RECOGNITION AND PREDICTION OF RUNOFF-PRODUCING ZONES IN HUMID REGIONS

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Abstract. Field experiments on representative and experimental basins have demonstrated the validity of the variable-source concept of storm runoff production in humid regions. Storm runoff is produced on small portions of a catchment which vary in extent during and between storms. In order to make the variable-source concept useful for flood prediction, water quality management, and land planning, it is necessary to develop routine methods for recognizing and quantifying the seasonal and in-storm variation of the saturated runoff-producing zones. The paper suggests various hydrological, pedological and vegetative indicators of this variation. They can be used for prediction of the maximum seasonal extent of saturated conditions, and also for day-to-day accounting, probability studies, and for analysing the probable effect of some environmental change. The techniques are illustrated with examples from representative and experimental basins in the United States and Canada.

La reconnaissance et la prévision des zones qui produisent l'écoulement dans les régions humides

Résumé. Des essais sur le terrain sur les bassins représentatifs et expérimentaux ont démontré la justesse du concept d'une source variable pour le ruissellement de surface direct dans les régions humides. Le ruissellement de surface direct est produit sur des portions étroites d'un bassin qui changent leur étendue pendant et entre les orages. Pour utiliser ce concept pour la prévision des crues, pour le contrôle des propriétés caractéristiques de l'eau, et pour faire les projets de l'utilisation des terres, il faut développer des méthodes pour reconnaître et quantifier la variation saisonnière et pendant-orage des zones saturées qui produisent cet écoulement. Cette mémoire suggère quelques indicateurs hydrologiques, pedologiques et végétatifs de cette variation. On peut les utiliser pour la prévision de l'étendue maximum-saisonnière des conditions saturées et pour leur explication de jour en jour, pour les études statistiques et pour l'analyse des effets probables de quelque changement du milieu natural. On a expliqué les méthodes avec des exemples des bassins représentatifs et expérimentaux aux Etats Unis et au Canada.

INTRODUCTION

Runoff processes

Since the 1930's the Horton (1933) infiltration approach to runoff production has dominated hydrology and its applications to the prediction of river discharges (Soil Conservation Service, 1972; Crawford and Linsley, 1966) and in land management (Schwab *et al.*, 1966; Hudson, 1971). The Horton model of overland flow has been confirmed by repeated field observation and detailed hydraulic study (e.g. Emmett, 1970). This form of runoff is known

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to occur in semiarid regions and on agricultural lands such as those of the Midwestern United States. On these lands infiltration capacities are usually less than the majority of rainfall intensities.

Because the infiltration capacities of soils on a catchment are rarely uniform, the production of Horton overland flow varies spatially. This is the basis of Betson's (1964) partial-area model of storm runoff, which is based on the original Horton analysis of overland flow, but which suggests that only a small portion of some catchments contributes storm runoff. Routine techniques have been developed for quantifying the pattern of infiltration capacities for purposes of predicting runoff or for recognizing the runoff-producing areas of special interest in land management. The techniques vary from direct field mapping with a sprinkling infiltrometer, to indirect evaluation by hydrograph analysis or from soil texture, vegetative cover and management practice (Musgrave and Holtan, 1964). Statistical prediction of infiltration capacity is then made possible by correlating measured or estimated infiltration capacity with antecedent rainfall.

In humid regions, the infiltration capacity of the soil remains high unless the dense vegetation cover is disturbed. Hence, Horton overland flow is confined to such locations as roads and parking lots, skid trails in forests, some ploughed fields, artificial fills, and other areas that have been denuded of their vegetation. Overland flow can also occur under snowpacks in areas where the infiltration capacity is lowered by the presence of concrete frost in the soil. Where overland flow is generated because of low infiltration capacity in humid regions, the runoff-producing areas are often clearly identifiable and the infiltration capacity can be evaluated routinely as indicated above.

In humid regions that have not been severely disturbed, Horton overland flow does not occur. At least three processes generate storm runoff in these regions, and their relative importance varies with topography, soil antecedent wetness and storm size. These sources of storm runoff, illustrated in Fig. 1, are *subsurface stormflow, return flow,* and *direct precipitation onto saturated areas.*

Subsurface stormflow occurs where rainwater saturates some zone of a permeable soil and displaces soil water to the stream at rates which make an important contribution to the storm hydrograph in the channel. Experimental field studies of this process and its relationship to the stream hydrograph are described, for example, by Whipkey (1965, 1969), Dunne (1969a, 1976), Dunne and Black (1970a, b), Hewlett and Nutter (1970), and Weyman (1970).

The importance of return flow and direct precipitation on saturated areas was suggested by Dunne (1969a, 1976) and Dunne and Black (1970a, b) from field studies in northeastern Vermont. They occur when subsurface stormflow is unable to remove all the incoming rainwater, and the consequent increase in the amount of water stored in the soil raises the water table to the soil surface in swales and on the lower parts of hillslopes. Subsurface water can then emerge from the soil surface as return flow*, and run overland at much greater velocities than are possible for subsurface stormflow. Rain falling on the saturated area also runs off as overland flow and is termed direct precipitation on saturated areas (see Fig. 1). These last two processes may be grouped together under the title, *saturation overland flow.* Rain falling on other parts of the basin is stored within the soil. A small portion of it displaces water as subsurface stormflow and a larger portion supplies baseflow between storms (Hewlett, 1961).

Where soils are well-drained, deep and permeable, and steep hillsides border a narrow valley floor, subsurface stormflow dominates the hydrograph volumetrically, but emerges from the ground surface over only limited zones of the catchment. The area that can supply saturation overland flow, therefore, is small. Contributions of direct precipitation to the saturated valley floor and small amounts of return flow may produce sharp hydrograph peaks,

* Musgrave and Holtan (1964) used the term *return flow* for water returning from beneath the soil to the ground surface on a hillside plot.

but the generation of saturation overland flow is limited by the narrowness of the valley floors and the fact that the soils are deep, permeable and situated on steep hillslopes, Several examples of the importance of subsurface flow have been described for such an area in the Southern Appalachians by Hewlett and Nutter (1970). In many other regions, where the saturated and near-saturated valley bottoms are more extensive, and where footslopes are gentler, and soils thinner, saturation overland flow is much more extensive, and becomes the primary contributor of storm runoff. Moreover, the importance of saturation overland flow increases with storm size (Dunne, 1969a, 1976; Dunne and Black, 1970a, b).

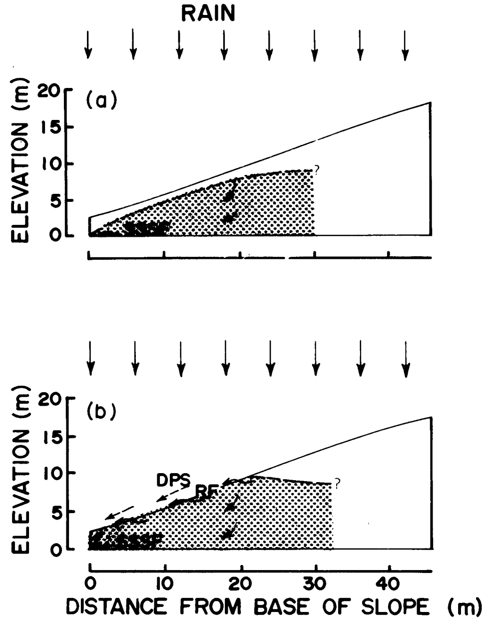

Fig. 1 – Field measurements of runoff processes during a 4.4 cm, 2-h storm on a well-drained hillside.
(a) Early in the storm the water table (dashed line) is approximately at its pre-storm position and subsurface stormflow (SSSF) has just begun. Flow lines, based upon piezometric measurements are shown by arrows in the saturated zone (shaded).
(b) Late in the storm the water table has risen to intersect the ground surface over the lower portion of the hillside. Subsurface stormflow has increased. Return flow (RF) and direct precipitation onto the saturated area (DPS) produce saturation overland flow. The saturated area expands upslope during the storm.

