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THE UNIVERSITY OF NEW SOUTH WALES

WATER RESEARCH LABORATORY

REPORT No. 63

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A Survey of Watershed Yield

by

H. D. Ayers

MARCH, 1962

THE UNIVERSITY OF NEW SOUTH WALES

SCHOOL OF CIVIL ENGINEERING

WATER RESEARCH LABORATORY

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A SURVEY OF WATERSHED YIELD.

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H.D. AYERS.

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FOREWORD

The assessment of the unregulated yield from a catchment is an everyday problem confronting the hydrologic engineer developing rural water supplies. Owing to the lack of sufficient basic research, he is forced in practice to employ empirical and arbitrary methods. This leads to waste and inefficiency in rural development. Recognising this weakness in water engineering, the Yield Sub-Committee of Technical Committee No. 6 (Hydrology) of the Institution of Engineers, Australia, urged that as a first step there should be prepared a comprehensive review of existing knowledge in this field.

Professor H.D. Ayers, Head of the Agricultural Engineering Department of the Ontario State Agricultural College, Guelph, Canada, during his sabbatical year with the School of Civil Engineering of the University of New South Wales, kindly consented to attempt this task, so far as the ungauged catchment problem is concerned, and the results of this survey are presented herein.

One difficulty which besets the engineering profession is that reports such as that of Professor Ayers are pigeon-holed in the University or Department in which they were prepared and do not reach a wide circle of readers. However, it is fortunate that in this case there did exist a Research Grant for "Rainfall and Runoff" studies, made available to the School of Civil Engineering of the University of New South Wales by the Rural Credits Development Fund of the Reserve Bank of Australia. Therefore money was available from this fund to enable this Report to be published and circulated widely amongst the professional public. This help of the Reserve Bank is gratefully acknowledged.

It is hoped that this discussion by Professor Ayers and the associated bibliography will be of assistance to practising engineers confronted with the problem of water supplies in rural areas for agricultural development and will also inspire research workers in Australia and elsewhere to develop lines of attack which will enable the practising engineer to approach these rural water supply problems on a more rational basis.

C.H. MUNRO
Professor of Civil Engineering
The University of New South Wales

I N D E X

A SURVEY OF WATERSHED YIELD.

A SURVEY OF WATERSHED YIELD.

1. INTRODUCTION.

1.1. DEFINITION

In the generic sense yield refers to the quantity of any product resulting from exploitation of natural resources. Specifically there are several definitions in current use in the hydrological literature. The Sub-group of the Hydrological Group of the Institution of Water Engineers of the United Kingdom, Law (1955), provides the following definition: "The uniform rate at which water can be withdrawn from a reservoir throughout a dry period of specified severity without depleting the contents to such an extent that withdrawal at that rate is no longer feasible." It will be noted that yield in this context refers to reservoired watersheds, so that it is a property of the watershed <u>and</u> the reservoir.

Johnstone and Cross (1949) state that yield is often used synonomously with streamflow, discharge and runoff. Wisler and Brater (1959) suggest that yield is usually considered in terms of the total volume of streamflow for periods of time of a year, or the average flow for periods of a month or more, while discharge and runoff are reserved for flow rates over short time periods.

Harrold (1957) used yield synonomously with stream flow to designate the quantity of water available for use on the land surface at the point on the stream at which it is measured, when periods of a month or year are considered.

Steel (1947), defines yield as that portion of precipitation on a watershed that can be collected for use, including direct runoff and water which first passes underground before appearing as streamflow.

The term yield is frequently qualified by such adjectives as minimum, safe and dependable. These expressions lack precision and should only be used when accompanied by a definition and specification of probability limits.

It is suggested that the term 'yield' be reserved to designate the total discharge normally expressed as an area-depth occurring during a specific time period of one month or longer duration <u>or</u> as a result of a specific storm event, on a particular watershed. It is important also that the value stated be qualified further as being the mean, median, minimum, maximum or by appropriate probability limits for the particular duration of the event.

As used herein the term yield will be restricted to the natural runoff from a watershed, so that artificial regulation will not be a factor. In this discussion, however, such indirect measures as land-use changes which may exert an effect on the water balance, will be considered.

1.2 IMPORTANCE

Water requirements for all purposes have increased rapidly in the industrially advanced areas of the world in recent years. Competition for water for irrigation, all types of manufacturing industries, domestic consumption, calls for a better understanding of the occurrence, distribution and limitations of surface water supplies.

Water use of a non-consumptive nature such as in the generation of hydro-electric power and for transportation is of a high economic value to a country. Its effective utilization for these and other purposes depends upon a satisfactory understanding of the yield characteristics of the drainage basins.

While stream gauging networks are extensive in certain countries of the world and records extend back for several decades on larger watersheds there are many regions where virtually no stream flow measurements exist. Many small watersheds and tributary basins will never be gauged, even in the United States. It is important therefore that experimental and analytical studies be pursued so that engineering project designs for water utilization will proceed upon a sound basis. This is important not only in areas where water use already plays an important part in the technological and social life of a country, but is equally important to the economic development of the newly emerging nations of the world.

1.3 SCOPE

The survey is limited to a review of the principles governing the generation of streamflow and a discussion of experiments and investigations concerning the interrelationship of the significant factors in the hydrologic cycle. Only a brief mention will be made of the common methods of analysis and interpretation of the results of stream flow measurements. Statistical methods applicable to storage studies will be omitted entirely because its very nature would justify a separate report.

Since flows of low magnitude are of extreme importance special attention will be given to the contribution of ground water to stream flow. Flood flows may form a high percentage of the seasonal water yield from some watersheds. Emphasis will be on volumes of water yielded rather than on instantaneous rates of discharge.

Precipitation as a direct factor affecting yield will be covered, in addition to the evaporation process as it relates to the water balance and the disposal of precipitation into stream flow or soil moisture storage.

Attention will be directed towards the effects of the physiographic and geological characteristics of watersheds on yield. The land-use and vegetative community will be discussed in the light of investigations into their effect on yield.

Reference to published literature will be made frequently, although an attempt will be made to avoid a repetition of material contained in the several fine text-books on Hydrology.

The survey is intended to focus attention on the present status of knowledge in this particular field. Criticisms will be advanced, which are in no way intended to reflect upon the authors cited. Rather it is hoped that fresh thought and effort may be stimulated in a very important field.

2. STREAMFLOW - DATA COLLECTION, ANALYSIS AND INTERPRETATION.

2.1 THE RUNOFF PROCESS

2.1.1 Origins of stream flow.

The water supply for stream flow is the precipitation incident upon the watershed. However, only a portion of the precipitation reaches the outlet of the watershed, the remainder eventually evaporating or leaking to neighbouring watersheds. The route followed by an element of water may be over the ground surface throughout its entire history within the watershed. The portion of streamflow which is derived in this manner is called surface runoff. It occurs during and immediately following intense rains or during periods of snow melt.

Many streams are supplied in part from bodies of ground water. The ground water is replenished from the surplus of water which infiltrates the soil surface after storage in the surface mantle of soil is satisfied. Ground water flow is very slow, so that there is a considerable lag between precipitation or snow-melt and the ground water flow component of stream flow. The ground water component is also known as base-flow and accounts for the entire flow when direct runoff ceases.

A further component of streamflow during storms is known as interflow. Some of the infiltrating water flows laterally at a shallow depth, due to the presence of rock or less permeable soil. This water often seeps into shallow rills or channels after a short distance of flow below the surface. At this point it is indistinguishable from surface runoff. At any one time particularly during and after a storm, stream flow may consist of components of surface runoff, interflow and ground water flow. The distinction between the paths of flow is to a large extent academic, however it is a useful basis for the separation of time variation in flow components and for the classification of types of streams.

2.1.2 Types of Streams.

Three types of streams are recognized.

Ephemeral streams are those which flow only as a result of surface runoff. They are typical of the natural drainage systems of arid or semi-arid regions in cases where soils of low permeability predominate.

Intermittent streams flow during wet seasons, but cease to flow after a short period with low rainfall. Ground water contribution is from aquifers of limited extent, which are capable of maintaining stream flow for very short periods only.

Perennial streams flow continuously even through long dry periods. Ground water flow sustains flow at such times, while substantial quantities of surface runoff may occur at times of high intensity storms.

2.1.3 The runoff cycle.

The march of events associated with the runoff cycle is by no means a consistent one. Hoyt (1942) presented a comprehensive description of the cycle and the phases as they related to the nature of stream flow. Another description is outlined in Wisler and Brater (1959). At best, a description is idealized or simplified. Most watersheds are so complex that there are several phases of the cycle in operation simultaneously, so that although it provides a basis for understanding the phenomena, it is beyond the scope of this survey to discuss the subject in detail.

2.2 STREAMFLOW DATA.

2.2.1 Collection.

The collection of streamflow data generally follows a demand for information to facilitate development and utilization of water resources. Seldom has there been adequate foresight to ensure that a long and representative period of observations on streams preceded their development. To a large extent the streams of the semi-arid regions of North America and Australia received first attention, since it was in these regions that water was of the utmost importance for irrigation and security of settlement. Streams in the humid regions where water was taken for granted as being in abundant supply, were neglected until public pressures focused attention on the problems of flood protection and town water supplies.

The nature of the observations initially was in the form of daily or weekly observations of river stages at selected points on large or important streams. By the use of current meter observations a stage-discharge relationship was developed so that the rate of flow could be determined. Local observers, such as farmers or ranchers were relied upon to make the observations of river stage. Many streams today are still observed in this manner, with daily or twice daily observations being common.

A more complete record is provided by automatic water stage recorders. Normally these consist of a float suspended by a cable over a pulley. The water level in the stream is recorded through the response of the float and pulley which actuate a pen providing a graphical record of stage. Installations of this type are expensive so are not used exclusively.

A less expensive installation is used in Australia to a considerable extent. Water levels are recorded through a pressure sensitive element inserted in the stream and connected to a remotely installed 'Bristol Recorder'. In either case the charts are usually changed by a local observer and forwarded to the appropriate government agency for tabulation.

2.2.2 Types of Data.

The raw data having been collected are tabulated in the form of flow rates. Corrections are made to the raw data to account for changes in stream section and the possible effects of temporary pondage or diversion, so that before publication the observations are as near as possible to the actual flow in the stream.

The form of data presentation varies with the agency providing the service. The U.S. Geological Survey publishes daily rates of flow in cubic feet per second, mean discharge for each month and acre-feet per month.

The location of gauging station, type of observation and area of drainage basin are also given.

The period of the stream flow record is also given as are the maximum and minimum flows with their dates. Similar types of information are provided by the Water Resources Branch, Canada Department of Northern Affairs and National Resources in publications each covering two years of observations.

The States of Victoria and Queensland have been publishing data for several years on monthly minimum, maximum and mean stream flow in cubic feet per second or acre-feet. In addition Queensland includes the corresponding rainfall and runoff in area-inches for certain typical catchments from 1921.

The State of New South Wales has just commenced publication of past records also as monthly values.

Although stream gauging is a Federal function in the United States and Canada, it is done co-operatively with State or Provincial agencies. On the other hand, stream gauging is a State responsibility in Australia except in the Northern Territory. The quality and availability of published data varies widely from State to State. In general the publication of streamflow data by States has been badly neglected.

