral flow systems (probably unsaturated) must operate within the B/C horizon. Unsaturated lateral flow can only occur in the higher soil horizons during the limited-duration change from downwards to upwards vertical movement. As drainage by gravity...
continues, moisture content decreases and the potential gradients in the unsaturated zone fall, resulting in a progressive decrease in discharge from the slope base. At the same time upwards capillary drainage to an evaporating surface extends down the soil profile. Given time all drainage of the slope would be effected by evapotranspiration. In practice this situation was not observed since slope discharge continued through rainless periods of up to 42 days.

**The pattern of hillslope flow during a storm**

During a storm, rain falls onto the slope where one of the unsaturated flow patterns described above prevails. After a period of prolonged drainage with high evapotranspiration the slope will exhibit reversed potential gradients and a light storm may be simply absorbed into the upwards capillary flow system. A more substantial rainstorm will satisfy near-surface water deficiency to the point where potential again increases up the profile. Alternatively, where reversed gradients have not already been established at the start of the storm, further rain will increase the normal potential gradient. In either case new water will move vertically downwards in an unsaturated state. The first situation is illustrated in Fig. 8 where a pre-storm reversed-potential system (at time 1) is turned into a normal downwards flow system (at time 2).

Although the details are not fully understood, it would appear that downwards unsaturated percolation following substantial rainfall overloads the lateral unsaturated flow system of the B/C horizon. Water then accumulates at the profile base until saturated conditions are produced. The limited evidence available suggests that under heavy infiltration conditions this accumulation...
process may occur initially at the base of the B horizon resulting in a temporary "perched" saturated zone overlying an unsaturated B/C horizon. Discharge from the B horizon at the slope base will then occur before any response is noted from the B/C horizon (Fig. 1). The "perched" situation is unlikely to last long since infiltration into the B/C horizon will continue at maximum rate and a vertically integrated saturated zone is probably in existence by the time of peak discharge.

Since a slight upslope moisture gradient normally exists on the slope, saturation will develop from the slope base. Because the moisture gradient is fairly low, initial expansion of the saturated zone upslope will be rapid. Downslope saturated lateral flow increases as the saturated zone grows and peak discharge at slope base must represent the equilibrium point between vertical unsaturated supply and downslope removal by saturated flow. The main phases of storm flow on the slope are generalised in Fig. 9.

Following peak slope discharge, the saturated zone contracts; that is, soil within the zone changes from the saturated to unsaturated state. Contraction occurs as saturated lateral flow exceeds unsaturated supply and will continue until a new equilibrium between unsaturated supply and the size of the saturated zone is reached. In practice this new point occurs when the saturated zone has shrunk to a small wedge adjacent to the discharging face. From here on, discharge from the slope is dominated by unsaturated lateral flow through the B/C horizon (at least until evapotranspiration becomes effective).
1.0 m upslope during discharge; \( t_4 \) = during

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Fig. 9. Generalised pattern of soil water tension and pressure on the lower part of the experimental slope during a storm. Saturated zone is shaded. Dotted lines indicate arbitrary soil water tension contours with \( \psi_0 \) at the water table: A. Before rainfall lateral unsaturated flow dominates. Tension increases with height up profile. B. During infiltration tension decreases towards top of profile giving vertical water movement. Unsaturated lateral flow continues in lower part of profile. C. Moisture is distributed vertically. Top of profile drains relatively, vertical flow causes growth of saturation at profile base. D. At peak discharge the extent of saturation is at maximum. Lateral saturated flow equals vertical and lateral unsaturated flow additions.

It has been noted previously (Weyman, 1970) that the recession curve of total discharge from the experimental slope has three components:

\[
q = 65.00 \ t^{0.442} \\
q = 63.75 - 99.98 \ \log t \\
q = 18.25 \ t^{0.729}
\]

for \( t = 0.1 - 1.0 \) days

for \( t = 1.0 - 4.0 \) days

for \( t = 4.0 - 30.0 \) days

where \( q \) is total slope discharge in cm³/min; and \( t \) is time from peak discharge in days.

The form of this recession is remarkably similar to that reported by Hewlett and Hibbert (1963) for drainage from an artificial slope. They concluded that a two-phase process was taking place: the drainage of large and small pore spaces, respectively, with a transition period between. In the present experiment, the
first stage of drainage is dominated entirely by saturated lateral flow which probably operates largely through the non-capillary spaces of the soil. After 4 days the situation is dominated by unsaturated lateral flow to a small saturated zone, the lateral component being capillary in operation. Between the two there is a transition phase where the saturated zone is contracting rapidly and no equilibrium exists between flow within the saturated zone and unsaturated supply to that zone.

SUMMARY AND DISCUSSION

(1) The surface infiltration capacity of the soil on the experimental slope is too high to permit the generation of infiltration-excess overland flow (Horton, 1945), at least under the observed range of rainfall conditions.

(2) The response of the hillslope to rainfall is dominated by saturated throughflow within the mineral layers of the soil.

(3) There is no evidence to suggest that lateral unsaturated flow contributes to the storm response of the slope. Analysis of the potential flow nets indicates that unsaturated flow during a storm will be entirely vertical.

(4) Analysis also indicates that unless the wetting front is saturated, the initiation of saturated lateral flow is dependent upon some break in the vertical permeability profile of the soil.

In this case suitable breaks occur at the profile base and at the base of the B horizon. Whipkey (1965) and Knapp (1970) have also observed the initiation of saturated flow above permeability breaks. Where no suitable permeability breaks occur no substantial throughflow has been found.