The saturated area which produces most of the storm runoff in humid regions varies through time, both during and between rainstorms. The water table rises to the soil surface over an expanding area as rainfall progresses. Saturation overland flow spreads first up the previously dry, low-order tributary channels, then up unchannelled swales and gentle footslopes of hillsides, especially those covered with shallow, poorly-drained soils. After the end of a storm, this saturated area contracts much more slowly as the soil drains. A conceptual model of this process was developed by the US Forest Service (US Forest Service, 1961; Hewlett and Hibbert, 1967) and called the *variable source area concept.* A similar conceptual model developed by the Tennessee Valley Authority (1965) was termed the *dynamic watershed concept.*

This model was field tested and extended by Ragan (1968) and Dunne (1969a), who presented measurements of the rates of runoff production by each process, as well as the associated changes of piezometric head and soil moisture content at various locations within small catchments. There is now a general consensus of opinion that in humid regions storm runoff is generated on relatively small areas of the catchment, and that these areas vary during and between storms.

Need for recognition and prediction of saturated zones

We have already indicated that the recognition and statistical prediction of areas producing Horton overland flow are routine. It would be valuable to make the recognition and prediction of saturated areas in humid regions equally routine. The size and location of these variable contributing zones are of interest for predicting rates or volumes of runoff in humid regions, either by physical modelling techniques or by reinterpreting traditional methods such as the unit hydrograph, the Φ-index, time-area diagrams, or the rational formula. Use of the simplest techniques may require only an estimate of the total area saturated at the beginning and end of a storm, while other methods will require estimates of the location and width of the saturated zone and the depth and travel time of saturation overland flow.

There are other reasons for knowing which areas of a basin yield saturation overland flow. The source of runoff in relation to sources of contaminants such as phosphates or bacteria is an important control of water quality (Dunne, 1969b; Kunkle, 1970). A knowledge of the areas of a catchment that produce saturation overland flow would allow major 'non-point' source areas of various contaminants to be delineated and perhaps steps can be taken to minimize such inputs. Better methods of predicting the location, magnitude and frequency of ground saturation at various times of the year would also improve the analysis of land suitability for ploughing, planting and harvesting. Houses are still built where saturation overland flow occurs from time to time, and so the routine recognition of variable sources would help in zoning to eliminate problems with septic tanks and other housing needs.

Here, we make some suggestions of methods that are being used to recognize and predict the size and location of variable saturated areas. The methods are not perfected yet, and we have not investigated all their possibilities. We offer them at this point as a stimulant to further discussion and trial, in the hope that soon the delineation of areas producing saturation overland flow will become as routine as the mapping of infiltration capacity in regions experiencing Horton overland flow.

The runoff-producing zone varies both during and between storms. The inter-storm and seasonal variation is easier to document and predict. We will deal less intensively with the more difficult problem of the expansion of the saturated area during a storm. The inter-storm variation of the saturated zone not only affects land capability as described earlier, but is, to a large extent, the antecedent moisture status of the catchment, which is an important control of the volume and rate of runoff.

SEASONAL AND INTER-STORM VARIATION OF THE SATURATED ZONE

The best method of evaluating the size, location and variation of the saturated zone either within or between storms is by repeated field mapping. The investigator can quickly identify the extent of saturated soils while walking through a small drainage basin, and he can plot their outline on an aerial photograph or plane-table map. We have tried taking photographs (both black-and-white and colour infrared) to record zones with differing degrees of wetness. Our efforts were not very successful, but others with more expertise in remote sensing might develop the technique for application over large areas. Since mapping on the ground can only be done occasionally and for small catchments, the field measurements must then be correlated with some other, easily observable, characteristic of the basin to allow mapping over larger areas, and prediction over time. The basin characteristics that we have found most useful are topography, baseflow, an antecedent moisture index and soils. Vegetation is also useful as a rough indicator in some areas.

Figure 2 is a map of the saturated zone at three seasons of the year 1967-1968 in a small catchment in northeastern Vermont. The basin is developed in a Pleistocene sandy kame terrace overlying a silty sandy lake deposit (Dunne, 1969a). The steep, straight sideslopes are covered by generally deep (0.6-6.0 m), well-drained sands and sandy loam soils. These soils are not saturated between storms, except at four places where the Pleistocene deposits are

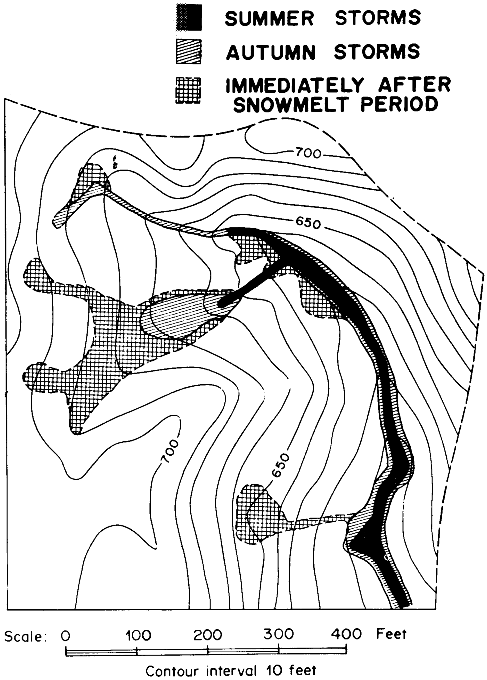

Fig. 2 – Seasonal extent of the saturated zone before summer storms, autumn storms, and immediately after the snowmelt season in a catchment with steep well-drained slopes and a narrow valley bottom. Basin WC-4, Sleepers River Watershed, Danville, Vermont.

thin and springs emerge from the underlying bedrock. On the gentle, lower sideslopes of the catchment, the silty sandy lake deposit is exposed and soils are shallow, poorly drained silt loams, which become saturated during wet periods of the year. The valley floor lies on a dense silty glacial till and in the lower part of the catchment is perennially saturated. The saturated zone varies from 3 to 11 per cent of the basin area. Figure 3 shows the variation of the saturated zone in a nearby catchment with gentle hillslopes and concave footslopes covered by poorly-drained to well-drained loams. The extent of the saturated zone and its variation from 15 to 51 per cent of the basin area are much larger than in the steep, well-drained catchment shown in Fig. 2. Also the expansion takes place not only by headward extension of the channels into swales, but also by lateral expansion up the hillslopes of the basin. This lateral expansion is important not only because of its magnitude and therefore its flood potential, but also because it affects water quality and land capability (Kunkle, 1970).