2.2.3 Reliability of Data.

A number of factors can reduce the reliability of published data, even though every effort is made to make corrections. As for unpublished data it is seldom subjected to the same critical examination, so that errors are more prevalent. Scouring and sedimentation of the gauging section are causes which can be detected, although the exact time of the change may be difficult to pinpoint. The use of double mass curve types of analyses however can facilitate the detection of the time of change of the stage-discharge relationship.

Reliance upon unskilled observers to make observations contributes a degree of uncertainty to the data. Occasions when observations have been missed mean incomplete records which reduce the value of the data.

2.3 ANALYSIS AND INTERPRETATION

2.3.1 Flow-duration Curves.

These curves indicate accumulatively the percentage of time during the period of record that indicated rates of flow were equalled or exceeded. Mean daily, weekly or monthly flow rates are commonly selected and plotted against the percentage of time that such flows were equalled or exceeded. If a sufficiently long period of record is available the flow-duration curve will be a reasonable representation of the possible future behaviour of the stream.

The unit of time selected for which the mean flow is plotted, is very important. Monthly means may not indicate a period during which the flow is zero even though there may be several days without flow. These occasions may be critical to the particular project under study.

The slope of a flow duration curve when the flow rate is plotted as the ordinate and the percentage of time as the abscissae, provides a useful means for comparing the flow characteristics of one stream with that of another. Where the duration curve is flat a large storage of surface water and ground water is indicated for the watershed. Flat slopes at the lower end of the curve are indicative of high ground water storage slowly but reliably sustaining flow in the stream.

Cross (1949) used flow-duration curves to compare the effects of the geological nature of watersheds in Ohio on dry weather flow. He used as an index the discharge in c.f.s. per square mile, exceeded 90 per cent of the time.

The major disadvantage of the flow duration curve is that there is no reference to the chronological sequence of events. For example one period of three months without flow would plot in the same position as three separate periods of a month each in which there was no flow. The significance to any storage investigations would be quite considerable.

2.3.2 Frequency analyses.

Much has been written about frequency studies in connection with flood flows. The same principles may be applied to low flows. The hope has always been that the discharge plotted against frequency of occurrence would provide a straight line when plotted on logarithmic probability paper. With this relationship, extrapolation for recurrence intervals (reciprocal of frequency) greater than the period of record could be accomplished with a degree of confidence. However, since hydrological data seldom, if ever, have a pattern of normal distribution such an undertaking is not practical. The skewness of data is evident in the case of low flows as well as for peak flows. However, frequency curves are useful in yield studies when average flows for several durations are plotted against frequency or recurrence interval. Cross and Webber (1950) present frequency curves for low flows for durations from one to 183 days (6 months).

2.3.3 Variability Parameters.

Several methods are available for expressing the variation in stream flow from a watershed. The commonest of these is the standard deviation, which is a measure of the dispersion of the observed values about the mean value and is given by :

$$s = \sqrt{\frac{\Sigma (x - \bar{x})^2}{n}}$$

where s = the standard deviation of the sample

$(x - \bar{x})$ = difference between observed value and mean of all the observations

n = number of observations.

The significance of the standard deviation is that in a population with nearly normal distribution the observed value is within the range $\bar{x} \pm s$, approximately two times out of three. The coefficient of variation C, is used to obtain a relative measure of variability.

$$C = \frac{s}{\bar{x}}$$

A stream-flow variability index, I_V was used by Lane and Lei (1950) to compare the flow characteristics of 220 streams east of the 100° meridian in the United States. From flow duration curves of mean daily flow, the discharge rates at percentages from 5 to 95 per cent by increments of ten per cent were selected. The common logarithms of each of the ten flows was determined. The standard deviation of the logarithms was computed for each stream. This value was called the Variability Index. Values ranged from 0.14 to slightly over 1.0. The magnitudes of the indices are small thus providing a simple means for comparing watersheds.

It is claimed by Lane that the Variability Index could be used to synthesize a flow duration curve. However, it would be necessary to make a close estimate of the Index based on watershed characteristics and to have some means of calculating or knowing the median value in the duration series.

There is also the same inherent weakness in the Index as in the flow-duration curve. The chronological order of events is completely ignored.

2.3.4 Mass Curves.

Mass curves and residual mass curves have been widely used in yield investigations. Text books by Linsley, Kohler and Paulhus (1958), Wisler and Brater (1959) and Johnstone and Cross (1949) contain suitable descriptions of mass curve analyses. The assumption inherent in the use of mass curves is that streamflow sequences in the future will be similar to those occurring in the past. With a sufficiently long period of record this may be a reasonable assumption to make. However, even then the possibility should not be overlooked of a sequence of dry years even more severe than anything past records would indicate.

3. THEORETICAL BASIS FOR RUNOFF GENERATION AND STREAMFLOW DISTRIBUTION.

3.1 THE WATER BALANCE

For the earth and atmosphere the total quantity of water is very nearly constant. The variation in time and space of solid, liquid and vapour proportions is great. Even within a single catchment there are continuous changes in the quantity and distribution of water. The hydrologic bookkeeping equation provides a quantitative description of the changes and processes in operation.

Johnstone and Cross (1949) present a rather complete equation:

$$S_1 + S_{s1} + S_{g1} + S_{ss1} + \int_{t_1}^{t_2} I\,dt + \int_{t_1}^{t_2} I_g\,dt + {}_1P_2$$

$$= {}_1E_2 + {}_1T_2 + \int_{t_1}^{t_2} D\,dt + \int_{t_1}^{t_2} D_g\,dt + S_2 + S_{s2} + S_{g2} + S_{ss2}$$

Subscripts 1 and 2 denote the beginning and end of the time period respectively.

S = Volume of water in storage in channels and reservoirs of the area under consideration.

S_s = Volume of water in storage on the surface of the ground on leaves and pavements etc.

S_g = Volume of water in storage as ground water.

S_{ss} = Volume of water in storage as soil moisture.

I = Instantaneous rate of inflow of water to the area above the surface including both channel and overland flow.

I_g = Instantaneous rate of inflow of ground water across the boundaries of the catchment.

${}_1P_2$ = Total equivalent uniform depth of precipitation over the area between time t_1 and t_2.

${}_1E_2$ = Total volume of water evaporated during the period.

${}_1T_2$ = Total volume of water transpired during the period.

D = Instantaneous rate of outflow of water from the area above the surface including both channel and overland flow.

D_g = Instantaneous rate of outflow of ground water across the boundaries of the area.

In this equation the only quantities which can be directly measured with any degree of success are I and D and P. Precipitation may be estimated on an area basis with a dense network of rain gauges. Estimates of the other quantities may be made by indirect means such as by meteorological observations, water table observations, soil moisture sampling, or by inference from the direct measurements.

Over selected time periods certain terms in the equation may be insignificant. For example during a storm the magnitudes of ${}_1E_2$ and ${}_1T_2$ will be small in comparison to the other terms, so can frequently be ignored.

Likewise with the judicious selection of sufficiently long periods of time the storage terms may be equated as follows:

$$S_1 = S_2, \quad S_{s1} = S_{s2}, \quad S_{g1} = S_{g2}, \quad S_{ss1} = S_{ss2}$$

With the catchment as the unit of area, frequently I, I_g and D_g are each very nearly equal to zero.

3.1.1. A Simple Water Budget.

A simplified form of the Water Budget is as follows:-

$$Q = P - E - \Delta S$$

where

Q = net outflow of water from the catchment during a specific time period.

P = Precipitation during the same time period.

E = Evaporation inclusive of transpiration from all surfaces in the catchment during the same time period.

ΔS = Net change in all storage on the catchment during the time period.

Evaluation of the three terms on the right side of the equation will lead to a solution for yield from a catchment.

Two aspects of the water balance computation require significantly different approaches. One involves the disposal of water during a period of precipitation, when the evaporation is relatively small. The other involves the disposal of water during periods of time between precipitation events when evaporation and ground water discharge are important. The two phases are not separable, since the magnitude of the storage changes effected by the evaporation and ground water discharge to the stream will influence the routing of precipitation during a storm and its disposal.

It will be our purpose to examine the major phenomena associated with the hydrologic balance of a catchment; precipitation, evaporation, storage and ground water discharge. Throughout, the effects on storage terms, and in turn the effect of storage levels on the processes will be considered.

3.2 PRECIPITATION

It will be assumed that precipitation is uniform over the catchment in each of its characteristics even though such a model is rarely if ever attained. The efficacy of precipitation in generating streamflow will depend strongly upon the nature of the precipitation and the condition of the catchment surface. The catchment surface will include all materials exposed to the precipitation andto a depth of approximately four feet, or the region of moisture storage and utilization by vegetation.

3.2.1 Form.

The major precipitation forms of significance are snow and rain. Snow normally remains in storage while temperatures are below freezing. Snow is subject to redistribution by winds, changes in density, structural properties, sublimation, and water equivalent between the time it falls and when it melts. The water released during snow melt will be considered in the same manner as low intensity rainfall. Even though the equivalent intensity of melting snow may be low, the usually low soil moisture storage opportunity and low infiltration capacity prevailing at such times frequently results in significant amounts of surface runoff.

3.2.2 Rainfall Intensity.

The rainfall intensity in conjunction with the prevailing infiltration capacity at the moment of contact of water with the catchment surface will determine the immediate path of water. However since in practice intensities change by the minute and infiltration capacities vary as quickly, it is necessary to resort to rather simple models to determine the stream flow component resulting from storm rainfall.

The separation of precipitation into components of direct runoff and losses (soil moisture and ground water accretion, interception, depression storage) is not strictly necessary in connection with yield studies, as is the case for the derivation of unit hydrographs. A portion of the water which infiltrates may recharge ground water to eventually be discharged as stream flow.

A simple model which is quite satisfactory for rains of low intensity is:

$$Q = P - (FC - S_{ssl}) \qquad \begin{array}{l} p < f_c \\ Q \nless 0 \end{array}$$

where p and f_c are respectively rainfall rate and infiltration capacity

FC = Field Capacity of the soil root zone.

The assumption is that all rainfall above that required to recharge the soil root zone is converted to streamflow, either by ground water discharge or direct runoff. If this model is to be accepted the minimum level of ground water storage is that at which the stream just ceases to flow.

The same model may also be used for rains in which $p > f_c$, it still being assumed that all rainfall in excess of that required to recharge the soil moisture to Field Capacity, provides water for stream flow.

In certain cases rains of very high intensity occur in arid regions with sparse vegetation, where the infiltration capacity is low and hence soil moisture recharge is limited. A modified approach is thus adopted:

$$\text{When } p > f_c \qquad Q^1 = (p - f_c)t^1$$

Where t^1 is the duration of time during which $p > f_c$

At times when $p \leq f_c$ the previous relationship holds.

An analogous model is shown in Figure 1 in the manner used by Sugawara (1961). It is assumedthat there is no interflow.

