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RELATION OF FIELD STUDIES AND MODELING IN THE PREDICTION OF STORM RUNOFF

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ABSTRACT

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Routine methods of predicting storm hydrographs do not require accurate conceptual models of storm runoff processes. Application of hydrology to the analysis of landform evolution and other scientific and land-management problems requires realistic concepts of runoff processes and their variation within drainage basins. These concepts need to be refined, developed, and formalized through more vigorous combination of rigorously designed field experiments and realistic physically-based mathematical models.

STATEMENT OF THE PROBLEM

The most frequent exercise in surface hydrology is the prediction of flood discharge from drainage basins. If a sufficient hydrologic record exists in the basin or region, such forecasts are made with adequate success on the basis of empirical methods such as flood-frequency analysis, unit hydrographs, and loss-functions coupled with flood routing. The methods do not rely upon well-defined conceptual models of the mechanisms generating runoff.

Since the development of Horton's (1933) infiltration theory of runoff, some flood prediction methods have incorporated parameters that are identified with infiltration, storage, flow resistance, and other aspects of runoff. These procedures are partially decomposable into elements that express the physics of runoff processes, and are therefore more flexible and informative than correlation alone. However, the parameters are usually obtained through back-calculation from hyetographs and basin hydrographs. It is possible to calibrate these methods for the prediction of future hydrographs, but such calibration is often possible whether or not the original conceptual model of the runoff process is valid for the basin of interest. Thus, a device such as the Stanford Watershed Model (Crawford and Linsley, 1966) or its derivatives (Burnash et al., 1973) can be calibrated to reflect an infiltration capacity and a flow resistance for overland flow in basins where the original conceptual model of storm-runoff generation is inappropriate.

The necessity of calibration and the potential for confusing model parameters with physical processes can degrade attempts to predict the effects of land use, such as forest cutting or fertilizer application on stormflow and water quality. If the model is based on infiltration theory, an increase in peak runoff after deforestation can be interpreted as a result of declining infiltration capacity even when the forest floor remains intact and infiltration is not a limiting factor. Or, lumped indices of pollutant entrainment can be calculated when the pollutants are being mobilized only from small zones of the basin with a particular chemistry and runoff mechanism related to moisture conditions. These limitations become important in predicting the success of various management options to reduce the undesired effects of land use.

Application of hydrology to the prediction of sediment loss, water quality, and landform evolution on hillslopes or entire drainage basins has increased the need to understand various characteristics of runoff *within* the catchment, such as the spatial variability of runoff generation, the flow depth, speed, and entrainment potential for sediment or chemicals. Table I is a short list of issues, the study of which requires a more realistic picture of runoff characteristics within drainage basins than can be obtained through back-calculation from data collected at the basin outlet.

A realistic and detailed picture of runoff characteristics within a drainage basin can be obtained for sample areas by intensive field studies and by mathematical modeling based upon rigorous definition of the physics of runoff. These activities should preferably be conducted jointly, but it is a reflection of the current nature of hydrological research that they are not so conducted.

The field studies involve the measurement of runoff rates and flow characteristics and of the concentration and pressure of water on hillside plots, ranging in size from one to several thousand square meters. They have sampled natural events as well as controlled simulated rainstorms. The mathematical models involve one component to describe the transient unsaturated—saturated regime of infiltration and subsurface flow, and another component to route surface flow over hillslopes and along channels by means of the kinematic approximation of the shallow-water flow equations (Wooding, 1965—1966). The requirements of data and computing power for the models are extremely high and the models are not used for routine design problems. However, they are valuable as hydrological tools in the following ways:

(1) They provide rigorous physical explanations for the apparently disparate results of field measurements in diverse regions, and allow them to be drawn together into a comprehensive theory of runoff generation.

(2) They increase efficiency in the use of data from properly designed field studies by allowing interpolation and extrapolation of parameters from a few studies.

(3) They suggest the kind and rigor of measurements needed to make field experiments definitive and to fill gaps in the range of conditions that need to be documented in the field.

(4) They provide theoretical explanations for the general form of empirical relations widely recognized in hydrology.

(5) They allow one to assess the effects of spatial—temporal variation of controlling factors on the variability of runoff.

(6) They explain and predict spatial heterogeneity of runoff generation that is important not only in the computation of flood hydrographs, but in the understanding of such processes as sediment production and the washing of pollutants from a landscape under natural or disturbed conditions.

(7) They allow anticipation of the effects of changing the controlling factors of runoff before such changes occur and therefore before their effects can be documented empirically.

(8) They predict other characteristics of runoff besides its discharge rate and timing; these characteristics include the depth and velocity of sheetflow (which are useful in sediment transport models of erosion), pore pressures (which are useful in understanding hillslope stability), and the residence time of water in soils and channels (which is of interest to geochemists interested in weathering reactions).

TABLE I

Issues that require realistic models of runoff processes

Geomorphology	(a) hillslope and hollow stability (b) hillslope evolution by water erosion (c) channel incision
Soil erosion	(a) distributed models of sediment transport along hillslopes (b) localized sources of sediment (e.g., forest roads)
Biogeochemistry	(a) effect of residence time on weathering reactions (b) entrainment of pollutants from land surfaces

AIM OF THE PRESENT PAPER

The value of theoretical models can be greatly enhanced if they are developed in close cooperation with field studies. Such cooperation ensures that the physics of the problem is well understood and that the model is an adequate description of field conditions. The interaction should also stimulate the evaluation of field parameter values necessary for modeling, and also the organization of field data to modulate the more discouraging aspects of the current concern with the stochastic aspects of runoff generation (Freeze, 1980).

This paper will review the present connection between modeling and field studies of runoff. It will describe the status of physically-based modeling of runoff founded on accurate descriptions of runoff and quantification of physical parameters from the field. Finally, some suggestions will be made of opportunities for further modeling and fieldwork, and particularly for cooperative work between these often-independent activities in hydrology. The review will cover the study of various runoff mechanisms during rainfall and then the special case of runoff generated by these same mechanisms under snowmelt conditions that cause significant differences from runoff generated by rainfall.