(5) The response of the hillslope to rainfall is considerably delayed in this instance because the lateral flow system is at some depth below the surface and peak discharge is relatively low because lateral velocity is very slow. Although a major part of the basin hydrograph (Weyman, 1973) throughflow does not provide stormflow. Whipkey (1965) and Knapp (1970) both observed proper stormflow but the contrast, particularly with respect to the operation of near-surface soil horizons, may reflect rainfall rather than soil differences. Whipkey simulated high rainfall intensities, which may have exceeded horizon infiltration capacity or created a saturated wetting front, while Knapp was working in an area of higher natural rainfall and consistently high soil moisture.

(6) Because downslope flow during drainage results in an upslope moisture gradient, the zone of saturation expands during the course of a storm in the form of a wedge. Discharge at the slope base is a non-linear function of saturation depth resulting from lateral flow through increasingly permeable horizons towards the ground surface. Velocity of downslope flow would increase markedly if the top of the saturation zone intersected the ground surface to give
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an expanding area of saturated overland flow: the “dynamic watershed” model (Tennessee Valley Authority, 1965). In this case, however, slope convexity permits saturated flow velocity to increase downslope: downslope discharge increases without increase in saturation depth which would bring the water table to the surface. On the other hand, the variable extent of the saturated flow system does result in non-linearities in the relationships between rainfall/antecedent soil moisture and runoff volume/hydrograph shape (Weyman, 1973) as would be expected from the “dynamic watershed” model.

(7) During drainage the saturated zone contracts and is replaced by vertical unsaturated flow to an unsaturated lateral flow system in the B/C horizon (again, the presence of a permeability break at bedrock seems crucial for the maintenance of lateral flow). Saturated flow appears to be dominated by the non-capillary spaces of the soil and capillary pores dominate unsaturated flow. A change in dominant soil state on the slope is, therefore, marked by a change in the recession rate of slope discharge as suggested by earlier work (Hewlett and Hibbert, 1963). There is no evidence to suggest that any part of the hill-slope flow occurs within macro-voids of the soil which are hydraulically unrelated to the main soil mass.

(8) Once unsaturated flow dominates the slope, discharge continues for prolonged periods without further recharge and can therefore supply stream baseflow. Drainage by upwards capillary movement to an evaporating surface becomes increasingly important as drainage progresses but, in the present case, never actually eliminated the process of downslope drainage by gravity.

CONCLUSION

It is now possible to provide some sort of answer to the questions raised at the beginning of this paper.

Firstly, it appears that throughflow will only occur where there are breaks in the vertical permeability profile of the soil. Unless the wetting front is itself saturated, soil saturation will only occur above some impediment to further vertical movement. Once saturated conditions are generated lateral flow is bound to occur since the equipotential lines within the saturated soil will be nearly orthogonal to the slope. A similar potential field is difficult to achieve in the unsaturated state since the existence of an unsaturated wetting front implies increasing mechanical potential towards the surface and downwards flow. On the other hand, since a permeability break occurs in the soil/bedrock interface in a wide range of situations, it is reasonable to suggest that throughflow, in one form or another, is a very widespread phenomenon.

The same line of reasoning also suggests that the hillslope hydrograph is likely to be dominated by saturated, rather than unsaturated, flow immediately after rainfall. Similarly, the velocity of saturated flow will clearly be higher
than that of unsaturated flow since the potential gradients will be greater and average hydraulic conductivity greater (non-capillary pore spaces being utilised). Baseflow, however, is still likely to be dominated by unsaturated lateral flow on the slope since the saturated zone can be expected to contract rapidly once rainfall has ceased.

The identification of saturated flow with non-capillary pore spaces and unsaturated flow with capillary pores is fairly obvious. Experiments of the type described here can only indicate that average hydraulic conductivity changes with a change in soil state and leads to a marked difference in flow velocity and drainage rate. Moreover, the analysis assumes that Darcian flow holds for both saturated and unsaturated states. A hydraulically distinct system of non-capillary spaces will require the application of other flow laws for part of the flow and renders the parameter “average hydraulic conductivity” rather meaningless (Chamberlin, 1972).

As far as the impact of throughflow on the basin hydrograph is concerned, the picture remains a little obscure. Whilst it is clear that throughflow will contribute to the recession limb and baseflow in many basins, the recorded instances of subsurface stormflow remain very few. In a hydraulically integrated flow system the primary controls on hillslope response are depth to the lateral flow system and the velocity of lateral flow (a function of potential gradient and lateral hydraulic conductivity). Depth to lateral flow will depend largely upon the vertical permeability profile of the soil and rainfall intensity. Ragan (1968) reported stormflow from a saturated litter horizon when rainfall intensity exceeded the infiltration capacity of the underlying soil horizon. If water has to move to the base of the soil profile before lateral flow occurs, response will be considerably dampened. Similarly, lateral velocity of flow will control the magnitude of hillslope response. Lateral velocities appear to be very low in the mineral horizons of the soil. Although lateral hydraulic conductivity will probably increase up the soil profile, it is only within the litter horizon that overland flow velocities are likely to be approached and at that point it is no longer clear whether flow is really subsurface or a form of overland flow (Pierce, 1965). It is only in a soil where vertical and lateral discrete non-capillary routes exist that the lower soil horizons will contribute to stormflow. Knapp (1970) indicates that rapid response in lower soil horizons may be related to non-capillary flow routes within a solifluction material. Otherwise it is probably safe to assume that overland flow, whether in the classic Hortonian sense or as saturated overland flow, will still dominate the storm response of most drainage basins.
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