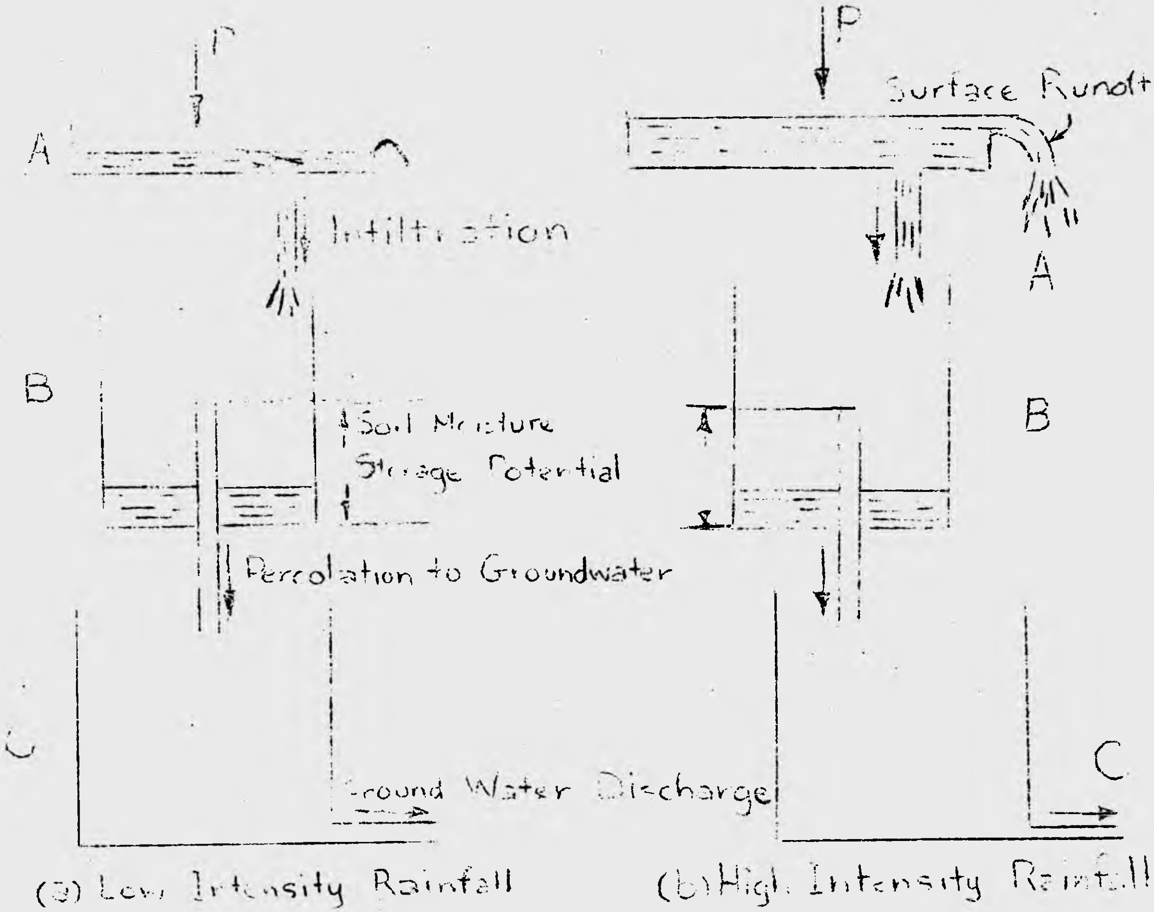


FIG. 1. RAINFALL INTENSITY AND RUNOFF.

In Figure 1(a) the infiltration rate equals the rainfall rate and all rainfall goes to replenish soil moisture. In Figure 1(b), rainfall rate is in excess of the infiltration capacity. The surplus is spilled to surface runoff even though soil moisture may not be completely recharged. In both cases, percolation to ground water starts when soil moisture is recharged.

It should be noted that the models represented are not intended to convey the time variation of the phenomena, but merely the mass effects.

3.2.3 Time of Occurrence.

This is best illustrated again by reference to Figure 1. The rainfall loss to soil moisture storage will depend upon the extent of depletion of soil moisture. Rains occurring at a time when soil moisture is completely filled will be more effective in generating stream flow, than when moisture is depleted. Rains of one inch on successive days will generate more stream flow than when separated by two weeks. The rate of creation of additional storage is a function of the evapotranspiration which has a seasonal variation. This will be discussed in detail later.

3.2.4 Areal Distribution.

Although the original assumption was based on uniformity of precipitation over the catchment, the effects of variations in soil characteristics and precipitation intensity can be accounted for by a model consisting of parallel elements similar to those of Figure 1, each with its own infiltration, soil moisture storage and discharge characteristics.

3.3 EVAPORATION

For our purposes evaporation and transpiration will be considered as one process. The evaporation of transpired water is only a special case of evaporation from a free water surface. Transpiration is limited in many cases by availability of water in the soil. Free water surface evaporation will be designated as E_o, potential evapotranspiration from soil and vegetated area as E_T, and actual evapotranspiration from soil and vegetated areas as E_v (soil moisture limiting).

Since evaporation represents a major item in the water balance equation, detailed discussion of this process is warranted if we propose to develop this approach to yield determinations. Only the more significant features will be presented since the investigation and literature on the subject is voluminous.

3.3.1 Sink-strength Formulas.

Dalton (1802) recognized that one of the necessary conditions for evaporation was that of a sink for absorption of water vapour. In other words the rate of evaporation is proportional to the dryness of the atmosphere. The strength of the sink may be represented by the difference in the vapour pressure of the water and the vapour pressure of the adjacent atmosphere.

Dalton's Law : $E_o = C(e_s - e_a)$ is the basis of most sink strength formulas, where:

E_o = evaporation rate.
e_s = saturated vapour pressure at the temperature of the water surface.
e_a = vapour pressure of the atmosphere adjacent to the water surface.
C = a constant.

The continuity of vapour movement across the water-air interface depends upon some mechanism for transport of vapour away from the evaporating surface. In nature this mechanism is wind. Subsequent to Dalton investigators included a correction for the effects of wind.

Rohwer (1931) after extensive investigations on the evaporation from open water surfaces proposed the following formula:

$$E_o = 0.771(1.465 - 0.0186B)(0.44 + 0.118V)(e_s - e_a)$$

E_o = inches per day.

B = atmospheric pressure in inches of mercury.

V = mean wind velocity in miles per hour at six inches above the water surface.

e_s, e_a are vapour pressures as previously used, expressed in inches of mercury.

A simpler expression was proposed by Meyer (1942) from studies on Minnesota lakes :

$$E_o = C(e_s - e_a)(1 + \frac{V}{10})$$

The factor C has a value of 0.36 when V is measured at an altitude of 25 feet above the water surface.

By assuming a logarithmic variation of wind velocity with elevation the two equations give evaporations at an atmospheric pressure of 30 inches of mercury, respectively of :

$$E_o \text{ (Rowher)} = 0.771\ (0.40 + 0.107V)(e_s - e_a)$$

$$E_o \text{ (Meyer)} = 0.771\ (0.47 + 0.079V)(e_s - e_a)$$

where V is the wind velocity at 6 inches above the surface.

Equivalent values of evaporation will be obtained at a wind velocity of 2.5 miles per hour.

Both expressions are empirical in nature and take into consideration only the mechanism for lateral transport of water vapour. Furthermore the temperature of the water at the evaporating surface must be known. For a free water surface this can be accomplished, but for vegetated areas it presents considerable difficulties.

3.3.2 Aerodynamic Determinations.

By the application of the principles of turbulent diffusion an attempt is made to express the vertical flux of water vapour by considering its similarity to the transport of mass and momentum. The best known expression for evaporation based on this consideration is that credited to Thornthwaite and Holzman (1942).

$$E = \frac{833k^2(e_1 - e_2)(V_1 - V_2)}{(T + 459.4)\log_e (z_2/z_1)^2}$$

E = evaporation in inches per hour.

k = von Karman's constant (0.4)

e_1 and e_2 correspond to the vapour pressure in inches of mercury at a lower and upper level respectively.

V_1 and V_2 are wind speeds in miles per hour at lower and upper levels of elevation z_1 and z_2 in feet.

T is the mean temperature in degress Fahrenheit of the air between z_1 and z_2.

The difficulties in instrumentation in terms of the zero displacement level or elevation z_1 are very great, particularly in its application to large land areas.

3.3.3 Energy Budget.

A supply of energy is necessary to evaporation. Approximately 5000 b.t.u. are required to evaporate a one inch depth of water from an area of one square foot. It is obvious then that an energy budget approach is a logical basis for calculating evaporation.

Cummings (1940) presented the following heat budget:

$$H_I = H_B + H_k + H_s + H_c + H_E$$

where

H_I = incident radiation.

H_B = back radiation (reflected short wave and re-radiated long wave).

H_k = energy used in heating the air.

H_s = energy used in heating the water.

H_c = lateral heat flow to or from the area under consideration.

H_E = energy utilized in evaporation.

For a land area H_s would be the energy consumed in heating vegetation and soil.

Direct measurements of H_I and H_B are possible with radiometers. Measurement of H_s is relatively simple in the case of water. With sufficiently long periods H_s for land surfaces can be neglected as insignificant by comparison with other terms.

H_c, the lateral energy transport can be neglected in certain cases where long time periods are considered or where large land or water areas are involved.

Because of atmospheric turbulence the energy consumed in heating the air H_k is difficult to evaluate. However Bowen (1926) postulated that the ratio of H_k to H_E, now known as Bowen's Ratio could be expressed by the following formula :

$$H_k/H_E = R = 0.61 \left(\frac{T_s - T_a}{e_s - e_a} \right) \left(\frac{p}{1000} \right)$$

where T_s and e_s are the temperature and vapour pressure of the water surface respectively, T_a, e_a are the temperature and vapour pressure of the air respectively, and p is the atmospheric pressure. Temperatures are in degrees Fahrenheit and pressure in millibars.

The determination of Bowen's Ratio still presents considerable difficulty. However the biggest drawback to the general application of the energy budget approach to the calculation of evaporation for hydrologic purpose is the paucity of data on radiation.

Incoming short-wave radiation may be estimated by indirect means, from duration of sunshine or cloudiness and latitude, Fritz and MacDonald (1949) Mateer (1955), Brunt (1939), Black et al (1954), Sapsford (1957). However Gabites (1956) pointed out that the standard deviation of the differences between computed and measured values of incoming radiation was 60 cal/cm^2/day for a period from January to June for four stations in New Zealand using the relationship from data at Rothamsted, Penman (1948).

$$H_I = H_A \left(0.18 + 0.55 \frac{n}{N}\right)$$

where H_I = incoming radiation

H_A = Angot value of radiation at the earth's surface in the absence of an atmosphere.

$\frac{n}{N}$ = ratio of actual to possible hours of sunshine.

The percentage error could be quite large when one considers that the monthly averages for H_I in the measurements made by Gabites ranged from 80 to 650 cal/cm^2/day.

The reflected short wave radiation will vary with the nature of the surface and may be as high as 20 per cent.

The Brunt (1939) equation for net radiation ($H_I - H_B$) in the Cummings (1940) formulation is usually depended upon. The term for long wave radiation exchange between earth and sky is an empirical one and probably suffers from defects which preclude its universal application. In the discussion of a paper by Hounam (1956), Priestley states that the latter term, H_B has not been satisfactorily tested for Australia. The same is no doubt true for other regions.

3.3.4 Combined Energy Budget-Sink Strength Approach.

The difficulties previously mentioned in connection with the Sink Strength approach with regard to measurement of the temperature of the water at the evaporating surface were overcome by Penman (1948) by combining the two approaches to give the following formula:

$$E_o = \frac{H\Delta + E_a\gamma}{\Delta + \gamma}$$

where E_o = evaporation from an extended water surface.