MECHANISMS OF RUNOFF GENERATION

Field research has identified the following mechanisms by which storm runoff may be generated:

(1) *Horton overland flow* is generated when rainfall intensity exceeds the infiltration capacity of the soil (Horton, 1933). This process is likely to occur most frequently in arid and semi-arid landscapes where vegetation densities and therefore infiltration rates are low, and in disturbed areas of humid landscapes, such as cultivated fields, paved areas, mine spoils, construction sites, and rural roads. However, even on grassy hillslides in humid regions, some clay soils have saturated conductivities that are low enough to be exceeded by relatively frequent rainstorms. For example, M.G. Anderson and Kneale (1980) report conductivities of 0.36—36 mm $hr.^{-1}$ for a pasture in southwest England. Horton overland flow may be confined to only a portion of the drainage area, an effect which Betson (1964) referred to as *partial-area* runoff.

(2) *Subsurface flow* contributes significantly to storm runoff where soil conductivity is high because of coarse soil texture or large structural openings in thickly-vegetated soils. The subsurface flow may enter the stream through the perennial groundwater body (Ragan, 1968; Sklash and Farvolden, 1980), or it may be diverted laterally by some layer with a vertical conductivity less than the rainfall intensity (Hewlett and Hibbert, 1963; Whipkey, 1965). Zones with concave topographic contours tend to contribute more subsurface stormflow than other zones, either because the pre-storm water table tends to be closer to the surface (Dunne and Black, 1970a) or because the geometry of the impeding layer causes subsurface flow to converge (Pierson, 1977; M.G. Anderson and Burt, 1978). Subsurface stormflow dominates the flood hydrograph and other phenomena such as landsliding and soil-water geochemistry most commonly in humid regions with dense vegetation and permeable soils on steep hillslopes. Such areas include the southern Appalachians, U.S.A. (Hewlett and Nutter, 1970); the Oregon Cascade Mountains, U.S.A. (Harr, 1977); the Southern Alps of New Zealand (Mosley, 1979); and the hills of southwestern England (Weyman, 1970; M.G. Anderson and Burt, 1978).

(3) *Saturation overland flow* develops where the soil becomes saturated by the perennial groundwater table rising to the surface or by lateral or

vertical percolation above an impeding horizon (Dunne and Black, 1970a, b; Bonell and Gilmour, 1978). Some of the water that has been moving slowly through the topsoil emerges and flows over the ground surface as return flow (Musgrave and Holtan, 1964) and is augmented by direct precipitation onto the saturated area, which expands and contracts seasonally and during rainstorms. Saturation overland flow tends to dominate storm runoff in thickly-vegetated landscapes with thin soils, high water tables, and long gentle concave hillslopes. Subsurface stormflow in these regions may still be important in contributing to the recession limbs of hydrographs from small basins as well as affecting landsliding, soil drainage, and water chemistry. Thus it is necessary to combine the study of subsurface flow and saturation overland flow.

Temporal and spatial variation of the zones that generate subsurface stormflow led to the use of the terms *variable-source concept* (Hewlett and Hibbert, 1967) and *dynamic-watershed concept* (T.V.A., 1964). Dunne and Black (1970b) and Dunne et al. (1975) emphasized the fact that in some regions those dynamic contributing zones generate saturation overland flow.

The environmental controls on the various mechanisms of storm flow are summarized in Fig. 1. The arrows between labels imply a range of storm frequency as well as drainage basin characteristics. For example, a hillslope may generate only subsurface flow during a gentle rainstorm, and Horton overland flow during a deluge; or subsurface flow alone during a short rainstorm and saturation overland flow during a long one.

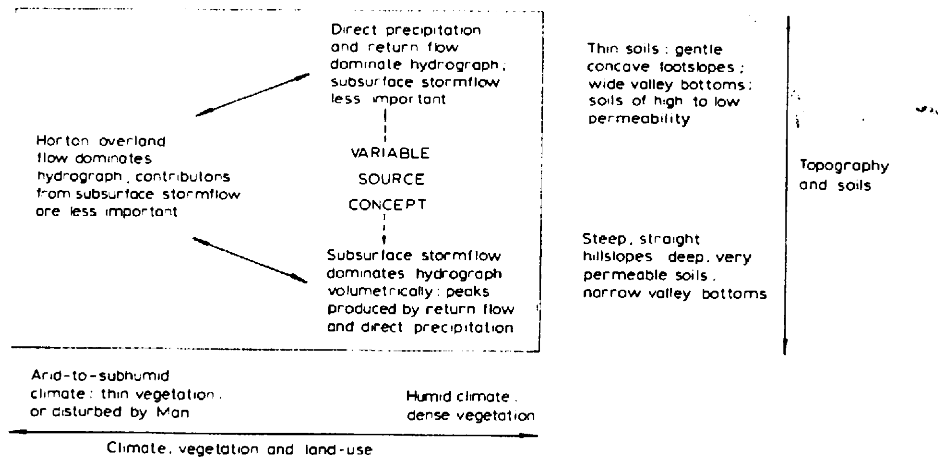

Fig. 1. Schematic illustration of the occurrence of various runoff processes in relation to their major controls.

STATUS OF RESEARCH

Horton overland flow

The understanding and prediction of Horton overland flow depend upon theoretical and field studies of infiltration and the hydraulics of thin films of flow. For many years after Horton (1933) had emphasized the importance of infiltration capacity, this variable was widely measured in the field with cylinder and sprinkler infiltrometers. The work qualitatively identified the major factors controlling infiltration. The most important controls are generally agreed to be: soil texture and structure, initial moisture content, and some aspect of vegetation density or organic content of the surface. Other properties associated with these variables, such as swelling behavior associated with various clay minerals or the densification of topsoil due to trampling or rainpacking, have been identified in some places. These empirical studies are useful for developing indices of runoff potential or for various tasks in land management to reduce erosion or pollution. However, the number of controlling variables, the inherent variability of results due to measurement and sampling errors, and the tendency of field scientists to emphasize the particular rather than the general aspects of their results has led to very few useful generalizations from these data (e.g., Musgrave and Holtan, 1964). Thus, it is not often possible a priori to make an assessment of infiltration capacity without a time-consuming field study.