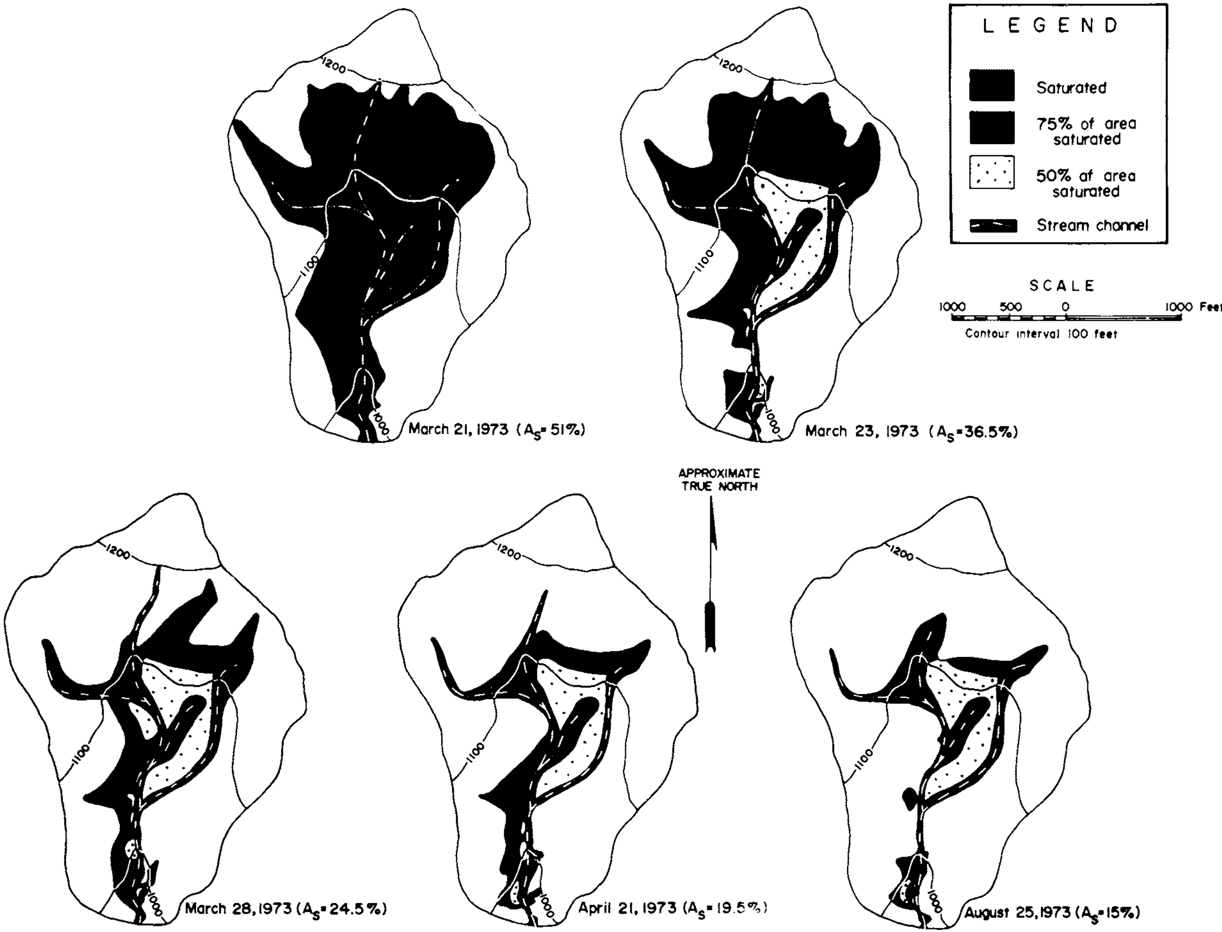

Fig. 3 – Seasonal variation of the saturated zone in a catchment with extensive, gentle footslopes mantled with shallow soils. Basin W-2, Sleepers River Watershed, Danville, Vermont.

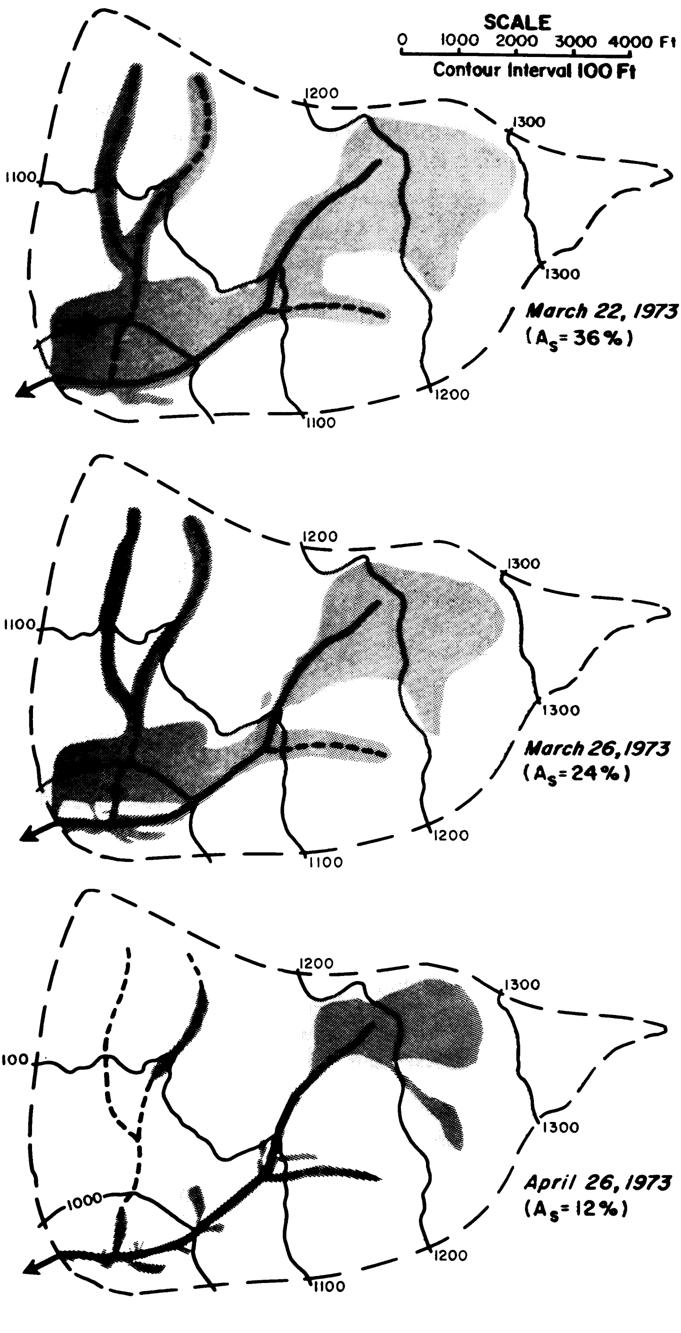

Fig. 4 – Seasonal variation of the saturated zone in a catchment at Randboro, Quebec.

Figure 4 shows the variation of the saturated zone in a small catchment in glacial till and schist at Randboro, Quebec. Figure 5 shows the variation in a small inter-drumlin swale at Peterborough, Ontario.

This part of our work has shown that for catchments with areas as large as 5-8 km², it is possible to define the seasonal or inter-storm variation of the saturated zone by repeated field mapping, and that the variation is large.

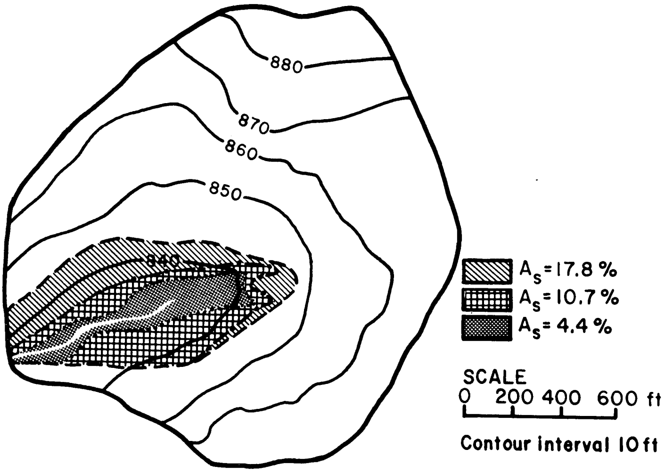

Fig. 5 – Seasonal variation of the saturated zone in an inter-drumlin swale near Peterborough, Ontario.

Relation to topography

The most obvious feature emerging from these maps is the influence of topography. The saturated contributing zones are generally located in valley bottoms and swales throughout most of the year, and extend further up the swales and up gentle footslopes at the wettest times of the year. In the steepest catchment (Fig. 2) the maximum extent of saturation is only 11 per cent of the basin area; in the less steep W-2 basin (Fig. 3) it is 51 per cent. In regions where the underlying rock structure exerts a strong control over soil depth, and the emergence of groundwater, the saturated runoff-producing zones may not be confined to low-lying parts of the catchment. Working in a part of Ohio where runoff is largely controlled by alternating strata of varying permeability, Amerman (1965, p. 504) found that 'Runoff-producing areas were located in seemingly random fashion on ridge tops, valley slopes, and valley bottoms.' Where such control is absent, however, there is a greater probability of developing a saturated zone on low-lying ground with a considerable drainage area above it to supply seepage throughout the year.