H = net radiation in evaporation units ($H_I - H_B$, of Cumming's formula).

Δ = vapour pressure gradient in mm. of mercury per degree F. at the air temperature T_a.

E_a = $(e_a - e_d)\ f(v)$ in which $f(v)$ is a wind velocity function.

γ = psychrometric constant for wet and dry bulb hygrometer equation (in ^{o}F and mm. of mercury, $\gamma = 0.27$).

The evaporation will be in the same units as the heat budget units.

Penman on the basis of experiments at Rothamsted determined values of E_T (potential evapotranspiration) for closely cut grass as

$$E_T = fE_o$$

where f is a seasonal factor having a range of values from 0.6 in the winter to 0.8 in the summer.

While Penman's formula would be considered semi-empirical, it represented a major advance in the application of sound physical principles to the determination of potential evapotranspiration.

Many papers have been written in which evaporation computed by Penman's formula has been compared with that computed by other means, or with measured results. A few of these are Makkink (1957a, 1957b) van Wijk and De Vries (1954) Stanhill (1958) Businger (1956), Closs (1956), Smith (1959) Law (1957), Makkink and Heemst (1956).

Penman (1952) improved upon the original formula with an expression for potential evapotranspiration independent of the seasonal multiplying factor f.

$$E_T = \frac{\Delta H_T + \gamma E_a}{\Delta + \gamma /SD}$$

where S = diffusion resistance of plant stomata when open.

D = a factor relating to the influence of closing stomata at night.

Makkink (1957b) in comparing the computed evapotranspiration with that measured by lysimeters in the Netherlands found a considerable discrepancy between values of H_I as determined from the relationship:

$$H_I = H_A(0.18 + 0.55 \frac{n}{N})$$ and the measured values of H_I.

Makkink (1957b) also found that:

E_T/E_o < 0.5 in summer.

> 0.8 in the winter.

From his work Makkink presents a somewhat simpler expression for evapotranspiration.

$$E_T = \frac{0.61 R_m \Delta}{\Delta + \gamma} - 0.12 \text{ mm per day}$$

The symbols are the same as in Penman's formula except that R_m is the measured incident radiation at the evaporating surface.

The evidence to date on Penman's formula appears to support its general use where the necessary basic data (temperature, vapour pressure, wind and cloudiness) are available, for the determination of evaporation from a free water surface. For the estimation of potential evapotranspiration from a vegetative surface of closely cut grass (water supply not limiting) the formula will give estimates on an annual basis within ten per cent and within twenty per cent for shorter periods of time such as a month.

3.3.5 Evaporation Based upon Temperature.

Probably the best known of the equations for determining potential evapotranspiration, E_T is that credited to Thornthwaite (1948) who uses the symbol P.E. for potential evapotranspiration from a low grass cover. Utilizing a heat index based on temperature, monthly values of E_T are calculated or determined from a nomograph. These values are in turn adjusted for the duration of daylight hours at the particular location, i.e. latitude.

The formula presented by Blaney and Criddle (1950) has been widely used in the Western United States for determining consumptive use (which approximates E_T by definition) in irrigated valleys of the region. The Blaney-Criddle formula is based on mean monthly temperature and monthly hours of daylight expressed as a percentage of the annual daylight hours.

$$u = \frac{ktp}{100}$$

u = consumptive use in inches for a month.

k = crop use coefficient.

t = mean monthly temperature ^{o}F.

p = percentage of daylight hours for the month.

Serra (1954a) presents a formula based upon temperature and relative humidity, which in effect is an empirical combination of the sink strength formula of Dalton (1802) and energy formula if we consider temperature to be a rough measure of energy.

There are a number of other formulas for evaporation and transpiration which depend upon temperature as the main parameter. Lowry and Johnson (1942) presented a formula primarily intended for the irrigated areas of the Western United States. Turc (1954) presented a formula based upon precipitation and temperature for actual rather than potential evaporation.

Van Wijk and De Vries (1954) critically reviewed both the Thormthwaite and Penman approach and concluded that formulas based upon temperature were quite unreliable because temperature lags considerably behind solar radiation. The net energy supply must be considered the most important single item affecting potential evaporation. The authors pointed out that for similar temperatures in spring and autumn, the solar radiation is much greater during the spring period. In Australia for example, February temperatures are normally slightly higher than December temperatures, yet higher levels of evaporation consistently occur in the month of December, when solar radiation is greater.

In spite of the serious drawbacks of the Thornthwaite formula, reasonable agreement between observed E_T and calculated values on an annual basis have been determined by some workers.

3.3.6 Evaporimeters and Lysimeters.

Evaporation pans have been used to indicate evaporation from large free water surfaces by the application of an appropriate multiplying factor. The value of the coefficient depends upon the type of evaporation pan and the exposure of the pan. Every type of pan suffers from the rather obvious defect inherent in a small sampling device, that of boundary conditions. Pronounced advective heat and moisture transfer may take place over the water surface. Sensible heat transfer through the walls of evaporation tanks may be quite considerable particularly in the above ground type such as the U.S. Weather Bureau Class A Pan. However if corrections are made for heat transfer through the tank walls, the tank can be used as an integrator of net radiation. This when used with Penman's formula is probably better than the calculated net energy based on duration of sunshine.

Lysimeters are tanks filled with soil, with or without vegetation. When troublesome border effects are eliminated by locating the device within a large area of similar vegetation, they can give a good measure of actual evaporation. Great care is necessary in the installation of lysimeters and measurement of the water balance for the device.

Many investigators have reported on evaporation measurements from lysimeters, including Harrold (1958) Makkink (1957a), Penman (1948), Suomi and Tanner (1958), Deij (1954), Law (1957), Stanhill (1958), Gilbert and Van Bavel (1954).

3.3.7 Actual Evaporation.

Of more general interest in the hydrologic aspects of yield is the actual rather than potential evapotranspiration. This is particularly true in the semi-arid and sub-humid regions of the world.

It is generally recognized that E_V must be limited in some way and at some time by availability of water to the plant roots. However there has been considerable controversy regarding the amount and degree of departure of E_V from E_T, as the soil dries out.

Veihmeyer and Hendrickson (1955) have steadfastly maintained that evaporation takes place at the potential rate over the full range of soil moisture from Field Capacity to Wilting Point. However, their experiments were carried out under special conditions with trees confined in a container, which may have involved rather artificial root concentrations. Further since the Wilting Point defined as the 15 Atmosphere soil moisture tension, is an arbitrary one in any case, it may be that the prune trees and pine trees used in the experiment were capable of extracting water over a much wider range of moisture tension without undue stress. Although it is difficult to criticize the work of Veimeyer and Hendrickson (1955) it is hardly possible to extrapolate their conclusions to the full range of vegetation.

Slatyer (1956) in a laboratory study found that evapotranspiration varied approximately linearly with available moisture for peanuts and cotton and curvilinearly for sorghum.

$$E_V = b(S_{ss} - S_{sso})\, E_W^{0.75}$$

where b = a constant.

$(S_{ss} - S_{sso})$ = water available in the soil for plant use.

E_W = pan evaporation (Australian pan)

West and Perkman (1953) also in Australia present equations for accumulated evapotranspiration as a function of accumulated pan evaporation.

$$W_s = 5.23 + 7.12e^{-0.219W_o}$$

W_s = moisture supply remaining in the soil at any time, expressed as a percentage (dry weight basis).

W_o = accumulated pan evaporation at a particular time.

e = exponential base of Naperian logarithms.

It should be noted that the pan is of a special type, the evaporation from which is approximately 1.4 times that from the standard Australian pan. Similar expressions are presented with different co-efficients for clean cultivated and for bare soil when the vegetation is controlled by oil. The difficulty with this expression is that apparently modifications are necessary to account for differences in the apparent specific gravity of the soil, in order to convert to volumetric units of water.

Smith (1959) working in the West Indies found that the following form of relationship was valid for grass covered plots:

$$E_V = k(S_{ss} - S_{sso})\ E_T$$

k = a constant

$(S_{ss} - S_{sso})$ = Available Moisture in the soil.

E_T = Evapotranspiration computed by Penman's formula.

This compares favourably in form with that presented by Slatyer (1956).

Marlatt et al (1961) found a similar relationship in the United States for snap beans:

$$E_V = E_T\left(\frac{S_{ss} - S_{sso}}{FC - S_{sso}}\right)$$

where $(FC - S_{sso})$ = Moisture available between Field Capacity and Wilting Point.

In this case the E_T was actually that computed by the Thornthwaite procedure and called P.E.

Hartmann (1960) presents data for a Bermuda grass sod in the Blacklands of Texas;

$$SM_t = SM_o K^t$$

where SM_t is the soil moisture at time t

SM_o is the initial soil moisture

K is a depletion constant which varies with the season, soil moisture and E_T. The relationship is given in the reference.

The formula may be recast as follows, using a time unit of one day and K, for the summer period, where $S_{ss} = SM_o$:

$$E_v = S_{ss} (0.04 - 0.003S_{ss} + 0.012 E_T)$$

The work of Hartmann is supported by Makkink and Heemst (1956) who working with lysimeters in the Netherlands found that for low levels of E_T (1.7 mm. per day) that $E_v \approx E_T$ for soil moisture tensions of up to 7 metres. As E_T increased to 2.4 mm. per day the relationship $E_v \approx E_T$ held only for very low tensions of 2 metres or less. The apparent explanation of this is that the rate of unsaturated flow of water within the soil is insufficient to satisfy potential evapotranspiration requirements. Higher tensions represent dryer conditions and lower values of hydraulic conductivity.

The evidence strongly supports the contention that $E_v < E_T$ at high moisture tension and at high values of E_T. Additional experimental work is required to evaluate the exact relationship.

3.4 WATERSHED STORAGE.

Four major types of storage are of interest in affecting the water balance. These have previously been designated as follows:

S = storage in channels and reservoirs.

S_s = storage on surfaces of leaves, buildings, pavements, etc.

S_{ss} = storage as soil moisture in plant root zones.

S_g = ground water storage.

The basic influence of any type of storage is retention for evaporation <u>or</u> detention and subsequent release as streamflow.

3.4.1 Channel and Reservoir Storage.

The magnitude of channel storage varies markedly during a period of storm runoff. This short term variation is of secondary importance in yield studies however so will not be considered further.

Reservoirs if they are of substantial area in relation to the size of catchment must be considered in any water balance computation. The criterion to be used to determine whether reservoir storage changes are significant would be a comparison of volumetric changes in storage with the order of magnitude of error in other measurements involved in the water balance. In yield studies based on a water balance approach, changes in reservoir storage equivalent to less than 0.01 inch on the watershed over the study period would normally be disregarded.

3.4.2 Storage on surfaces.

Adhesion of precipitation to vegetative surfaces and surfaces of other materials accounts for an initial abstraction from precipitation. This water is normally subsequently evaporated. Evaporation of water intercepted by vegetation effectively dissipates energy which would otherwise result in evaporation of water from soil moisture storage, so that the effective loss is frequently much less than the apparent loss due to interception.