The development of useful generalizations from field measurements of infiltration capacity was originally hindered by a lack of rigor in the popular conceptual models of the infiltration process (e.g., Green and Ampt, 1911; Horton, 1940). Klute (1952), Philip (1957) and Rubin and Steinhardt (1963) developed methods for computing infiltration based on a rigorous theory of the physics of vertical flow in a saturated or unsaturated porous medium. Both analytical and numerical methods were developed for these computations. The full numerical calculations are complicated because both the soil hydraulic conductivity and pore pressure depend upon moisture content, and consequently vary during the storm until the topsoil becomes saturated and the infiltration capacity declines to the saturated hydraulic conductivity. It is time-consuming to document the necessary relationships between conductivity, pressure and moisture content, and most modeling efforts have relied upon a few data sets (e.g., Brooks and Corey, 1966). The data requirements increase rapidly if the soils are anisotropic or layered. The demanding data requirements and the complexity of the numerical computations limit the widespread application of this approach to field problems.

R.E. Smith (1972) has illustrated how numerical computations of infiltration based on a rigorous physical theory can be useful for bringing some order among field measurements and for extracting information from infiltration experiments, thereby enhancing their predictive value. Using measured hydraulic properties for a few soils in a numerical model of soil-

moisture flow, he determined that a simple four-parameter curve could be fitted to the results and described by a single formula. The parameters of the formula could be obtained from a single infiltrometer measurement, and could then be used for predicting the form of the infiltration curve at different rainfall intensities and initial moisture contents. Much recent research has concentrated on developing a physically reasonable infiltration equation (e.g., Morel-Seytoux, 1976), which is simple enough for routine field application.

The utility of these physically rational infiltration models has not yet been widely tested in the field. This appears to be partly because field studies of infiltration have not included measurement of the critical parameters required for the models and for comparison with data from other sources. The sophistication of field studies is lagging behind that of the modeling, and improvements are necessary in the design of infiltration experiments and the measurement of critical variables that control the physics of infiltration. A second reason for the lack of testing is the failure of models to address the issues which many field researchers claim to be the overwhelmingly important controls of infiltration, namely the roles of crusting and vegetation.

It has been shown by laboratory experiment that when some bare soils are exposed to raindrop impact a thin surface crust of dispersed fine particles forms a less permeable layer one to several millimeters thick. McIntyre (1958a, b) demonstrated that such a crust developed on a sandy loam could have a saturated conductivity two to three orders of magnitude lower than that of the undisturbed soil below. Hillel and Gardner (1969) have theoretically analyzed the role of these crusts in controlling infiltration. Morin and Benyamini (1977) have presented field data on the effect of crust formation on infiltration but the evidence is circumstantial. Furthermore, it is not possible to transfer the available laboratory results on crusting directly to the field. The detailed laboratory examinations of crust formation (McIntyre, 1958a, b; Farres, 1978) have occurred under ponded water or on unsaturated soils. In the field, overland flow winnows fine particles from a surface that is disturbed by raindrop impact. Dunne and Dietrich (1980a) proposed that an increase of infiltration capacity with an increase in simulated rainfall intensity may have been due to removal of a soil crust under the higher turbulence and shear stress generated during the more intense rainfall. Current infiltration models do not include the effects of crusting, and the available field studies do not adequately define the necessary conditions for crust formation or destruction. Thus, its hydrologic significance, though widely referred to, remains unknown.

It is claimed that vegetation is the most important and multifarious influence upon infiltration, being responsible for intercepting the energy of raindrops, depositing a surface mulch, and altering the pore-size distribution of soil through aggregation and root penetration. Yet there is not a rigorous theory describing the influence of vegetation and its interaction with other variables, and there are no detailed measurements of the physical parameters controlling flow into a vegetated surface. Therefore, there is little immediate hope of rationalizing the often conflicting results of field studies on the effect of sparse vegetation covers on infiltration. Johnson and Niederhof (1941) and Marston (1952) failed to uncover any relationship between these variables in soils with high infiltration capacities and a narrow range of cover density. Moore et al. (1979) found no differences in infiltration capacity related to vegetation cover density between 0 and 57% in southern Kenya, and Kincaid et al. (1963) showed little if any increase of infiltration capacity until the cover exceeded ~ 20%. On the other hand, H.L. Smith and Leopold (1942) and Dortignac and Love (1961) documented obvious changes in infiltration with vegetation density.

Other controls of infiltration that have been shown to be important by field workers include the cracking and swelling of clay soils (Bouma et al., 1978; Dunne and Dietrich, 1980a) and the progressive inundation of micro-topography as runoff rates increase (Moldenauer et al., 1960; Hawkins, 1981).

In the solution of these problems, there exist many opportunities for cooperation between field workers, who need to make some critical measurements during their experiments, and those interested in simulating the runoff process by means of realistic distributed models. At present, the two groups are focusing their attention on different issues.

Another issue which requires the combined attention of theorists and field researchers is the spatial heterogeneity of infiltration within a drainage basin. Musgrave and Holtan (1964) demonstrated the feasibility of using a sprinkling infiltrometer to map infiltration capacity in small basins, although no analysis of errors was provided. Sharma et al. (1980) were not so successful when using a cylinder infiltrometer. Nielsen et al. (1973) reported that the spatial variability of unsaturated conductivity was particularly high. R.E. Smith and Hebbert (1979) and Freeze (1980) have analyzed the potential variability of runoff caused by random variation of infiltration capacity along hillsides. Fortunately, the variance of runoff rate decreases through a storm as the mean unsaturated conductivity converges on the saturation value. The unsaturated conductivity is much more variable at a given degree of saturation or time. The variation of time to ponding also depends upon the saturated conductivity and initial moisture content for a given rainfall intensity. As the ratio of rainfall intensity to average saturated conductivity decreases, there is a greater variability in response and more runoff occurs sooner than on a hillside with a uniform conductivity. Smith and Hebbert also pointed out several important consequences of spatial trends that may occur in addition to random variation. For example, if the infiltration capacity tends to increase downslope, the precipitation excess generated upslope provides infiltrable water that reduces the ponding time and its variance along the flow path downslope. These papers emphasize the potential for field work to document typical frequency distributions of the parameters controlling infiltration and also to compile maps and hillslope profiles

on which typical spatial patterns of infiltration capacity have been defined. For this work, the concepts used in soil mapping and hillslope geomorphology should be useful. In both of these fields the concept is well established that soils distributed along a hillslope profile are often related genetically and in their physical properties. Thus in soil science, the catena concept is widely used in some landscapes (Paton, 1978), while in geomorphology Furley (1971) among others has studied typical spatial patterns of soil properties along hillslopes.