It is difficult to develop quantitative predictions from topography, however. Troeh (1964) has suggested the use of polynomial equations to describe topography for the purpose of relating it to soil drainage, and this may be useful for some purposes, but is unlikely to become a routine technique on the scale of a catchment.

Relation to soils

Soil morphology is useful in indicating the distribution of saturated areas in a drainage basin. When soils become saturated, the free oxygen in the profile is consumed by soil organisms and reducing conditions develop. Anaerobic micro-organisms are able to gain energy by converting oxidized elements such as iron and manganese into their reduced forms. The oxidized forms are insoluble, whereas the reduced forms are more soluble and are able to move through the soil. The mobilized iron and manganese are removed from the reduced sections of the soil and either lost from the soil profile or reprecipitated in oxidized sections of the soil. Since iron and manganese are important colouring agents in soils, the reduced sections become pale grey and the oxidized sections show a brown or reddish brown colour (Crompton, 1952). This grey and brown mottling, or gleying, is characteristic of soils subject to seasonal waterlogging.

Brown mottling in the dark brown surface horizon of soils may be useful in estimating the distribution of the saturated area of a drainage basin. In general, the more extensive the brown mottling, the longer the period of saturation (Moore, 1974). This relationship may be obscured in some places because the development of reducing conditions and the mobilization of iron and manganese are dependent on many factors such as temperature, time, and the amount of organic matter available to the microbial population. Reducing conditions develop in saturated organic-rich soils in the laboratory in 2 or 3 days, but it may take up to 2 weeks under field conditions where soils are cold in the winter and spring (McKeague, 1965; Meek *et al.,* 1968). Saturated areas in some parts of the drainage basin may exist for only a week or two each year, and gley morphology is unlikely to be very distinct in these zones. Variations in vegetation and land use practices may also disrupt the establishment of a relationship between the water regime and the morphology of the surface horizons. In many areas, however, soil morphology is a useful indicator of the zones that remain saturated for periods of one to several weeks.

Subsoil gley morphology is a more useful indicator of the probability of surface saturation because it reflects the height of the water table and hence the likelihood of the water table rising to the soil surface for short periods in response to heavy rainfall. Several studies have shown that there is a correlation between soil-water regime and quantitative indices of subsoil gley morphology (e.g. Crown and Hoffman, 1970; Latshaw and Thompson, 1968; Moore, 1974; Simonson and Boersma, 1972). Since the monitoring of the soil-water regime is time consuming, soil profile morphology is used as a criterion of soil drainage (Canada Department of Agriculture, 1970; Soil Survey Staff, 1967).

The depth at which mottling occurs in a soil profile reflects the depth of the water table for periods of about 2 months or more during the season in which saturated areas are likely to be most extensive (Daniels *et al.,* 1971; Simonson and Boersma, 1972). Reducing conditions develop more slowly in subsoil horizons, because of the low availability of organic matter, so that gley morphology reflects longer periods of saturation than in the surface horizon. Furthermore, the development of the gley morphology is affected by the colour of the parent material, and the presence of fossil gley morphology is related to previous soil-water regimes (Daniels *et al.,* 1973; Moore, 1974; Schelling, 1960). From an estimate of the water table depth and the water holding characteristics of the soil, a prediction can be made of the area that is likely to be saturated under certain rainfall conditions (see later).

Comparison of Figs. 2 and 6 shows that there is a generally good correlation between the maximum seasonal extent of the saturated zone and the area of 'somewhat poorly drained soils', as mapped by the US Department of Agriculture, Soil Conservation Service. The saturated area ranges up to 11 per cent of the basin area, while the 'somewhat poorly drained soils' cover 15 per cent. There are no major discrepancies between the two maps, though minor ones occur near springs. The degree of correspondence is particularly encouraging because the original soils map was compiled for a different purpose from aerial photographs at a scale of 1 : 8000, while the map of the saturated zone was compiled by field surveying at a scale of 1 : 800.

Figure 7 is a soil map compiled by the Soil Conservation Service for the basin shown in Fig. 3. Comparison of the two figures shows that at the season of maximum snowmelt runoff, the saturated area expands into those 'well drained soils' that are on gentle gradients, but that it soon retreats to the zone covered by 'moderately and somewhat poorly drained

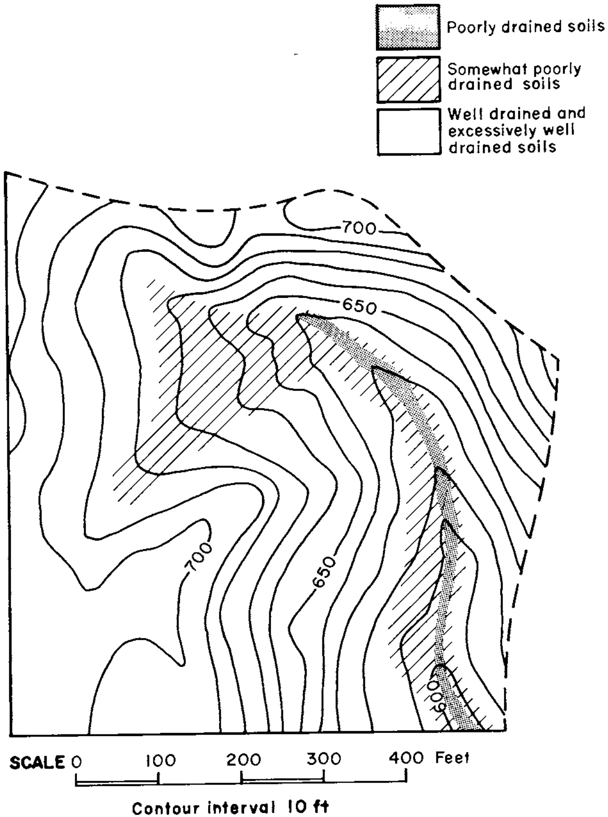

Fig. 6 – Soil drainage map of the WC-4 catchment shown in Fig. 2.

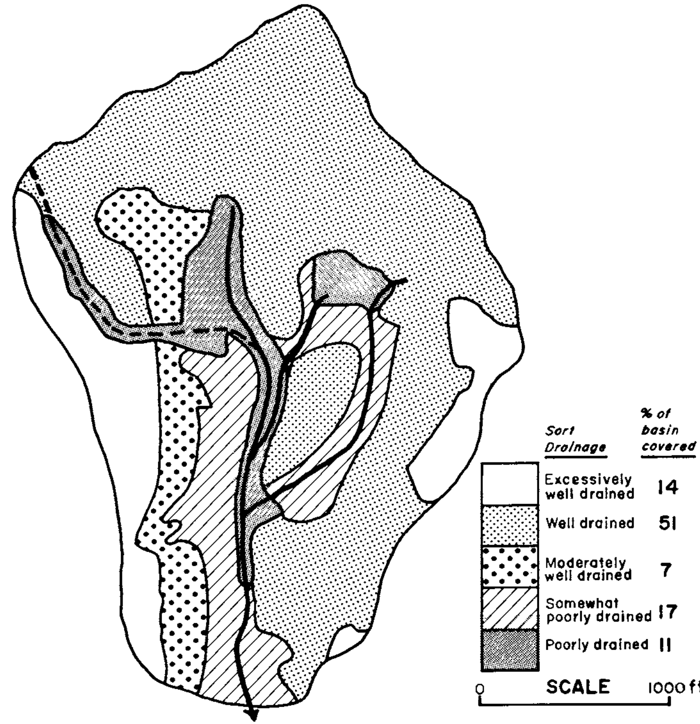

Fig. 7 – Soil drainage map of the W-2 catchment shown in Fig. 3.

soils'. The saturated area during summer agrees quite closely with the zone of 'poorly drained soils'. Considering that the maps were compiled at widely differing scales, and that Fig. 3 was produced without looking at the soils map, the maps agree surprisingly well, even in detail.