Intercepted snow fall may form a substantial loss of potential streamflow, particularly where the vegetation is of a coniferous type. Wilm (1944) reported that there was a difference of two inches of water equivalent in the snow under a Lodgepole pine forest and that in a nearby cleared area. The higher storage at the beginning of the melt period was in the snow pack on the cleared areas.

The method of determining the differences leads to some doubt as to the probable net effect over large areas. Wilm made comparisons in isolated cleared plots which could have accumulated on excess of snow at the expense of the surrounding forested area. Nevertheless it is well recognised that coniferous stands of forests intercept considerable quantities of snow and that sublimation is likely to dispose of a large portion of this during the winter months. At the same time it seems quite reasonable to expect losses through interception of snow fall on deciduous trees or low level vegetation to be much lower.

Interception of rainfall has been investigated by a number of people including Horton (1919) and Trimble and Weitzman (1954). The latter found that in a mixed stand of hardwoods in the Appalachian Mountains of the Eastern United States the interception was about 20 per cent in winter and summer. Stemflow was not measured, so that no values of net interception could be determined. Linsley, Kohler and Paulhus (1958) state that the net storage capacity of a well developed forest canopy will be from 0.03 to 0.06 inches. Horton's (1919) empirically derived values for interception on cultivated farm crops range from 0.03 inches for corn to 0.33 inches for cotton. The value for cotton appears to be excessive.

Experiments reported on interception can be notoriously misleading, since stemflow, throughfall, and drip may be entirely neglected in the placement of rain gauges. Burgy and Pomeroy (1958) reporting on a laboratory study of interception losses found that interception amounted to 0.041 inches, on three types of vegetation (1) sunflowers, (2) pure stand of soft chess (3) mixed stand of grasses and legumes. They also found that STD (stemflow, throughfall, drip) increased linearly on a 1:1 basis with rainfall after a precipitation of 0.25 inches at which time interception storage was satisfied.

The net effect of interception storage on a seasonal basis or annual basis is likely to be insignificant on natural watersheds if the assumption that the evaporation of intercepted water is compensated by a reduction of soil moisture depletion.

3.4.3 Soil Moisture storage.

Moisture is retained in the soil root zone by cohesive and adhesive forces. The upper limit of storage under conditions of good drainage is termed the Field Capacity. Under such conditions water in excess is removed under the action of gravity forces.

Volumetrically the upper limit of storage (Field Capacity) could range from 3 inches per foot depth of coarse textured soil to 7 inches for the same depth of a very fine textured soil. The range of available storage is not likely to be of the same order. Soil water is relatively available to plants to tensions of approximately 15 atmospheres, regardless of texture. At this tension the fine textured soil will probably have in storage something in the order of $4\frac{1}{2}$ inches while the coarse textured soil about $1\frac{1}{2}$ inches per foot of depth. The range of available water then respectively is $2\frac{1}{2}$ inches and $1\frac{1}{2}$ inches for fine and coarse textured soils.

The lower limit for storage of course can be much lower than that represented by the 15 atmosphere tension. Although the evaporation rate E_v is much less than E_T during extended periods of drought, nevertheless soil drying occurs well below the so called Wilting Point.

Texture has been mentioned as the major factor affecting soil moisture storage. Other properties such as soil structure and organic matter content are also highly important.

The fluctuations in storage of soil moisture are directly related to precipitation and evaporation which have been previously discussed.

3.4.4 Ground water storage.

The water which occupies the larger pore spaces and openings in soil and rock and which moves under gravity forces is known as ground water. The magnitude of this form of storage depends upon the geological character of the basin. Direct determination is difficult except by extensive sampling from borings.

On gauged rivers the amount of storage may be determined by integration of the ground water recession portion of the hydrograph:

$$S_{gt} = -\frac{q_{gt}}{\log_e K_r}$$

Where S_{gt} = the ground water storage at time t.

q_{gt} = discharge from ground water at time t.

K_r = recession constant peculiar to the stream and unit of time selected.

Also $q_{gt} = q_o K_r^{\,t}$

where q_o is the initial flow at the beginning of time period.

The value of K_r is not consistent varying slightly with the discharge and varying seasonally. The seasonal variation is probably due to discharge of ground water directly through vegetation by evapotranspiration where the water table is shallow such as on the banks of streams. The base flow however can be used as an indicator of ground water storage.

Observations of the water levels in wells within a catchment serve as an index of ground water storage. In the application of the Water Balance approach selection of time periods, at the beginning and end of which ground water storage is equal, will cause this term to cancel from the equation.

3.5 GROUND-WATER FLOW

Since the sequence and distribution of low flows is extremely critical to any yield investigations it is worthwhile to examine the principles associated with ground water flow towards a stream. The mathematical and descriptive models presented are necessarily very much simplified over actual field cases.

3.5.1 Darcy's Law.

The basis of both saturated and unsaturated flow of water in porous materials is Darcy's Law (1856).

$$V = -K\frac{d\phi}{dl}$$

where V = volume rate of flow of water through unit cross-section of the material.

K = hydraulic conductivity.

$\frac{d\phi}{dl}$ = energy gradient of the soil water.

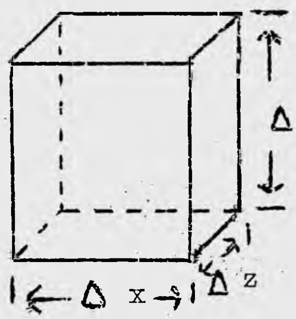
Since at this time we are interested only in saturated flow the more usual form of Darcy's Law,

$$V = -K\frac{dh}{dl} \text{ will be used.}$$

where $\frac{dh}{dl}$ = potential gradient.

3.5.2 Continuity Principle.

Consideration of continuity of flow within an element of saturated medium as illustrated leads to:



$$K_x \frac{\partial^2 \phi}{\partial x^2} + K_y \frac{\partial^2 \phi}{\partial y^2} + K_z \frac{\partial^2 \phi}{z^2} = 0$$

the Laplace equation.

where K_x, K_y, K_z are respectively the hydraulic conductivities in the three coordinate directions

and $\frac{\partial \phi}{\partial x}$, $\frac{\partial \phi}{\partial y}$, $\frac{\partial \phi}{\partial z}$, are the hydraulic gradients in the three coordinate directions.

For an element of medium at the water table (upper surface of the saturated zone in an unconfined aquifer), the expression may be presented as

$$\frac{dc}{dt} = K_x \frac{\partial^2 \phi}{\partial x^2} + K_y \frac{\partial^2 \phi}{\partial y^2} + K_z \frac{\partial^2 \phi}{\partial z^2}$$

where $\frac{dc}{dt}$ = rate of change of moisture within the element with time as the medium is drained.

It is more usual to simplify the expression to one of two dimesnsions so that

$$\frac{dc}{dt} = K\left(\frac{\partial^2 \phi}{\partial x^2} + \frac{\partial^2 \phi}{\partial y^2}\right)$$

with the additional assumption of isotropicity of hydraulic conductivity.

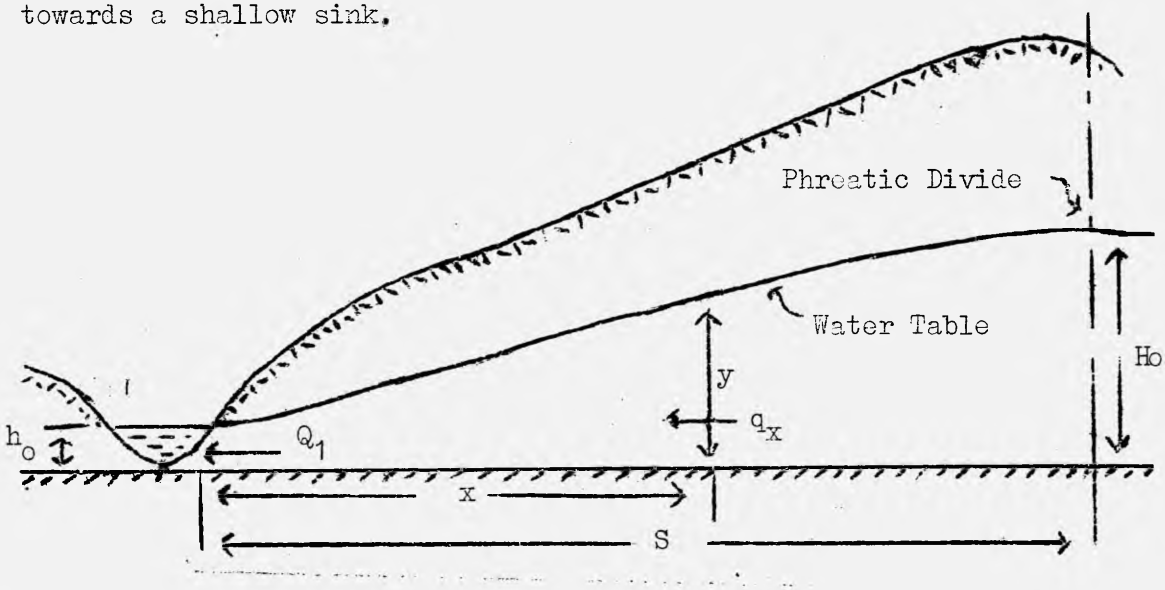
3.5.3 Steady-State Ground Water Flow.

The Dupuit-Forchheimer theory is frequently used to develop an expression for the geometry of the water table when ground water flows towards a trench or ditch. This theory has been extensively used in connection with the development of formulae for use in agricultural drainage. There are two basic assumptions in the theory:-

1. All streamlines in a system of gravity flow towards a shallow sink are horizontal.

2. The velocity along these streamlines is proportional to the slope of the free water surface, but independant of the depth.

These assumptions have been criticized by Muskat (1946), and critically examined by van Schilfgarde, Kirkham and Frevert (1956). Although the two assumptions are somewhat in contradiction if followed through rigidly, they do provide a rather simple expression for flow towards a shallow sink.



In the illustration we have a homogenous medium bounded at the elevation of the stream bed by a horizontal impermeable layer. It is further assumed that a uniform and steady rate of rainfall is removed equally well at all distances from the stream. Then the rate of flow, Q_1, into unit length of stream from the right, can be related to the flow q_x past any vertical section at distance x from the sink as follows:

$$q_x = \left(\frac{S - x}{S}\right) Q_1$$

Also

$$q_x = - y v_x$$

$$q_x = yK \frac{dy}{dx}$$

Equating the two expressions for q_x

$$\left(\frac{S - x}{S}\right) Q_1 = yK \frac{dy}{dx}$$

Integrating from $x = o$, $y = h_o$ to $x = S$ and $y = H_o$.

$$S = \frac{K(H_o^2 - h_o^2)}{Q_1}$$

or $Q_1 = \frac{K}{S}(H_o^2 - h_o^2)$ which is known as the ellipse equation. A number of investigators including Hooghoudt (1937), Aronovici and Donnan (1946) developed this equation independently.
Muskat (1946) criticizes the ellipse form because a surface of seepage at the sink has been ignored. He points out that there must be a seepage surface above the water level in the stream, otherwise an infinite velocity must occur at the point where the water table intersects the stream bank.

Schilfgarde et al (1956) state further that capillary flow above the water table has been ignored. The latter may be more important than the neglect of a seepage surface.

However in spite of these shortcomings, the ellipse equation is probably reasonably satisfactory where flow is steady and where the lateral dimension S is large compared to the depth. Horizontal flow certainly predominates.