The depth, velocity and discharge of Horton overland flow are usually calculated analytically or numerically from the kinematic approximation to the shallow-water equations (Wooding, 1965—1966; Woolhiser, 1975). It is necessary to define the geometry of the hillslope or basin and to specify a resistance equation, such as the Darcy—Weisbach equation. The technique is relatively straightforward and reliable, but the data requirements vary dramatically according to the purpose for which the results are used, and therefore the degree of realism required in the description of the runoff within the basin. If it is necessary to predict the basin hydrograph, the geometry of the catchment has been successfully represented as two planes (Wooding, 1965—1966), a cascade of planes (Woolhiser et al., 1970), or a convergent surface (Woolhiser, 1969). These approximations often ignore the differences between sheetflow and channel flow, which are not necessarily important for hydrograph prediction. The friction factor varies with the Reynolds number of the flow, and the relationship between the two is characteristic of the surface; it is usually obtained by maximizing the fit between computed and observed results. Infiltration capacity is also usually obtained by analyzing rainfall and runoff data. However, considering the number and length of hydrologic records from plots and small basins, there seems to be a remarkable dearth of useful, homogeneous data sets for such calibration (R.E. Smith and Woolhiser, 1971; Singh, 1976, p.905).

These sorts of approximations are less useful for purposes that require some knowledge of the nature of flow within the drainage basin. For example, soil conservation scientists studying the physics of erosion as a combination of rainsplash, sheetwash and rill erosion (Mutchler and Young, 1975; Foster and Meyer, 1975) need to predict flow depth and its variation along and across a hillslope. The effects of microrelief and rilling are crucial in this case, and the physical significance of a single friction factor is questionable. Some field researchers have conducted plot and laboratory studies to document the variation of depth, velocity, and flow resistance on natural surfaces (Ree, 1939; Izzard, 1944; Parsons, 1949; Emmett, 1970; Dunne and Dietrich, 1980b). Each of these studies involved direct measurement of mean flow depth and velocity, and the calculation of resistance coefficients.

Figs. 2—4 show typical measurements of the characteristics of sheetflow on plots under artificial rainfall. The surfaces on which the experiments were conducted are described by Dunne and Dietrich (1980b). Flow depths were measured with a thin ruler and velocities with dye. Fig. 2 shows the variability

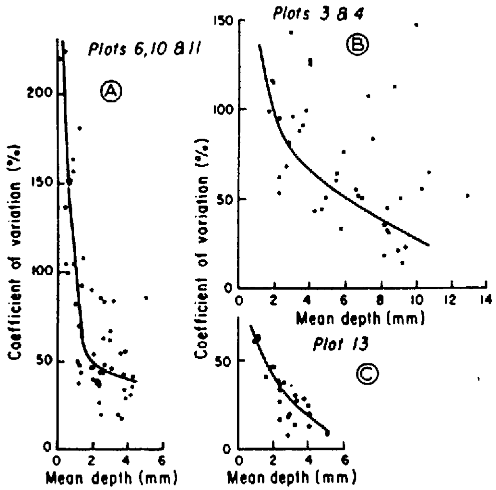

Fig. 2. Coefficient of variation of depth plotted against mean depth of sheetflow based on five measurements along a contour (Dunne and Dietrich, 1980a,b). Plot *13* was bare; plots *6*, *10* and *11* had a clumped ground cover of 10, 12, 16 and 41% during various experiments; plots *3* and *4* had covers of 36% as well as gilgai microtopography.

of flow depth across the plot, based on only five measurements at each distance from the upper boundary. The inverse relationship with depth reflects both measurement errors and the fact that in the shallowest flows even the smallest surface protuberances divide the flow. However, on the bare plot (*13*) and those with a sparse vegetation cover (*6*, *10* and *11*), the coefficient of variation declines below 50% of the mean for flow depths averaging more than 2 mm, and the sheetflow approximation is tenable. There were no rills on the hillsides used for these experiments. On the plots (*3* and *4*) with gilgai microtopography and vegetation densities of 36%, the variance of depth was much greater, as the water flowed in anastomosing patterns around clumps of vegetation or around protuberances several centimeters high.

The same kind of variation is shown by depth profiles along the plots. The depths increase with runoff rate (Fig. 3A) and with surface roughness (comparison between Fig. 3B and the curve of Fig. 2A for which runoff rates were equal). Regression analysis indicated that, on all surfaces, depth was strongly correlated with the square root of distance from the upper plot boundary, suggesting that the flow regime is intermediate between laminar, for which the exponent should be 0.33, and turbulent, for which it should be 0.67. The dye injections confirmed this as the dye tended to flow in smooth trajectories until disturbed by raindrops and wakes behind small obstacles.

The variation of friction factor ($f = 2gds/u^2$) with Reynolds' number ($N_R = ud/\nu$) indicated that the flow was laminar for $N_R < 1000$ (Fig. 4,

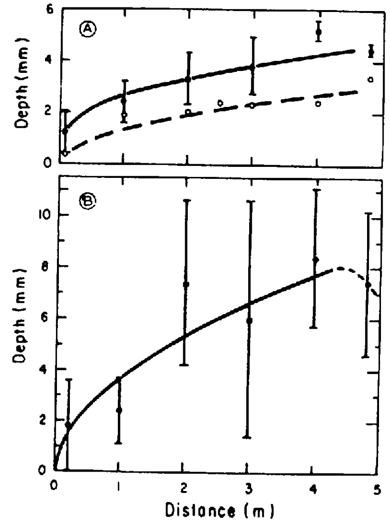

Fig. 3. Steady-state profiles of sheetflow depth along hillside plots described by Dunne and Dietrich (1980a,b). *Points* represent means of measurements along a contour and *bars* indicate the standard deviation of the measurements.

A. *Solid curve* refers to a runoff rate of 85 mm hr.$^{-1}$ and *dashed curve* to a rate of 27 mm hr.$^{-1}$ on a bare sandy clay loam.