Relation to vegetation

In some regions it is possible to use plants as indicators of soil drainage and runoff-producing zones, using the methodology of 'physiographical ecology', as illustrated by Hack and Goodlett (1960), Zimmermann (1967), and Zimmermann and Goodlett (1975). The use of plant indicators of soil wetness should be approached with care (Zimmermann, 1970). Plants reflect only broad regimes of soil moisture rather than specific values at one season of the year. They also reflect regimes of other factors such as light or nutrients. The history of land use in a region may also complicate distributions. In northeastern Vermont, for example, the abandonment of fields has allowed a spread of northern white cedar, white spruce, balsam fir and other conifers to well-drained sites that before colonization were occupied by northern hardwoods (Zimmermann, 1967). Another complication is the fact that an indicator of wetness in one region may not indicate the same conditions elsewhere. The role of wetness in limiting plant distributions may vary regionally. In an area such as Vermont, soil moisture may be limiting at germination during April, May and early June. In Maryland, where vegetation receives approximately the same annual precipitation but suffers a severe moisture deficiency during the summer, the limiting factor may be the moisture status of the soil during late summer.

In spite of the difficulties encountered, however, plants may be useful indicators of soil wetness within a region. Zimmermann (1970) suggests that the presence or absence of individual tree species are the most useful indicators. This is fortunate for most routine hydrological applications because individual tree species may be mapped rapidly over large areas from the ground or sometimes from aerial photographs.

In northeastern Vermont, for example, northern white cedar, white spruce and balsam fir that are not located on old-field sites outline approximately the maximum extent of the saturated zone in most years. During very wet years and at the base of long slopes the saturated zone may extend into a hardwood cover which includes ash, butternut, yellow birch and basswood. Paper birch, beech, and hop-hornbeam, which are common in the area do not occupy wet zones, except for an occasional individual on a blowdown mound (Zimmermann, 1967). The minimum, late summer extent of the saturated area is outlined quite clearly by sedges, with occasional willows and elms.

Relation to hydrological parameters

The previously described techniques of prediction can be used to estimate seasonal maxima or minima. For a day-to-day accounting of the extent of saturation, some hydrological technique involving a routinely measured parameter is necessary. Figure 8 shows a correlation between baseflow and the extent of the saturated zone for the two Vermont drainage basins and the Peterborough basin described previously, indicating that baseflow is a good predictor of the saturated zone. The nature of the relationship varies with the topography and soils of the catchment. In the steep, well-drained catchment (WC-4) the saturated zone is always small and cannot change by very much. In the basin with gentle hillslopes and moderate to poorly drained soils (W-2) the variation of the saturated zone is large, especially when baseflow exceeds 0.07 m^3 s^{-2} km^2. Several hydrologists (Gregory and Walling, 1968; Blyth and Rodda, 1973) have shown that the length of flowing streams in a basin varies with stream discharge. Because the saturated area can spread laterally by large amounts, however (Fig. 3), and because the areal extent of saturation is of interest in many hydrological problems, the correlation in Fig. 8 is more useful. With the aid of such a correlation, streamflow records can be used to make statistical studies of the duration of saturated conditions (see Fig. 9), or

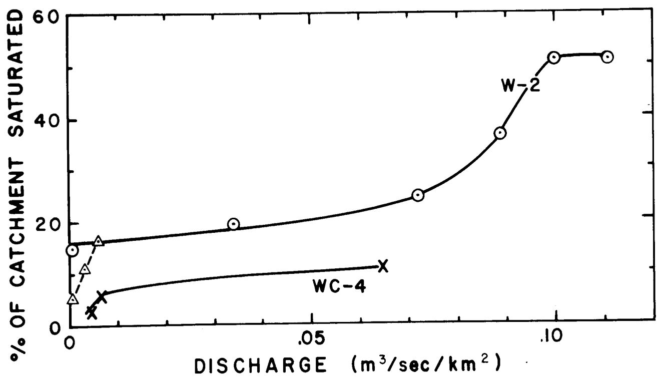

Fig. 8 – Relation between baseflow and the extent of the saturated area for the two Vermont catchments, W-2 and WC-4, and the Peterborough catchment (dashed line)

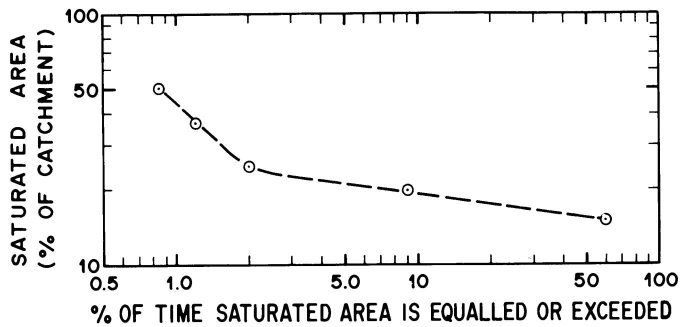

Fig. 9 – The probability of occurrence of saturated zones of various sizes in the W-2 catchment, Sleepers River Watershed. The diagram was constructed using Fig. 8 and a flow duration curve for the basin.

of the probability of various degrees of saturation occurring before design storms in any month of the year.

In the absence of stream records, one may use a water balance calculation (Thornthwaite and Mather, 1957) and relate the area of saturation to the calculated value of moisture surplus or deficit. Alternatively, the independent variable may be an antecedent precipitation index of the form

$$I_t = I_0 K^t$$

where

I_t = the antecedent precipitation index of the catchment on day t (cm);
I_0 = the initial value of the antecedent precipitation (in cm) at some time of maximum saturation, such as the end of the snowmelt season, when t is set equal to zero. I_0 is usually chosen by trial and error;
K = a constant, characteristic of the catchment, or perhaps of all catchments with similar soils and topography in a region;
t = time (days).

When rainfall is added to the basin, its depth is added to the value of I_t for that day. The measured extent of the saturated zone can then be correlated with the computed value of I_t for the day of measurement, as shown in Fig. 10. It is likely that an antecedent precipitation index could be used to indicate the conditions under which soils of each drainage class become saturated in a particular region.

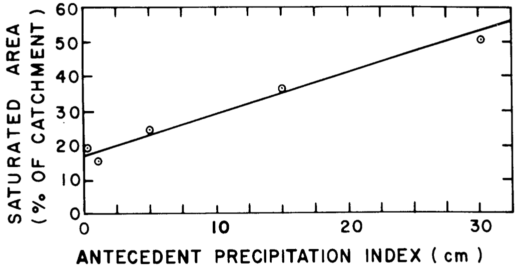

Fig. 10 – Correlation of the mapped extent of the saturated area in catchment W-2 with the calculated antecedent moisture index, using values of $I_0 = 30$ cm, $K = 0.80$.

Other hydrological prediction techniques that seem appropriate involve correlation of the saturated area with some other parameter of moisture such as soil moisture levels, or water table elevations. Dickinson and Whiteley (1970) found that soil-moisture measurements could be related to an estimate of the saturated area. We have used the water table elevations in an index well for the same purpose in the small, well-drained catchment described earlier (Fig. 11). Then, to the extent that water table elevations or soil moisture levels can be predicted from meteorological and pedological characteristics, the extent of the saturated zone can be computed on a routine basis. Boersma (1967) has shown the feasibility of predicting water table fluctuations by a water budget procedure, and Nelson *et al.* (1973) have used meteorological variables to do the same thing by multiple regression. When the day-to-

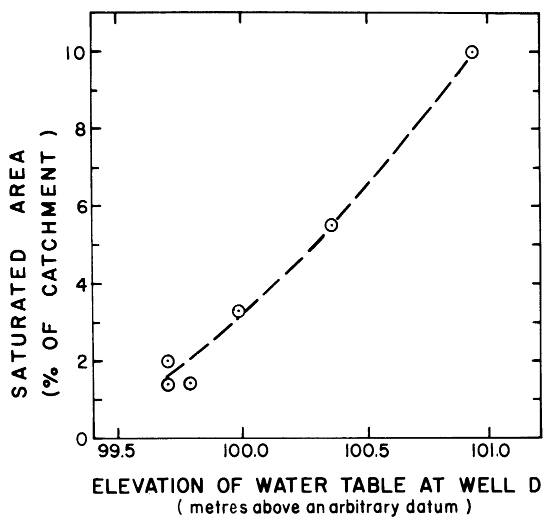

Fig. 11 – Relation between the extent of the saturated area and the elevation of the water table at an index well on the WC-4 catchment, Sleepers River Watershed.

variation of the saturated area can be related to streamflow, soil moisture, water table elevations, or meteorological variables, a long-term record of the zone of saturation can be generated and subjected to probability analysis. Estimates of the seasonal variation of average or maximum saturated areas would be useful for many purposes in hydrological design, land capability classification and planning. Figure 12 shows such seasonal variation for a short record in a Canadian catchment. Longer records would allow the construction of frequency distributions for each month.