The assumption of horizontal flow can not be followed where the impermeable layer is at some distance below the stream-bed, since the effect of flow convergence or radial flow in the vicinity of the stream must be considered. Hooghoudt (1937) considered this case and combined the radial flow and horizontal flow approaches.

The biggest restriction to the use of the ellipse equation as presented however is that steady state rainfall seldom occurs, and we are more interested in the time variation of discharge of ground water - a case of a falling water table or unsteady flow.

3.5.4. Unsteady State Ground Water Flow.

The analogy with heat flow is usually adopted to express the relationship for a falling water table.

Referring to the previous Figure we may state:

$$\frac{\partial y}{\partial t} = \frac{Ky}{\varepsilon} \left(\frac{\partial^2 y}{\partial x^2} \right)$$

where all symbols are as previously stated and ε = specific yield of the aquifer.

This equation in slightly different form was used by Glover as reported by Dumm (1954) to develop a formula for the spacing of tile drains on irrigated land. Although Glover's analysis included the case of an impermeable barrier at some distance below the drain, when radial flow occurs, the fundamental Dupuit-Forcheimer assumptions provide the basis for derivation. It has been shown by Hooghoudt (1937) that when the depth of the impermeable barrier is greater than

S/10, the effect of flow convergence becomes marked. This may be important in tile drainage where S is relatively small, but need not be so in drainage of ground water by river channels. Glover's solution is also based on the assumption of an average elevation for the water table at the mid-point between drains during the time period selected. From this standpoint considerable error could be involved if the height of the water table at the phreatic divide fluctuated widely over the period.

3.5.5 Storage-Discharge Relationships.

It is customary to assume a linear relationship between storage in ground water and the discharge from ground water into the stream:

$$S_g = K_{s\cancel{q}}\ q$$

More familiar forms are those which represent discharge as a function of the discharge at some time period earlier:

$$q_t = q_o\, K^t{}_r$$

where q_o is the discharge at some earlier time period, usually at the beginning of the ground water recession period.

q_t is the discharge at t, time units later, and

K_r is a recession constant peculiar to the stream.

Integrating this expression we obtain:

$$S_{gt} = \frac{-\ q_{gt}}{\log_e K_r}$$

where S_{gt} is the storage remaining in the basin at time t.

It will be noted that:

$$-\frac{1}{\log_e K_r} = K_s \text{ of the first equation, } S_g = K_s q$$

This K_s is sometimes called the storage delay time, since it has the dimensions of time.

While linearity may be a convenient assumption and reasonably valid over short segments of the recession curve, there is abundant evidence that the value of K_s is not constant over the entire recession range. This is illustrated by Linsley, Kohler and Paulhus (1958) p.155.

This is further borne out by reference to the previous discussion of Steady State Flow of Ground-water (3.5.3) and the Ellipse equation:

$$Q_1 = \frac{K}{S}(H_{01}^2 - h_{a1}^2)$$

The rate of flow at a later time is:

$$Q_2 = \frac{K}{S}(H^2_{O2} - h^2_{O2})$$

and if h_O is small compared to H_O then for deep elements of ground water storage the relationship,

$$Q = K^1(S_g)^2 \quad \text{would seem appropriate.}$$

For areas with considerable swamp in the vicinity of the stream and tributary a linear relationship would probably be more appropriate.

If the expression $Q = K^1(S_g)^2$ is differentiated:

$$dQ = 2K^1 S_g dS_g$$

and if $dQ = Q_1 - Q_2$

and $dS_g = Q_2$

where Q = (discharge per unit of time) x (one time unit)

$$Q_1 - Q_2 = 2K^1 S_g Q_2$$

But $S_g = \sqrt{Q_1/K^1}$

$$Q_1 - Q_2 = 2\sqrt{K^1 Q_1}\, Q_2$$

$$Q_2 = \frac{Q_1}{2\sqrt{K^1 Q_1} + 1}$$

or approximately

$$Q_2 = K_g \sqrt{Q_1}$$

where with appropriate dimensions for Q, $K_g \approx \dfrac{1}{2\sqrt{K^1}}$.

There are no known studies to confirm the analysis presented above.

3.6 SUMMARY OF THEORETICAL BASIS FOR RUNOFF GENERATION AND STREAMFLOW DISTRIBUTION

The water balance equation forms the basis for the foregoing presentation. In its simplest form the equation,

$$Q = P - E - \Delta S$$

provides a useful approach to yield studies.

However the inherent difficulties in evaluating the right-hand terms over short or long periods of time, and on a catchment wide basis are major obstacles to the application of this approach. This will be more apparent in Section 4. REVIEW OF EXPERIMENTS where the literature on the subject from the applied standpoint is examined.

4. REVIEW OF EXPERIMENTS AND INVESTIGATIONS OF CATCHMENT YIELD.

4.1 YIELD FROM SNOW

The pioneering work on the influence of snow on catchment yield was instigated by Church (1912). Snow in the mountain ranges of the Western United States provides the major source of water for irrigation purposes. It was natural then that forecasts of water supply for irrigation purposes should receive early attention.

Snow begins to accumulate in the mountain catchments during October and for the most part remains on the ground until late in April. The snow melts and is released as stream flow and surface infiltration through May, June and into July. A very high percentage of the annual yield occurs during these three months.

4.1.1 Amount of Snowfall.

Early attempts to predict streamflow directly from snow fall measurements were not entirely satisfactory. Suitable observation sites, particularly at the higher elevations were difficult to obtain. Accessibility of representative sites during the winter months was likewise a deterrent. There were few local inhabitants in the areas of highest snowfall who could be relied upon to make observations.

The U.S. Weather Bureau however does use precipitation data for making seasonal and annual yield forecasts. Bernard (1949) and Kohler and Linsley (1949) describe the procedure adopted. After adjustment of station precipitation and catchment runoff data by double-mass curve techniques, station weights are established by a least squares regression analysis of Q versus P_1, P_2, P_3, P_n, for the n stations. Similar methods are used to establish calendar month weights for October through June, and for establishing a relative index of effectiveness of the precipitation of the previous year.

The method can only be applied to catchments where concurrent discharge and precipitation data are available. Seasonal yield from snow-melt is based on using precipitation data as indices rather than as direct physical measures of water available for streamflow. This weakness prohibits the application of a relationship for a given catchment to an ungauged catchment, since the station weights are a function of the location, orientation and siting of the individual gauges.

4.1.2 Snow Courses.

A snow course is a series of sampling sites at short intervals of distance established so that the snow pack is representative of that in the area, and consistent from year to year as an index of snow water equivalent for a considerable area of the catchment. Several courses established over a catchment are sampled at monthly intervals for snow water equivalent, the final observations being about April 1 in the Northern Hemisphere. These measurements with subsequent streamflow measurements are used in a forecasting equation. The snow course observations are frequently weighted according to elevation of the course on the catchment.

Clyde (1932) suggested for the Logan River in Utah that:

$$Q = \text{Snow (\% of Normal)} \times Q(\text{normal})$$

He realized the necessity for very careful selection of snow courses and that a unique relationship probably existed for each catchment. Clyde (1939) published data for the Logan River, indicating the maximum variation of the actual stream flow from that forecast was 10 per cent, with an average of 5 per cent for a 14 year period based on the simple relationship above.

Hall (1944) discriminated between snow water equivalent as determined by survey on April 1, for different elevations on the Mokelumne Catchment of California. He developed a graphical relationship based on:

1. Snow water content above 6,000 feet.

2. Snow water content 4,000 ft. to 6,000 feet.

3. Normal April-June Precipitation.

Using the sum of (2) and (3) as the independent variable, the per cent of total catchment snow water content realized as streamflow from snowmelt was indicated graphically as the dependent variable. The author shows two such relationships, one for 1935-39 and one for 1940-42, without explaining the cause of the change.

4.1.3 Antecedent Conditions.

A number of investigators have recognized that improvements in the correlation of stream flow with snow survey data could be obtained if some weight was accorded to the ground water conditions of the preceding year. The usual method of doing this is to include a term in a regression equation for the forecast seasonal flow, which gives the stream flow prior to the first autumn snow fall. Clyde and Work (1943) and Church and Boardman (1943) mention the importance of this factor.

Nelson, Wilm and Work (1953) use the term 'soil priming' to indicate the influence of Autumn precipitation on ground water levels and soil moisture on the forecast stream flow for the Columbia River at the Dalles, Oregon.

$$Q = 14.239 + 2.627X_1 + 3.850X_2$$

where Q is in millions of acre-feet.

X_1 = weighted average water equivalent of snow.

X_2 = August-October precipitation departure from normal for the preceding year.

This equation gave a value of R^2 (Coefficient of Determination) of 0.788 and a standard error of estimate of 7.32×10^6 Acre-Feet or 12 per cent of the mean for 1937-1951.

Light (1960), for the mid-western non-mountainous area of the United States drained by the Mississippi River, attempted to use a graphical multiple correlation developed from storm rainfall studies as a means of forecasting runoff from snow melt in the spring of 1959. The independent variables used in the correlation were Antecedent Precipitation Index (A.P.I.) of Linsley, Kohler and Paulhus (1958), season of the year, duration of rain and amount of rain. The dependent variable was runoff. The A.P.I. selected for the forecast was that occurring on November 26, 1958, immediately prior to freeze-up. The actual yield from snow melt in Light's (1960) study was less than that forecast. The reasons attributed by Light were that intermittent melting had contributed an unusual proportion of melt water to ground water recharge, and that the presence of dry air masses over the area had depleted snow water amounts by sublimation.

A fact not mentioned by Light is that the assumed A.P.I. of the preceding autumn may not be indicative of that immediately preceding melting. Considerable moisture migration within the soil profile, particularly during periods of the winter when the ground surface is not covered by snow, may result in an appreciable reduction in the A.P.Index through the winter months.

4.1.4 Concurrent Meteorological Conditions.

The rate of melting as influenced by the heat balance, and precipitation during the melt period are the major factors which result in errors in seasonal yield forecasts. Croft (1946), based on studies in the Wasatch Mountains of Utah, asserts that rapid snow melt generates more stream flow than slow snow melt. Frame (1944) presented data for the 1943 season in the Columbia, Kootenay and Okanagan basins and the coastal belt adjacent to Vancouver, Canada, indicating that the abnormally cool weather of May and June resulted in yields much below those forecast on the basis of snow course surveys. It would be assumed that with lower melt rates, a greater share of water infiltrates to recharge soil moisture and ground water. The accretion to ground water would eventually be yielded in the form of a higher level of baseflow following the melt season.

For the same region Frame (1944) points out that the 1942 yield was about 70 per cent above the forecast value because of abnormal precipitation and high temperatures during the April-July period. Frame (1945) presented data for the same area which showed streamflow to be only 70 per cent of the forecast value, when precipitation was only 69 per cent of normal for April-May-June.

Paget, (1946) incorporated antecedent and concurrent precipitation into a forecasting curve for the Kaweah River.

$$Q = K \left[(P_A - 8) + S + 1.25 P_S \right] + C$$

where K and C are constants

P_A is precipitation of previous autumn

$(P \nless 8)$

S = Snow water equivalent on April 1st

P_S = April-May precipitation

The relationship was developed on the basis of 15 years of record, giving a maximum error of 10 per cent.