B. Data refer to a runoff rate of 85 mm hr.$^{-1}$ on a plot with a 36% ground cover on gilgai microtopography.

which also summarizes data from other field and laboratory surfaces). The relationship expected from theory, $f = K/N_R$, was fitted to the data from each set of plots, and the value of K varied with vegetation (Fig. 5). An analysis of errors is given by Dunne and Dietrich (1980b).

Direct measurements of flow properties on small plots can also reveal typical, repeated patterns of K-values along hillslope profiles, as surface morphology and vegetation density change. Although such patterns are not quantified yet, they are obvious on many hillsides in the Kenyan savannas. Where there is no significant change in surface texture, a graph such as Fig. 5 could be used to define the variation of K and therefore of friction factor for chosen runoff rates (Reynolds' numbers) at various seasons along typical hillslope profiles. The delineation of such common patterns in the landscape is one way in which field researchers can employ their experience with regional relationships between hillslope shapes and dimensions, soils, and vegetation to reduce the variability affecting model predictions. There is opportunity for fruitful cooperation here between field studies and stochastic—conceptual modeling of the type reported by R.E. Smith and Hebbert (1979).

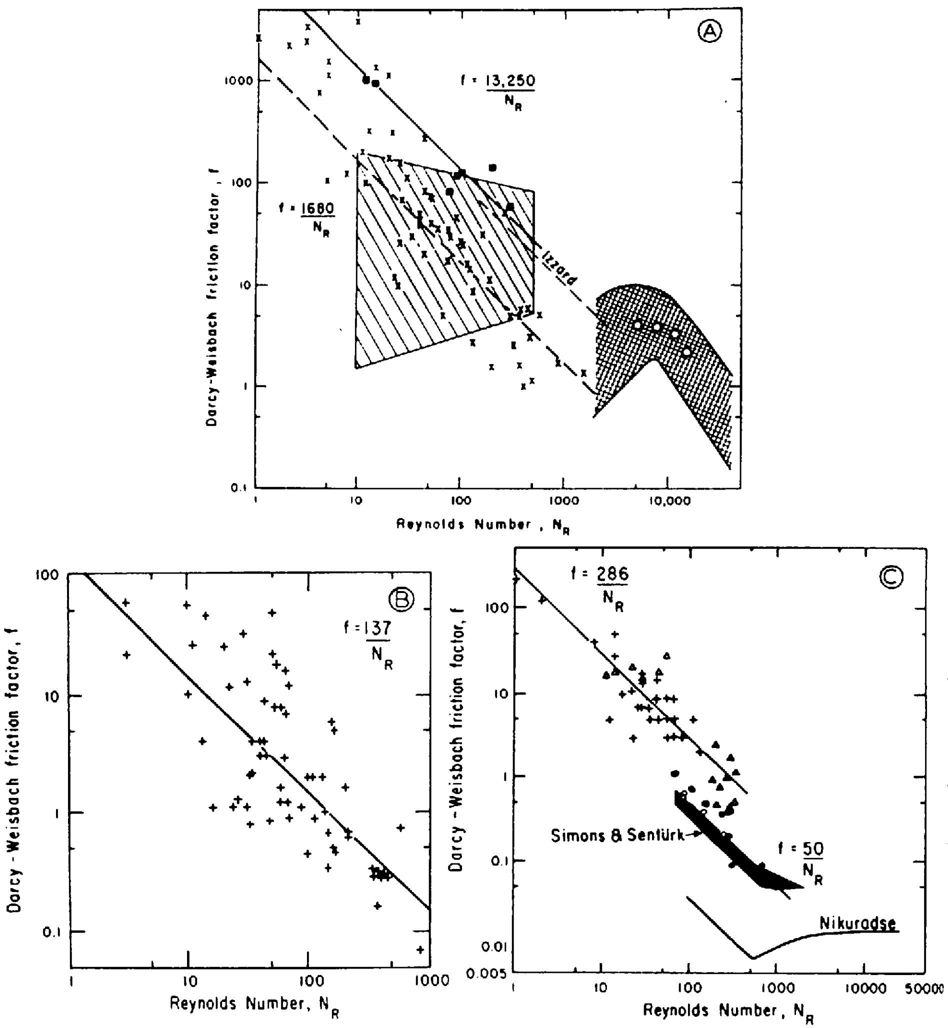

Fig. 4. Relationship of Darcy—Weisbach friction factor for various hillslope surfaces in Kenya, described by Dunne and Dietrich (1980a,b).
A. *Solid squares* and *lines* are from a grass cover of 77% on a clay soil and agree well with Izzard's (1944) data from turf. *Crosses* and *dashed line* are from a 36% grass cover on the same soil. *Lined area* indicates Emmett's (1970) measurements on Wyoming hillsides with a cover of 8—35%. *Cross-hatched* area indicates measurements by Kao and Barfield (1978) in a flume lined with artificial grass. *Open circles* are Ree's (1939) data from a flume lined with natural grass.
B. Data from a sandy clay loam with ground cover densities of 10, 16 and 41%, but the densest cover consisted of heavily clumped plants.
C. *Triangles* are from a bare gravel cover. *Crosses* are from a bare sandy clay loam with a sparse (1—2%) gravel cover. *Solid circles* are from gravel-free sandy zones. *Open circles* and *shaded region* are laboratory results from sandy surfaces.

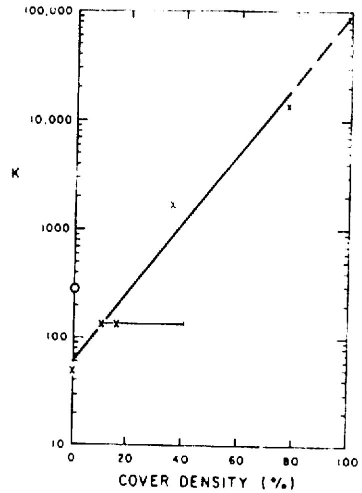

Fig. 5. Variation of K ($= fN_R$) with ground cover density on plots in Kenya rangelands. The *circle* at zero cover refers to surfaces complicated by gravel, and the *horizontal bar* indicates the range of clumped cover on one of the plots.

Subsurface stormflow

Early interest in the hydrologic importance of subsurface stormflow arose among fieldworkers in steep forested terrain such as the Appalachian Mountains (Hursh, 1944) and Indonesia (see Roessel, 1950 for review of earlier work). But the persuasiveness of the Horton infiltration theory and of early modeling efforts focused attention towards overland flow as the dominant source of flood runoff, despite persistent failure to observe this phenomenon over vast areas of the world, including much of the northeastern U.S.A. where Horton worked.