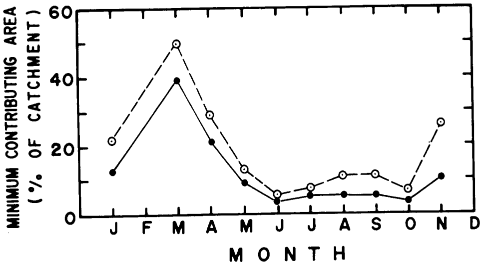

Fig. 12 – Monthly mean (●) and maximum (⊙) contributing areas for a 18 km2 catchment in southern Ontario. Compiled from data in Dickinson and Whiteley (1970).

VARIATION OF THE SATURATED ZONES DURING STORMS

Field measurements indicate that, as predicted by the conceptual models of the US Forest Service (1961) and Tennessee Valley Authority (1965), the saturated zone expands during a storm. This expansion can be mapped by direct observation or by interpreting the records from lines of wells across a catchment. Figure 13 shows a map constructed by the former method, and Dunne (1969a, 1976) has presented maps and cross sections of the expansion recorded by wells and field mapping during sprinkling experiments on plots in the W-2 and WC-4 catchments at Danville, Vermont. The same processes have been observed in the Peterborough catchment.

Such mapping can only be done for relatively small catchments, and is unlikely to become routine, though it is an instructive exercise to make such a map for a small catchment before making hydrological predictions for a region. A few lines of wells for indicating maximum water table heights would help in the construction of such a map. Betson *et al.* (1968) have described simple, easily installed instrumentation for recording maximum water table elevations. Attempts are also being made to monitor the maximum expansion of the saturated area in the Peterborough area by sinking weighted plastic cups into the ground in a line normal to a stream channel. The cups are covered to prevent direct precipitation access, but have small openings in the side at ground level so that if the water table rises to the surface, or if saturation overland flow occurs around a cup, it will fill with water. The outermost filled cup in the transect line indicates the maximum lateral extent of the saturated area during a storm. We have also mapped the maximum extent of the saturated zone by placing large numbers of stakes around its edge during the last few minutes of a storm. The stakes can later be mapped with greater speed, accuracy and comfort. A map of this kind for a large storm or for a snowmelt season gives an indication of the topographic zones, soil types, and vegetation assemblages into which the saturated zones may extend during rare periods of high runoff.

Prediction of storm runoff rates from variable sources is not an easy task, however, and at this time there is no routine method of calculating the expansion of the saturated zone during a storm or of calculating the runoff produced by each of the processes described earlier. The choice lies between sophisticated mathematical models with demanding information requirements and rough approximations based on simplification of field conditions.

Freeze (1972a, b) has presented a mathematical model of runoff generation in small catchments by saturated and unsaturated subsurface flow, return flow, and direct precipitation on saturated areas. The model was tested in a field situation during a snowmelt season (Stephenson and Freeze, 1974). In view of the difficulty of obtaining the necessary input data, however, the Freeze model remains a research tool, rather than a technique for routine prediction of the saturated zone or the runoff produced therein.

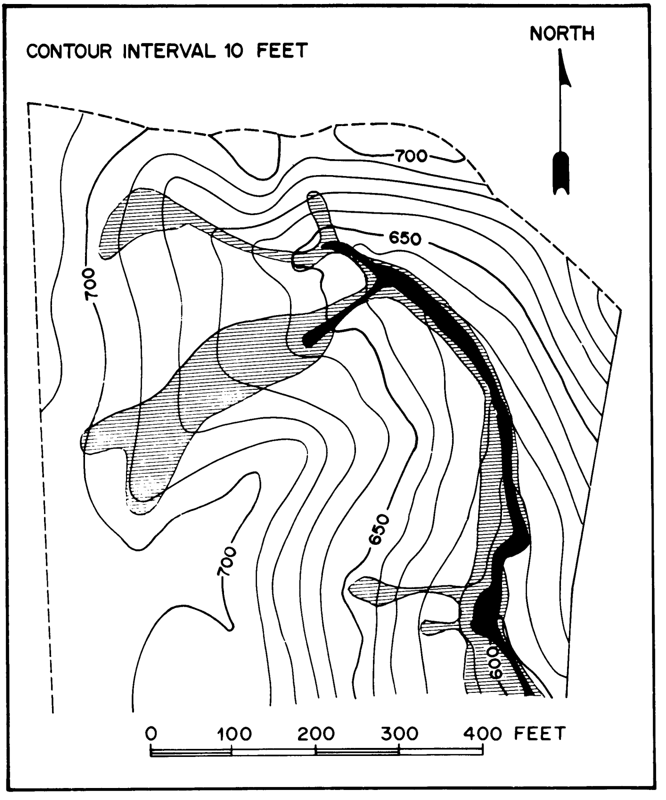

Fig. 13 – Expansion of the saturated area during a 4.6-cm rainstorm on the WC-4 catchment, Sleepers River Watershed. The solid area indicates the saturated zone at the beginning of the storm; the lined area indicates the expansion.

The Tennessee Valley Authority (1965, 1970) have estimated the contributing area at various times during a storm by dividing the measured volume of storm runoff by the volume of precipitation during that time interval. Direct precipitation on saturated areas is therefore assumed to be the dominant contributor of stormflow, while 'interflow and storm baseflow' are assumed to make up 'delayed runoff'. The contributing area was then estimated from the accumulated rainfall and the rainfall intensity (1965) or from rainfall volume alone (1970). Lee and Delleur (1972) have used essentially the same technique with hydrographs from Indiana streams. Dickinson and Whiteley (1970) also simplified the problem by computing the minimum contributing area from measured stormflow and rainfall, thereby defining storm runoff as direct precipitation on saturated zones, ignoring return flow and subsurface stormflow contributions. They correlated this calculated area with a basin moisture index composed of the sum of the antecedent moisture in the upper 114 cm of soil at an indicator site and one-half of the precipitation during the storm.

Similar calculations have been made for a number of other drainage basins in southern Ontario. In these calculations direct runoff volumes under the rising limbs of floods were divided by precipitation totals falling before the time of peak discharge, corrected for litter storage, depression storage and interception losses. This concentration on only the rising segment of a peak is admittedly arbitrary but was done because a considerable proportion of runoff on the falling limb would be supplied by return flow and subsurface flow and would therefore not be directly related to the extent of the saturated area. In other words using only the rising segment would yield a *minimum* runoff from direct precipitation on saturated areas. Runoff/rainfall ratios were calculated in this way for Kawartha Heights Stream near Peterborough for 11 rainstorms in October and November 1973. The range of values, from 0.77 to 4.91 per cent for storms of 0.29 to 0.93 in., compares very closely with the estimates of minimum and maximum saturated area measured by direct field mapping over the same time period of 0.5-4.9. Similar calculations were made for two adjacent basins near Orillia, Ontario. The average runoff/rainfall ratio for two storms was 0.99 per cent for a basin of 7.5 mi^2 and 0.74 per cent for a basin of 10 mi^2. These ratios compare quite well with the percentage saturated area estimated from field mapping: 0.88 per cent for the smaller basin and 0.46 per cent for the larger. This is an encouraging result because these basins are larger than others discussed in this paper.