Revisions in the April 1st forecast may be made periodically through the melt period by considering departures from precipitation normals subsequent to the date of the new forecast.

For example Nelson, Wilm and Work (1953) previously cited, have the following equation for the Columbia

$$Q = 7.402 + 2.879X_1 + 1.943X_2 + 11.083X_4$$

where Q = yield after May 15, millions of acre-feet.

X_1 and X_2 are as previously stated

X_4 = Precipitation departure from normal for April 1 - May 15

Ford (1959) illustrates the use of the multiple regression approach to forecasting seasonal runoff. For the Colorado River at Cameo, Colorado

$$Y = 0.120X_2 + 0.129X_3 + 0.178X_4 + 0.171X_5 - 1.970$$

where Y = discharge in millions of acre-feet, April-July

X_2 = July-Sept. precipitation of preceding year.

X_3 = October-January precipitation

X_4 = Snow water equivalent from average of April 1 and May 1 forecasts

X_5 = May-July precipitation

In this case R^2 (Coefficient of Determination) = 0.884, and standard error of estimate = 0.169 or 7.3 per cent of the mean value of discharge.

4.1.5 Summary of Current Status.

While great strides have been made in forecasting yield from river basins from snow melt, the procedures are purely empirical and require a prior record for a number of years of a number of relevant factors so that a reliable correlation can be developed. Snow surveys provide an index value of the water available for streamflow, while precipitation measurements are merely samples of catchment precipitation. Where snow surveys and precipitation measurements can be highly representative of conditions on the catchment a reasonable estimate of the water balance can perhaps be made on a monthly or seasonal basis for a particular area. However there is no acceptable method available for the prediction of yield from snow melt on ungauged catchments. The complexity and interrelationship of the many variables involved has so far prevented a completely integrated solution and understanding of the basic physical phenomena involved in the transition from water in the solid state on the ground surface to streamflow.

Hourly changes in the snow itself alter the water content, melt rate and thermal conductivity. Marked changes in soil infiltration capacity take place in a few hours as the depth of snow is depleted and more energy is available to thaw the ground surface, thus increasing the receptivity to water.

Until considerable progress can be made in expressing the time variation of the infiltration capacity of an initially frozen soil as a function of snow cover condition with time and the energy relationships involved in snow melt it will be necessary to rely upon the development of suitable correlations for each catchment based on measurement of significant variables.

4.2 YIELD FROM RAINFALL - ANNUAL AND MONTHLY.

Most of this section deals with literature on annual yield but by implication, relationships of a similar form could be developed for monthly yield. It would be expected that there would exist a unique relationship for each of the twelve months. In a later section 4.4 DRY WEATHER FLOW, the month as a unit of time is frequently adopted, although shorter and longer time periods are also used.

4.2.1 Linear Relationships.

The simplestbasis for annual yield determinations is based upon a two variable regression of annual yield on annual rainfall.

One form which has application in certain cases is

$$Q = P - b$$

where Q = annual yield

P = annual precipitation

b = a constant.

Implicit in this form is the fact that above a certain minimum value of precipitation equal to b units, all the surplus is streamflow. The minimum value of P = b, would be that required to satisfy evaporation requirements. In cases where rainfall at all times is sufficient to satisfy evaporation requirements, then $b = E_T$, and it would be expectedthat the annual evaporation would fluctuate relatively little.

A case in point is illustrated by Sutcliffe and Rangeley (1960) for the Tongariro River of New Zealand, where,

$$Q = P - 20 \text{ inches.}$$

In the nine years of record the annual precipitation was never less than 64 inches, fairly evenly distributed throughout the year, so that $E_V = E_T$ for most of the time.

It would seem, though, that this type of relationship can be applied only where rainfall is abundant and well distributed with respect to evaporative demands. This would normally mean that $P > 2E_T$.

A more versatile form of relationship is

$$Q = a(P - b)$$

Again there is no runoff when $P < b$, and yield is a constant proportion of the surplus of P over b. That is both evaporation and stream yield increase with rainfall.

Again Sutcliffe and Rangeley (1960) use this form for the Tana River of East Africa

$$Q = 0.406\ (P - 17) \text{ inches.}$$

The computed minimum precipitation for the correlation period was 28.43 inches. Actual evaporation is no doubt much less than the potential in many years. It should be noted that the standard error of estimate in this case was 3.39 inches or 29.2 per cent of the mean runoff, while in the case of the Tongariro River the standard error of estimate was 13.92 inches, but only 16.5 per cent of the mean value. It would be expected that less reliability can be placed upon a simple two variable linear relationship where lower rainfalls are experienced, because of the dependence of evaporation losses upon rainfall and because of the natural high relative variability of annual rainfall normally associated with low rainfall regions.

The Thames River in England has been the subject of several studies of this nature. Gold (1951) quotes the following but does not give the source.

$$Q = 0.67(P - 14.6) \text{ inches.}$$

$$Q = 0.67(P - 13.6) \text{ inches.}$$

$$Q = 0.57(P - 10.6) \text{ inches.}$$

The mean annual rainfall for the area is about 29 inches.

The error in using this simple relationship is stated by Gold to be in some cases 50 per cent of the mean annual runoff. Since rainfall intensities are low, it would appear that the variation of actual evapotranspiration with rainfall is responsible for the variation in stream yield.

Glasspoole (1960) gives the following for the Thames:

$$Q = 0.46(P - 7.8)$$

There is a considerable difference between the relationships quoted by Gold and that of Glasspoole, except near the mean value of precipitation, probably due to the method of establishing the slope of the regression and the judgement of the investigator.

4.2.2 Non-Linear Regressions.

Two forms which lend themselves readily to regression analyses are:

$$Q = aP^n$$

and $$Q = a(P - b)^2$$

The latter form is suggested by Coutagnie (1954) and the example given is for the Neosha River of Kansas, U.S.A.

$$Q = 0.017(P - 17.5)^2 \text{ inches.}$$

and for le Vaal of the Union of South Africa

$$Q = 0.02(P - 18)^2$$

It is remarkable how nearly identical the two regressions are. The coefficient of determination, $R^2 = 0.89$ in the latter case, is exceptionally good for a two variable regression. Again the relation is limited to $P > b$.

This type of relation would be best suited to areas where rains of high intensity result in abnormal precipitation in those years of higher than average precipitation. In other words high intensity storms have a higher proportion of runoff than lower intensity storms.

An upper limit for which such a relationship is valid would be necessary since by differentiating

$$Q = 0.02\,(P - 18)^2$$

$$\frac{dQ}{dP} = 0.04P - 0.72$$

The maximum value of $\frac{dQ}{dP} = 1$, that is one unit of yield per unit of rainfall, occurs when $P = 43$ inches. Beyond this

$$Q = P - 30.5 \qquad P > 43.0''$$

Below this $Q = 0.02(P - 18)^2$ $\qquad 43 > P > 18$

Harrold (1957), gives the following relation for several small catchments in Ohio, U.S.A. based on 18 years of measurements.

$$Q = b \log P + a.$$

The catchments ranged from 29.0 acres to 17,540 acres in size. The coefficient of determination, R^2 ranged from 0.62 to 0.72, which is not particularly good, so it would be assumed that the standard error of estimate is fairly high.

Webber (1958) found for the North Pine River of Queensland that

$$Y = 2.447X - 4.717$$

where Y = logarithm (base 10) of annual discharge in millions of gallons.

X = log. of Annual rainfall at Petrie in hundredths of inches.

The R^2 was 0.75 and standard error of estimate 0.3108 log units.

4.2.3 Yield from Seasonal Precipitation.

The weakness of the simple relationships between annual yield and annual precipitation can be partially overcome by some form of seasonal weighting of precipitation. Mention has been made of the procedure adopted by the U.S. Weather Bureau, discussed by Bernard (1949), with particular application to yield from snow.

Glasspoole (1960) for the Thames River gives the following multiple regression:

$$Q = 0.18P_{wl} + 0.51P_{sl} + 0.73P_{wo} + 0.13P_{so} - 13.0 \pm 0.92 \text{ inches.}$$

where Q = annual yield in inches.

P_{wl} and P_{sl} are precipitation for winter and summer of previous year respectively.

P_{wo}, P_{so} are precipitation for winter and summer of current year.

The percentages of the total variation ascribable to the seasons according to Glasspoole are

R^2_{wl} - 3.7%

R^2_{sl} - 24.4%

R^2_{wo} - 63.2%

R^2_{so} - 1.6%

Glasspoole states that 92.9% (sum of coefficients of partial determination) of the variability in streamflow is accounted for in the regression equation. However this is not true since there is no doubt some correlation between the serial rainfall events, even though they appear to be completely independent.

Nevertheless the use of seasonal weights has reduced the standard error of estimate from 25 per cent of the mean for a simple two variable relationship to 9 per cent of the mean for a five variable relationship.

Srebronovic (1957) determines annual yield for the region between the Sava and Drava rivers in Croatia on the basis of computed monthly discharges based on current and preceding monthly precipitation.

$$Q = a\left[\log.\ (H_1 + 10)\right]\left[(H + 10)^b\right]$$

where Q = Monthly runoff

H = Monthly rainfall

H_1 = Monthly rainfall of preceding month

a, b, are constants which vary with each month.

The standard error of estimates indicate that for several months during the year the error could exceed 50 per cent of the mean monthly discharge. The presence of snow during the winter results in a considerable error during the winter and spring months in particular.

4.2.4 Effect of Precipitation of Antecedent Year.

Some investigators have reported that there is relatively little effect of precipitation for an antecedent year on the yield for a given year, when yearly totals are considered. Webber (1958) reported that annual discharges on the North Pine River of Queensland were not serially correlated, which indicates at least that the correlation between rainfall of antecedent year and current year discharge is slight.

Gold (1951) in investigating the effects of rainfall by quarter-years found that rainfall in the fifth quarter preceding a given quarter had negligible effect on the runoff of the given quarter. However on an annual basis the selection of the water year is no doubt very critical in determining whether antecedent precipitation is influential.

For example Gold (1951) gives the following correlations between annual yield and annual rainfall on the Thames for the respective water years:

Jan. 1 - Dec. 31	R = 0.71
April 1 - March 31	R = 0.91
July 1 - June 30	R = 0.89
Oct. 1 - Sept. 30	R = 0.68

It would be expected that for water years beginning

50 per cent) could be explained by precipitation in the preceding) to year.

It is interesting to note that Sharpe et al. (1960)

It is interesting to note that Sharpe et al. (1960) found the best correlation between annual yield and annual precipitation for the May 1 - April 30 water year, (R = 0.946) and the poorest for the Aug. 1 to July 31 water year, (R = 0.817). This was for the Delaware River Basin of Kansas. It appears that a water year beginning when the ground is moist gives the best correlation. In the case of both the Thames and Delaware, consistency from year to year in this aspect of watershed condition is to be expected in the spring of the year.

Hoyt and Langbein (1944) concluded from a broad study of streamflow in the United States and Canada that annual events were largely independent.

Siren (1960) of Finland calculated correlation coefficients for the Kymijoki River of Finland showing a high correlation between autumn precipitation and subsequent water year runoff.