The gradual accumulation of field experiments (e.g., Hewlett and Hibbert, 1963; Whipkey, 1965; Dunne and Black, 1970a; Weyman, 1970) documented the rates and volumes of subsurface flow in a range of environments (see Dunne, 1978 for a summary). Recent fieldwork (Harr, 1977; Mosley, 1979) has extended into environments of even greater soil conductivity, hillslope gradient, and rainstorm size, all of which increase the magnitude of subsurface stormflow.

Modeling of subsurface stormflow awaited the application of finite-difference methods for the solution of the Richards equation for transient partly-saturated flow. Freeze (1969) developed a finite-difference technique for unidirectional flow recharging a water table, and then (Freeze, 1972a,b) for two-dimensional subsurface flow from hillslopes to a stream. The work generalized the few data then available on the effect of soil conductivity and hillslope form on the magnitude of subsurface stormflow. Only in soils with particularly high conductivities could subsurface contributions dominate the storm hydrograph. The method was applied to a natural hillslope on which pore-water pressures, soil moisture and runoff were measured (Stephenson and Freeze, 1974). In spite of the fact that the hillslope was the site of a detailed investigation of runoff mechanisms, the investigators encountered many problems in calibrating their model. Most of the problems arose from the difficulty of specifying the geometry of the subsurface flow region and the hydraulic properties of the soil and weathered rock.

Beven (1977) used the finite-element method of solving the Richards equation applied to the same problem, and in addition to the variables examined by Freeze, he emphasized the role of initial moisture conditions and of topography in affecting the speed and magnitude of subsurface contributions to streamflow. Convergent topography generated particularly high runoff rates.

Field research (Hack and Goodlett, 1960; Dunne and Black, 1970a, b; M.G. Anderson and Burt, 1978) and qualitative discussion (Kirkby and Chorley, 1967) have emphasized the role of topographic hollows in generating sub-surface flow. There is considerable interest in these hollows because in some mountainous terrains they are the location of most landslides, especially of those which evolve into destructive debris flows. Pierson (1977) and M.G. Anderson and Burt (1978) have conducted field investigations into the development of high pore pressures in steep hollows. Although these studies emphasize the point that pore pressures and subsurface flow are enhanced in hollows, it is difficult to use the field work results alone for quantitative predictions, either in adjacent hollows, or in hypothetical situations with combinations of the well-known governing factors. Beven (1977) employed a two-dimensional finite-element calculation with the slope width increasing upslope to simulate the effect of flow convergence in topographic hollows. The calculated subsurface outflow exhibited a higher later peak than the flow from a straight hillslope under the same conditions. Humphrey (1981) used a finite-element method to compute transient pore pressures in saturated—unsaturated flow in hollows typical of the Oregon Coast Range. Limitations of computer storage capacity confined the study to treatment of either long narrow hollows or small elliptical hollows set into planar hillslopes.

Given the computer limitations and the difficulty of collecting the necessary field data for numerical modeling of interesting problems in subsurface flow, it is salutary that simplified methods are now being developed for approximate, but adequate computations of transient subsurface flow. Beven (1981, 1982) has illustrated the usefulness of the kinematic-wave approach for calculating subsurface stormflow and water-table elevations. Humphrey (1981) also showed that in extremely permeable soils on steep slopes, his finite-element approach yielded essentially the same water-table form as a much simpler analytical method by which the unsaturated domain is assumed to be crossed by a vertical flux that travels at a constant speed, or at a variable speed which is easily calculated. This influx to a thin saturated layer at the base of the soil profile is then routed downslope by means of a convolution technique. However, as the hillslope gradient and the ratio of soil conductivity to rainfall intensity decrease, the water table rises strongly, these simple methods for routing water through the unsaturated zone rapidly become untenable.

Saturation overland flow

The occurrence of saturation overland flow was documented in the field by Dunne and Black (1970b) and Dunne et al. (1975), and was modeled by Freeze (1972b). It occurs over a contributing area that varies during and between storms, and it is easily incorporated into the variable-source or dynamic-watershed concepts. Prediction of its extent and magnitude is important in the construction of spatially distributed runoff models and in understanding sources of non-point pollution (Cahill et al., 1975). The magnitude of this contribution to the storm hydrograph is enhanced by large rainstorms, wet antecedent conditions, long hillslopes with gentle gradients, and shallow soils with a conductivity that exceeds the rainfall intensity but is too low to convey all of the rainfall as subsurface flow.

Because of its close association with subsurface flow, saturation overland flow is computed with the numerical models of transient saturated—unsaturated flow referred to above. The predicted water table rises to the soil surface over an expanding zone at the base of the hillside and creates a seepage face from which water emerges as return flow (Fig. 6). The width of this expanding seepage face perpendicular to the contour determines the saturated area onto which direct precipitation contributes to saturation overland flow. Freeze's (1972b, fig. 11) model accurately reproduces these processes. However, the difficulties of computation are aggravated by the need for a close network of nodes in the numerical representation of the topsoil so that the intersection of the rising water table and the soil surface can be well defined. It is also difficult to measure the conductivity of the porous organic-rich surfaces that typify the wet zones in which saturation overland flow is generated. The hydraulics of flow through thick ground vegetation is also poorly documented. Many of the contributing areas have thickly-vegetated rough surfaces with low gradients, and there may be considerable delay between the emergence of saturation overland flow and its entry to a stream channel.

Because of the complexity of using the complete physically-based numerical model, approximate models are now being developed. One of these, developed by Beven and Kirkby (1979), divides a basin into small units (200—2000 m^2) of homogeneous soil and topography, and computes an approximate water balance for each unit. Runoff may be generated by any of the processes described in this paper, and the dynamic spatial aspect of runoff is described by the model. However, an extensive program of simple field measurements is necessary to obtain the empirical parameters.