Computation of the minimum saturated area from observed hydrographs is subject to several objections. It ignores the depression storage within the saturated zones themselves, as well as contributions from return flow and subsurface stormflow. Nevertheless, it remains the easiest method currently available for approximating the contributing zone and, though circular, the method is convenient for predicting stormflow volumes for use with the unit hydrograph or similar techniques. The method, however, gives only a minimum extent of saturated conditions, and perhaps more important, provides only a lumped value for a catchment. For several purposes, as described in the introduction, it would be useful to know the location within the catchment of zones that become saturated during a storm of a chosen magnitude. To predict these locations one has to employ information from Soil Survey Reports and topographic maps, and rough approximations to hydrological processes.

The problem is one of outlining those soils whose storage capacity becomes exhausted after various amounts of rainfall. These areas can then be added to the pre-storm saturated zone, outlined by techniques described in the preceding section.

The storage capacity of a soil at the beginning of a rainstorm can be roughly estimated from the depth of the water table and the porosity and field capacity of the soil. It is obvious that the distribution of soil moisture above the water table varies with depth, and that it may be depleted near the surface by evapotranspiration. If this kind of information is available for the soils of interest, it can be used, but for routine computations on most soils it will not be. Especially when critical wet seasons and large storms are being considered, the assumption that the whole soil profile above the water table is at field capacity seems an adequate approximation. Where the water table is close to the ground surface a soil moisture value larger than the field capacity might be used, or the capillary fringe might be assumed to intersect the ground surface over the lower portion of the hillslope. If the field capacity and porosity have not been measured, they can be estimated from a graph such as Fig. 14. The depth of the water table at the time of the design storm may be evaluated from field observations, by computations as previously described (Boersma, 1967; Nelson *et al.,* 1973), or by examination of the gley morphology of the soil profile as outlined in the previous section. For critical periods of high moisture levels, the depth of the seasonal high water table is listed for each soil series in the county Soil Survey Reports (e.g. Soil Conservation Service, 1973). When a range of depth, say 30-60 cm, is listed in these reports an interpolation can be made between a 30-cm depth at the lower end of the soil series and a 60-cm depth at the upper end (Fig. 15 (a)).

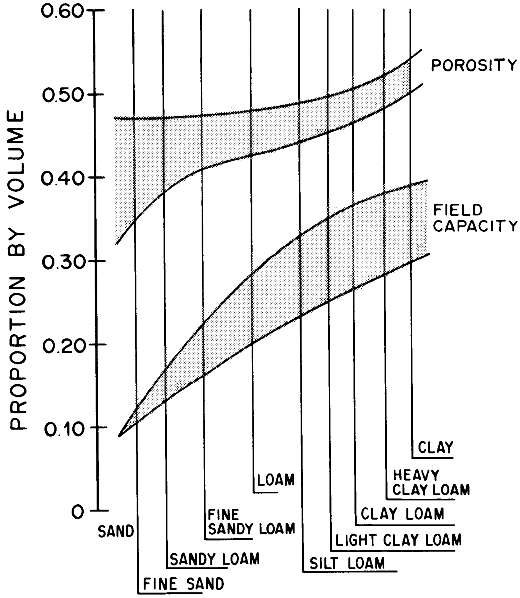

Fig. 14 – Suggested ranges of porosity and field capacity for soils of different textures. Based upon a diagram by US Department of Agriculture (1955).

As an example, let us consider a soil series shown in Fig. 15. A 150-cm deep Sammamish silt loam consists of an upper 75 cm of silt loam with a saturated hydraulic conductivity of 1.5-5.8 cm/h (geometric mean = 0.9 cm/h). The soil covers the lower parts of hillslopes with gradients of approximately 10 per cent and the minimum depth of the seasonal high water table is 30-60 cm (Soil Conservation Service, 1973). If a design calculation is to be made of potential runoff rates during seasons of high antecedent moisture conditions, the pre-storm situation can be roughly approximated for a 30-m long footslope, as shown in Fig. 15(a). Assuming that the water table is planar, and that soil moisture in the vadose zone is at field capacity, the storage potential of the soil at any distance downslope can be calculated from the depth of the water table, the porosity and the field capacity (0.45 and 0.32 respectively from Fig. 14). The variation of storage potential with distance along the slope is shown in Fig. 15(b). Pre-storm baseflow per metre width of hillside (Q) can be estimated from the saturated cross-sectional area of the soil, the saturated hydraulic conductivity, and the gradient of the water table as

$$
\begin{aligned}
Q_{\text{topsoil}} &= 0.45 \text{ m} \times 0.035 \text{ m/h} \times 0.04 = 0.00055 \text{ m}^3\text{/h/m} \\
Q_{\text{subsoil}} &= 0.75 \text{ m} \times 0.009 \text{ m/h} \times 0.04 = 0.00027 \text{ m}^3\text{/h/m} \\
Q_{\text{total}} &= Q_{\text{topsoil}} + Q_{\text{subsoil}} = 0.00082 \text{ m}^3\text{/h/m}
\end{aligned}
$$

Assuming an average drainage density for humid, hilly regions of 6 km/km^2, the pre-storm baseflow from the soil can be estimated as

$$Q_{\text{baseflow}} = 0.00082 \text{ m}^3/\text{h/m} \times 2 \times 1000 \text{ m/km} \times 6 \text{ km/km}^2$$
$$= 9.8 \text{ m}^3/\text{h/km}^2 \text{ or } 0.003 \text{ m}^3/\text{s/km}^2$$

Such a calculation does not give the amount of baseflow being supplied from deeper groundwater bodies. This must be estimated separately, if it is important.

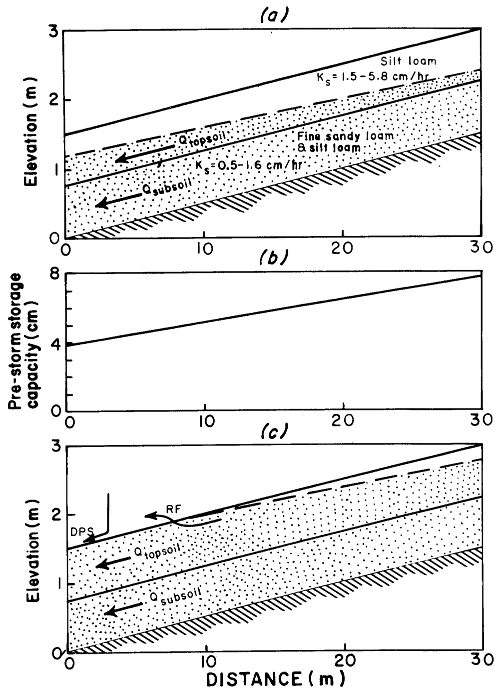

Fig. 15 – Simplified runoff conditions for computing the approximate runoff from a Sammamish silt loam on a 1 in 20 slope during a 5-cm rainstorm.

The reaction of the water table to a design rainstorm of (say) 5 cm/h for one hour can be estimated by assuming that in wet shallow soils, the water table rises almost immediately when rain enters the soil, and that the rate of downslope subsurface flow is small relative to the rate of rainfall. There is field support for the first approximation in the work of Ragan (1968) and Dunne (1969a), and for the second in the results of various experiments on subsurface storm flow in which the percentage yield of subsurface stormflow during a rainstorm is small (see Dunne, 1976 for summary). With these approximations the rainfall would fill the storage capacity of the soil and the water table would rise to the surface during the design storm over the lower 9 m of hillside, as shown in Fig. 15(c). Subsurface flow would increase to approximately 0.003 $m^3/h/m$, or 0.01 $m^3/s/km^2$, or to some larger value if the drainage density increased significantly during the storm. The increase of drainage density could also be estimated in the same way by examining the soils in the bottoms of tributary valleys with ephemeral channels. Even if the drainage density were to double, however, the subsurface stormflow would be small relative to other contributions.