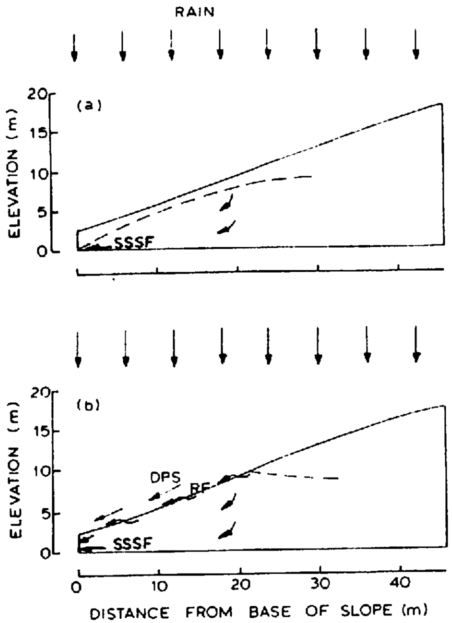

Fig. 6. During a 44-mm, 2-hr. artificial rainstorm on a Vermont hillslope, the water table rises to the ground surface over a zone that expands upslope. Beneath the central and upper parts of the hillside the water table slopes away from the stream. Flow lines are based on piezometric measurements.

a. Early in the rainstorm, subsurface stormflow (SSSF) has just begun.

b. Late in the storm, the water table intersects the ground surface allowing return flow (RF) from the expanding saturated zone. Direct precipitation onto the saturated area (DPS) also contributes to saturation overland flow.

Snowmelt runoff

Snowmelt dominates the hydrologic year in cold temperate and subarctic regions. It generates the greatest flood hazard (often in conjunction with ice jamming), and raises groundwater levels to their highest levels, often causing landslides and waterlogging that impede access, building, ploughing, and other uses of the land.

Field investigations of snowmelt runoff are difficult and uncomfortable, and therefore runoff processes have received relatively little empirical study. Consequently, there has been little stimulus for the development of models. On the other hand, the more tractable problem of melting at the snowpack surface has been the focus of much fieldwork and theoretical study. Yet runoff processes can have at least as great an influence on peak flows and their timing as do the melt parameters that have received so much study.

Rates of snowmelt depend on the influx of energy to the snowpack as

affected by weather, vegetation and topography. Models of the snowpack energy balance as affected by these variables are being developed in conjunction with field studies (E.A. Anderson, 1968; Föhn, 1973; de la Casinière, 1974; Price and Dunne, 1976; Obled and Rosse, 1977; Hendrie and Price, 1978). The models predict average hourly, and particulary daily snowmelt rates well, but their application is limited by the need for expensive sophisticated instruments operated under harsh conditions. Hourly melt rates are needed for inclusion into physically-based runoff models.

The paths of snowmelt runoff are the same as those described for rainstorm runoff, but the presence of the pack, and several other aspects of the interaction between the pack, the meltwater and the soil may alter the timing and relative importance of various runoff processes. The migration of water through the snowpack has been studied in the laboratory and in the field by Gerdel (1945), Sharp (1951), and Colbeck and Davidson (1973). It is dominated by vertical unsaturated flow unless impeded by ice layers or the soil surface, which may divert water laterally as a saturated flow. Colbeck (1972, 1978) formulated a kinematic-wave model to route meltwater through snowpacks, and several field studies (reviewed by Wankiewicz, 1978) have demonstrated its usefulness. These studies have helped to explain and draw together field results which indicate that the presence of the snowpack delays the arrival of the meltwater wave at the ground surface by up to 12 hr. or more, depending upon snow depth, temperature, grain size, and the melting rate (Sharp, 1951; E.A. Anderson, 1968; Dunne and Black, 1971; Dunne et al., 1976), and that there is then an initial abrupt efflux from the pack. The peak rate of this efflux is equal to the peak rate of melting if the pack is shallow, but may be damped slightly for packs several meters deep.

Melt rates are generally lower than most flood-producing rainfall intensities. Therefore, Horton overland flow is less likely, unless the infiltration capacity is lowered by soil frost. Thus, it is important to predict the occurrence of soil frost and its hydraulic properties. Some models have been developed to predict the occurrence of frozen soil (e.g., Cary et al., 1978), but the hydrologic characteristics of the frozen soil remains unpredictable, in spite of a few measurements (Burt and Williams, 1976). Yet, long ago Post and Dreibelbis (1942) and Trimble et al. (1958) pointed out the crucial difference between dense "concrete frost" that renders even vegetated soils almost impervious, and "porous frost" which does not lower the infiltration capacity. The occurrence and characteristics of soil frost are known qualitatively to depend upon weather, vegetation, soil characteristics, and snowpack depth, but a quantitative understanding and prediction of the hydrologic role of soil frost requires some field investigations guided by theoretical studies. Until this is done, and our knowledge of water flow paths is more secure, the calibration of watershed models of snowmelt runoff will rest on a weak foundation.

Where the flux of melt water from the base of the pack exceeds the infiltration capacity of the soil, water accumulates as a thin saturated layer and migrates downslope. The response of this saturated flow to melt was documented by Dunne and Black (1971) in Vermont, U.S.A., but nothing of general significance was known about the process until Colbeck (1974) extended his kinematic approach to the computation of this lateral saturated flux. Dunne et al. (1976) then applied the combination of Colbeck's methods for unsaturated–saturated flow to the prediction of hillslope runoff hydrographs from homogeneous snowpacks in subarctic Canada. Analytical solutions to the problem were sufficient in this case, but the data and computing requirements escalate rapidly as a result of inhomogeneities in snowpack depth and hydraulic properties, and hillslope shape. At this level of complexity, predictions are better obtained with the numerical models of unsaturated and saturated flow, such as Beven and Dunne (1981) applied to the computation of runoff from a Vermont hillslope. The role of such calculations is then reduced to that described earlier for physically-based models. They cannot be incorporated into routine predictions of snowmelt floods, although they can be used to structure simpler, empirical predictive relationships that reflect at least approximately the physics of runoff (E.A. Anderson, 1978).

Where the infiltration capacity remains high, all meltwater enters the soil and generates subsurface flow. This situation has been the subject of a field and modeling study by Stephenson and Freeze (1974). Weekly measurements of snowmelt, distributed on a daily basis by calculation were used as input to the model, which predicted the slow seasonal rise and fall of subsurface flow. The modeling exercise was carefully described so that the difficulties of applying computer models even to thoroughly instrumented research sites could be clearly seen. Beven and Dunne (1981) were able to use a finite-element method to reproduce the general features of hydrographs generated by a combination of subsurface flow and overland flow through the base of the pack over concrete frost on a Vermont hillslope (Dunne and Black, 1971). Both processes responded to daily melt cycles. The density of the frost lowered the infiltration capacity of the soil to less than 0.25 mm $hr.^{-1}$, which forced most of the runoff to travel through the base of the pack.