The greatest change in runoff rates would result from the onset and spread of saturation overland flow due to direct precipitation on saturated areas, as portrayed in Fig. 15(c). At the end of the storm, this runoff rate per metre width of hillside would be

$$Q_{DPS} = 0.05 \text{ m/h} \times 9 \text{ m} = 0.45 \text{ m}^3/\text{h/m} \ (1.5 \text{ m}^3/\text{s/km}^2)$$

These calculated rates of spreading and of saturation overland flow are minima because in a soil as fine-textured as a silt loam, the capillary fringe would be very close to the surface at the base of the hillside and even less rainfall would be needed to saturate the profile. The procedure can be refined by a few simple calculations to incorporate such conditions.

A comparison can be made with a nearby Puyallup fine sandy loam on a 30-m long hillslope with a gradient of 1 in 5. According to the Soil Survey Report, the depth of the seasonal high water table is 120-150 cm as shown in Fig. 16(a). The baseflow from such a soil would be approximately 0.017 m^3/h/m of hillside or 0.057 m^3/s/km^2. At least 30 cm of rain would have to fall before the water table reached the ground surface at the base of the hillslope. Further rain would cause an upslope expansion of the saturated zone and overland flow contributions. Subsurface stormflow would increase to approximately 0.056 m^3/h/m or 0.187 m^3/s/km^2, which is a reasonable value for deep sandy loams on steep slopes (Dunne, 1975). Similar approximate calculations for different combinations of topography, soils and rainfall indicate that the rate and volume of stormflow increase with storm size and hillslope length, and decrease as depth to water table increases. These results are in agreement with field observations. Runoff production from various parts of a soil catena can be considered in the same way. Antecedent conditions can also be varied if the water table depth and soil moisture status can be estimated for other seasons, but in most design problems, one is concerned with extreme conditions.

As a field test of the simplified procedure, the amount of rain required to induce saturated conditions during a 2.18 cm/h storm was calculated for the hillside and storm shown in Fig. 1. The calculated requirements ranged from 1.98 cm at the base of the slope to 3.45 cm half way up the slope and 9.65 cm three-quarters of the distance upslope. The measured values were 1.90 cm at the base, and 3.81-4.06 cm at the intermediate location. The soil at

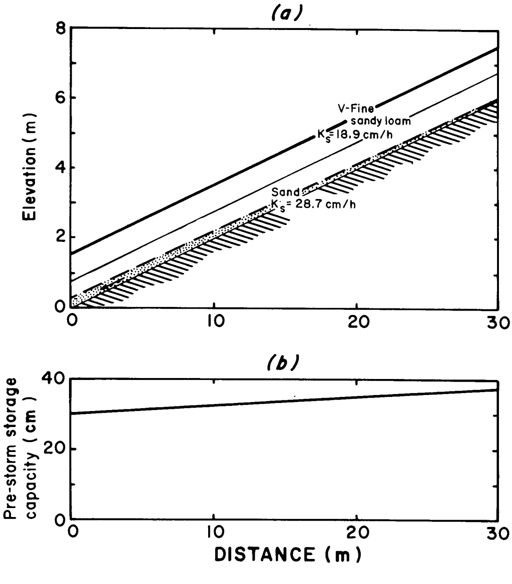

Fig. 16 – Pre-storm conditions on a Puyallup sandy loam on a 1 in 5 slope, showing the magnitude of the storage potential.

the highest location did not become saturated during this 4.4 cm storm, but another storm of 8.23 cm brought the water table at this station to within 30 cm of the ground surface. Similar calculations for two other plots were equally successful.

The most difficult process to quantify is the rate at which return flow emerges from the soil over the lower saturated zone. At present there is no obvious way of simplifying this calculation. A few field measurements may give some idea of the range of the return flow contribution. On a steep hillside swale covered with a sandy loam soil (shown in Fig. 1), we have measured return flow contributions to saturation overland flow ranging from 72 per cent of peak rate in a storm with an intensity of 2.18 cm/h to 42 per cent of peak rate in a storm with an intensity of 8.00 cm/h. On a gently sloping silt loam, return flow contributed approximately 50 per cent of both the runoff rate and total volume of saturation overland flow from storms of 1.52 cm/h and from 1 to 3.5 h duration.

As an indication of how large the return flow contributions might be, we used data from a 100-year storm on the Coweeta Hydrologic Laboratory in the southern Appalachians. Hewlett and Nutter (1970) presented data on runoff from a sandy loam on a 60-m long hillside plot during the 50.9 cm rainstorm. Hewlett and Helvey (1970) listed rainfall and runoff from small basins during the same event. Using values of porosity, field capacity, gradient and soil depth given by Hewlett (1961) and Hewlett and Nutter (1970), and a reasonable estimate of 30 cm/h for permeability, it is possible to calculate that the 2.13-m deep soil on the hillside plot requires at least 53.2 cm of rain to exhaust its storage capacity. Therefore no saturation overland flow should result from the 50.9 cm storm. This is in agreement with the field observations. If the soil was originally at field capacity the 50.9-cm rainstorm should have raised the water table to within 9-10 cm of the surface and the peak rate of subsurface stormflow should have been 0.21 $m^3/h/m$, which is equivalent to 1 $m^3/s/km^2$ (or to 0.5 $m^3/s/km^2$ if the hillside plot is half as long as an average hillslope in the basin as the authors describe). The authors do not give a measured value for this peak rate, but for two storms of 10.2 cm, and 14.3 cm on the same hillside their measured values are within 4 and 6 per cent of the values predicted by our simple method.

The value of 1 $m^3/s/km^2$ approaches the limiting rate at which water can move through a 213-cm deep sandy loam forest soil, since the saturated permeability cannot be much higher than 30 cm/h (see Dunne, 1976 for a summary of field measurements). Yet in a storm with a recurrence interval of more than 100 years Hewlett and Helvey (1970) report peak runoff rates from small Coweeta basins of 2.4 to 2.7 $m^3/s/km^2$. In order to generate runoff at this higher rate there must have been considerable contributions from return flow and direct precipitation on the small saturated areas at the base of hillslopes described by Hewlett and Nutter. The total of these contributions must have been approximately 1.5 $m^3/s/km^2$. It is not possible to separate the contributions exactly, but an approximation can be made by using the peak rainfall intensity (2.03 cm/h) for the storm, given by Hewlett and Helvey. If the whole 1.5 $m^3/s/km^2$ (0.54 cm/h) originated as direct precipitation on saturated areas, the extent of this saturated area would have to be 0.54/2.03 = 27 per cent of the basin area. By comparison with Figs.2, 8 and 13, this figure seems too high for basins like those at Coweeta with deep well-drained soils, steep hillslopes, and narrow valley bottoms. If return flow contributed 50 per cent of the saturation overland flow (as on the steep sandy loam plot of Fig. 1 in a large storm) the saturated area would cover only 13.3 per cent of the basin which is a reasonable value for the type of terrain. Until more data become available it seems appropriate to assume that return flow contributions can approximate those of direct precipitation on saturated areas during large storms.

AREAS FOR FUTURE RESEARCH

Quantitative understanding of the variable-source concept is far from complete and its application to hydrological calculations is not developed. More field observations are needed on the interrelationships of subsurface stormflow, return flow and direct precipitation on saturated areas. We have already pointed out the dearth of measurements of return flow and the lack of any simple usable technique for calculating its contribution to the hydrograph.

Another poorly understood process is saturation overland flow itself. It is usual to assume that 100 per cent of this water reaches the stream instantaneously. Yet the thick mat of vegetation and the roughness of the ground surface reduce both the velocity and the yield. On a hillslope with a gradient of 1 in 2½, we have measured velocities of overland flow ranging from 3 to 15 cm/s. On hillslopes with gradients below 1 in 20, velocities less than 0.15 cm/s are common. On a rough marshy area we have measured 2.54 cm of depression storage, 1 cm of detention storage and mean overland flow velocities (by salt injection) of 0.15 cm/s. There is then a considerable opportunity for storage of saturation overland flow within runoff producing areas, and this mechanism needs to be quantified.

Further work would also be useful on techniques for estimating the seasonal variation of the saturated zone. We have suggested some hydrological and pedological techniques which might be developed. The use of vegetative indicators is another possibility, but its use is fraught with many difficulties.

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