After a less severe winter, concrete frost was not so well developed, and the infiltration capacity of the Vermont hillslope was higher. A greater proportion of the meltwater entered the soil and raised the water table to the surface over the lower part of the hillslope. Return flow occurred over a saturated zone that expanded and contracted daily (as shown by measurements in wells). The return flow was augmented by percolation onto the saturated zone to produce saturation overland flow through the base of the snowpack.

In summary, it is now possible to predict roughly which runoff process or processes will dominate the snowmelt hydrograph if frost conditions and the associated infiltration capacity are known. It is possible with a small amount of fieldwork to assess the infiltration capacity at the melt season, but much field and theoretical work is needed before frost and its hydraulic

properties can be predicted. Both Horton overland flow and subsurface flow during snowmelt can be predicted for simple situations, but physically-based computations for more frequent natural situations require large amounts of field data and computer capacity.

RELATIONS BETWEEN FIELD STUDIES AND MODELING

Understanding and prediction of runoff processes would be enhanced by closer cooperation between scientists working on field studies and those interested in modeling. Bridging of this unfortunate gap in the science requires a change in the way hydrology is conducted and supported.

TABLE II

Examples of issues that might profitably be studied through a combination of fieldwork and modeling in the near future

Horton overland flow regimes:

(1) Critical field measurements that would allow theoretical developments to draw together disparate information on the effects of: (a) soil crusts; (b) sparse vegetation; (c) microtopography; and (d) cracking soils on infiltration
(2) Realistic description and measurement of the nature of overland flow disturbed by raindrops; adequacy of the sheetflow approximation
(3) Improved description of flow resistance and the Reynolds number characterizing the transition between laminar and turbulent flow
(4) Methods for describing representative regional hillslope profiles with characteristic sequences of topography, microtopography, surface properties affecting infiltration and flow resistance, and vegetation density; computation of representative runoff conditions for typical assemblages
(5) Methods for quantifying the variance of factors around the typical trends referred to in (4)

Variable-source regimes:

(1) Measurement of moisture—pressure—conductivity relations, pore pressure fields, and water flux during field investigations of subsurface flow
(2) Statistical description of representative regional hillslope profiles and associated soils; computation of representative runoff conditions for typical associations
(3) Methods for quantifying the variance of factors around the typical trends referred to in (2)
(4) Study of the hydrologic significance of topographic hollows, based on their frequency and range of form within a physiographic region
(5) Methods for representing saturation overland flow and quantifying flow resistance

Snowmelt regimes:

(1) Measurement of moisture—pressure—conductivity relationships, grain size, and density for natural snowpacks
(2) Study of the effects of snowpack stratigraphy on meltwater percolation
(3) Measurement of the distribution, controlling variables, and conductivity of frozen soils
(4) Modeling of the influence of various spatial patterns of snow depth, snowmelt, and continuous and discontinuous soil frost on snowmelt runoff

Field research needs to be pursued vigorously in a wider range of environments, such as snow regions, tropical rainforests, and disturbed temperate forests where land management is raising some important issues the resolution of which requires an understanding of runoff mechanisms. Field research will continue to unearth a broader range of conditions than are currently acknowledged, such as the extremely high conductivities documented by Harr (1977) and Mosley (1979). It will also emphasize mechanisms such as crust formation and flow in macropores that need to be incorporated into runoff theory.

However, field experiments and data collection programs need to be designed in the light of rigorous physically-based models. At present, there are too many field investigations which describe a particular situation without any hope of making quantitative generalizations beyond the study site. Field experiments cannot be conducted on every combination of even the major variables controlling runoff. Thus, it is necessary to extend the few obtainable data on the basis of a firm foundation of physical theory. If field studies were routinely designed around a rigorous theory, it is more likely that critical variables would be measured, which now usually are avoided because of the difficulty of measurement or because the fieldworker has neither insight into nor interest in their use for modeling to generalize his results.

For personnel on the modeling side of this gap, there is a need to present models, computational details, and their data requirements more clearly, and to become involved in fieldwork, at least in the planning and data analysis phases. There is much to be gained from more frequent discussion with field hydrologists about mechanisms that are not treated in current models, but which continue to engage the attention of field workers. Table II lists some suggestions for cooperative work between the Cains and Abels of hydrology.

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SOME PROBLEMS IN TRANSFERRING HYDROLOGICAL RELATIONSHIPS BETWEEN SMALL AND LARGE DRAINAGE BASINS AND BETWEEN REGIONS

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ABSTRACT

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Several studies are described that provide information on some problems involved in transferring hydrological relationships between small and large drainage basins and between regions. The studies illustrate the fact that basin processes vary considerably from one region to another, and over small distances within one region. These processes include the type and characteristics of storm runoff, small-scale variations of infiltration and runoff, spatial aspects of nonlinear flood response, annual runoff relations and channel transmission losses. Examples are also given of problems in identifying similarity relationships resulting from interaction of model parameters with basin size, and from random and systematic errors in available hydrological data. The results indicate the need for better knowledge of actual hydrological processes at basin scale.

INTRODUCTION

The validity of the transfer of research results from small to larger basins, and from one region to another, is of obvious practical importance. A large proportion of hydrological research has been carried out on small basins, but larger basins are generally of greater practical importance, and it is here that the results of research need to be applied for efficient design. Also, many of the published research results have been obtained in limited regions of a relatively small number of countries, and there is a practical need to apply the results on a wider geographical basis.

While the question of similarity of hydrological response at basin scale is thus of great importance, very few publications directly address the problem on the basis of field evidence of hydrological processes. There is an abundance of literature that is of indirect relevance, including those studies dealing with the development of all types of regional relationships. However, most of these studies are of an empirical nature. Vorst and Bell (1977) reviewed a wide range of regional relationships relating to both flood parameters and runoff volumes. They found that only three basin variables were consistently significant in predicting hydrological characteristics. These were

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