For the water year Jan. 1 - Dec. 31 the following correlations were found:

Precipitation Period.	R
Aug. 1 - Nov. 30 (preceding)	0.65
Jan. 1 - Dec. 31 (water year)	0.57
Aug. 1 (year preceding) - Dec. 31 (of water year)	).85

4.2.5 Temperature as a Factor in Yield.

In recognition of the inverse association of yield with water losses, and of the direct relation of water losses (Primarily evapotranspiration) with temperature and precipitation, a number of investigators have suggested empirical formulae for annual yield based on these parameters.

Vermuele (1894) was one of the earliest to propose such a formula :

$$Q = P - E$$

(Vermuele used F for Q, and R for P)

and E = annual retention in inches.

Vermuele apparently recognized E as being largely attributable to evaporation, although he preferred the more general inclusive term of retention to account for changes in soil moisture and ground water storage.

Vermuele expressed E as a function of temperature and rainfall :

$$E = (15.50 + 0.16P)(0.05T - 1.48)$$

where T = mean annual temperature in degrees Fahrenheit. The formula of Vermuele is remarkable for the era, since he did recognize a dependence of evaporation upon rainfall and temperature. He later (1899) modified the expression for E to :

$$E = (11 + 0.29P)M$$

where M varies from a value of 0.77 for an annual temperature of 40° F to 1.51 for a temperature of 61° F. Vermuele published values for monthly retention for the Sudbury, Croton and Passaic Rivers of New Jersey. The basic formula was based on data from this area of the United States.

Somewhat later Justin (1914) developed a formula for annual runoff in the North-eastern United States:

$$Q = \frac{0.934S^{0.155}P^2}{T}$$

(Justin also used F for Q and R for P).

T = mean annual air temperature °F

S = slope of drainage area, being equal to maximum elevation difference divided by the square root of the area.

The introduction of the slope term is no doubt an important concession to the contribution of storm flow on the rather rough and sometimes rocky areas of the North-eastern United States, where erosion of steeply sloping lands has materially reduced infiltration capacity of the soil.

Grunsky (1922) also developed an empirical formula based on both precipitation and temperature for California catchments:

$$Q = r^1 + 0.012P(55 - t^1)$$

where r^1 is the runoff in inches resulting from precipitation in the climatic year for low elevations of the State of California.

The author indicates values of r^1 graphically.

P = precipitation for the climatic year.

t^1 = mean temperature Dec. 1 to June 1, ^oF.

The climatic year of Grunsky's formula is from July to June or August to July rather than the calendar year.

In Grunsky's formula, the effect of increased evaporation in reducing runoff is taken account of in the temperature factor. However, the effect is also related to the availability of water. The effect of high temperatures on actual evaporation increases with greater amounts of precipitation.

The author fails to give examples of computed and observed discharge so that it is difficult to criticize the formula. However, it should be fairly obvious that no account of storage changes is provided, so that on watersheds with high storage of ground water, considerable error could result in annual calculations.

Wundty (1937) presents a graphical relationship between Q/P and precipitation and temperature.

Turc (1954) developed a formula for evaporation based on runoff records for 254 streams in different parts of the world, based on precipitation and temperature. Implicitly $Q = P - E$.

The weakness here again is that these formulas may be adequate to determine average annual yields but are certainly very inadequate to determine year to year yields, because storage changes are in no way accounted for.

4.2.6 Average Yield from Catchment Evaporation.

Using long term data on streamflow and precipitation, empirical relationships for evaporation as a function of precipitation and temperature have been developed. Basically this is a water budget application of the formula:

$$Q = P - E - (\Delta S + \Delta S_s + \Delta S_g + \Delta S_{ss})$$

With a sufficiently long period of record it is reasonable to assume that the summation of storage charges is small in comparison to the other terms so that:

$$E = P - Q$$

The work of Turc (1954) and Thornthwaite (1948) was directed towards establishing a relationship for (a) evaporation as a function of temperature and precipitation (Turc), and (b) potential evapotranspiration (Thornthwaite) as a function of temperature and latitude.

However, in both cases streamflow data were used in some degree as the basis for deriving the relationship.

In Turc's investigation 254 basins in Europe, Africa, America and Java were selected to obtain the formula for evaporation:

$$E = \frac{P}{\sqrt{0.9 + \frac{P^2}{L^2}}}$$

where $L = 300 + 25t + 0.05t^3$

E will be in millimetres per year when

P is in mm. per year and t is mean annual air temperature in $^{\circ}C$.

Thornthwaite's (1948) work has been previously mentioned under Section 3.3.5. In a subsequent publication Thornthwaite and Mather (1955) illustrate the application of the formula for potential evapotranspiration to the determination of the yield of a drainage basin. The procedure involves a monthly water budget based on certain boundary conditions.

(1) Actual evapotranspiration is proportional to amount of water stored in the soil.

(2) Surplus calculated for a given month distributed as streamflow with 50 percent during the current month and 50 per cent of the balance during succeeding months.

(3) Surplus only occurs after soil moisture has been recharged to Field Capacity.

Aside from the obvious weaknesses involved in the method of calculating evaporation in Turc's formula, the application to yield determination on a year to year basis is not feasible because storage changes are entirely neglected. As a means for determining the long term average yield it would seem to have some merit, however, and should be restricted to applications of a broad geographic survey type.

As for Thornthwaite's (1955) approach about the same comments are appropriate. The refinements in adjusting evapotranspiration and distributing streamflow exponentially are an improvement over Turc's method. However careful evaluation of the evaporation calculated to determine the extent of agreement with actual conditions for the particular geographic region is essential. Further there are great differences in streams with respect to the ground water depletion characteristics. Judicious selection of a water year when consistent storage conditions are likely to prevail will overcome the drawback to a certain extent.

4.2.7 Monthly or Daily Water Balance.

The first attempt by Penman (1950) to apply his (1948) method of calculating evapotranspiration to the determination of streamflow was in connection with the Stour catchment of England.

A monthly budget was carried out by a distribution of catchment evaporation according to proportions assumed appropriate to the Stour as follows;

(1) riparian vegetation (20%).
(2) deeprooted vegetation - trees etc. (30%).
(3) shallow rooted vegetation - pastures etc. (50%).

It was assumed that evaporation at the potential rate took place; at all times on the area of riparian vegetation; until moisture was depleted by 8 inches below Field Capacity on deep rooted crops and; until moisture was depleted by 5 inches below Field Capacity on shallow rooted crops. Penman called the 8 inches and 5 inches the 'root constants'. Evaporation rate was reduced after the depletion of the root constants. There was no attempt to make temporal distribution of surpluses, rather it was assumed that all surpluses drained off during the current month.

The agreement between calculated and observed values of streamflow seem to be very good, although this could be considered no more than an exploratory study. There is considerable dependence upon selecting the appropriate 'root constants' and the proportions of riparian, shallow and deep rooted crops. The latter could be done from land use surveys or aerial photographs. The particular root constants appear to be too high for many areas, although at the moderate levels of potential evapotranspiration prevalent in England they could be reasonable. However, in areas of high potential evapotranspiration the actual rates of evaporation are likely to be much below the potential. (See Section 3.3.7.).

Linsley and Crawford (1960) used a daily water hudget primarily to obtain distribution of daily flows, but from which monthly or annual yields could be determined. In their case actual evapotranspiration was assumed equal to the potential evapotranspiration rate. The value assumed was the normal evapotranspiration varied semi-monthly, computed by the method of Kohler (1957) from nearby pan data. Other adjustments were made for infiltration capacity limitations as a function of soil moisture storage, for storms of high intensity producing direct runoff, and for time distribution of surface and ground water flow.

The agreement between computed streamflow and actual yield over ten years was within six per cent of the actual, which is extremely good. Some criticism of the method for computing actual evapotranspiration could be raised except for the logical explanation advanced by the authors. In the region of the investigation, Los Trancos Creek, near Palo Alto, California, a distinct rainy season and dry season characterize the water year. During the wet season, evaporation of moisture from the upper soil horizons at the potential rate would appear logical, since these are the first to be recharged after the dry season. As precipitation exceeds evaporation, the deeper horizons are gradually recharged, while the upper horizon is maintained at or near capacity. Furthermore potential evaporation rates are low during the rainy season and evaporation at potential rates could be expected where moisture stresses are low.

During the dry season there is no direct runoff and soil moisture from the entire profile is depleted prior to the next wet season. The actual rate of depletion is unimportant, since there is no soil moisture or ground water recharge.

Although the ten year totals show good agreement with the actual values, some years are in error by a considerable amount.

YIELD OF LOS TRANCOS CREEK * (INCHES)

Year	1931-32	32-33	33-34	34-35	35-36	36-37	37-38	38-39	39-40	40-41
Estimated	7.06	1.59	1.07	5.41	2.95	9.78	13.66	1.35	15.11	17.52
Actual	5.95	1.55	1.17	3.92	4.26	8.14	14.32	1.03	13.44	16.03
Difference	+1.11	+0.04	-0.10	+1.49	-1.31	+1.64	-0.66	+0.32	+0.42	+1.49

*From Linsely and Crawford (1960).

The procedure appears to be basically sound for extending yield records to periods preceding that for which gaugings are available.

Since the greatest difficulty in a moisture balance approach, is in the determination of evaporation, additional papers will be mentioned, which although with a less refined approach than that of Penman, deserve consideration.

Langbein (1942) related monthly evapotranspiration to Field Moisture and Evaporitivity as determined from a U.S. Weather Bureau Class "A" pan. Field Moisture was related to base flow in a stream by a graphical correlation. Presented as it is, it is difficult to determine the degree of correlation existing. It is suspected, however, that the relationship is a unique one for the particular catchments and the rainfall pattern. It is possible to visualize a catchment subject to frequent light showers, insufficient to provide any ground water recharge, but maintaining a high level of evapotranspiration -- even though the base flow is very low.

Hursh, Hoover and Fletcher (1942) also presented a method for an annual water balance, in which evaporation was determined by Meyer's (1942) method and transpiration by weighting of mean monthly temperatures with the annual temperature. The procedure on the whole, is not to be recommended, even though for its time it was probably a reasonable approach.

Grindley (1960) used the moisture deficits calculated by Penman's method (max. deficit of 6.5 inches) and normal precipitation, to forecast the date at which streamflow would begin following the exceptionally dry summer of 1959 in the British Isles. Reasonably good agreement was obtained between forecast date and actual date, with this method.

4.2.8 Summary of Yield from Rainfall.

The simple two-variable linear regression of annual yield on annual rainfall is suited to cases where rainfall is high and is more than adequate to satisfy evaporation requirements at all times. It is suggested that, if $P \geq 3E_T$ on an average annual basis the relationship:

$Q = a(P - b)$ should prove quite reliable.

The curvilinear relationship between annual yield and rainfall is better suited to areas of lower rainfall (20 to 60 inches annually), because the effect of storm rainfall is better accounted for in the years of above average precipitation.

Multiple regressions of annual yield on seasonal rainfall provide a better estimate than when based on annual rainfall only, because the evaporative losses from rainfall and the subsequent recharge of storage losses follow a definite seasonal pattern.

The influence of precipitation from the antecedent year on annual yield depends primarily upon the ground water storage conditions and the choice of a water-year.

Calculation of annual yield, by taking account of temperature and evaporation are not likely to provide reliable estimates of yield unless the water balance approach is adapted and the accounting period is on a monthly or shorter term basis. The shorter term basis is required especially where high rainfall intensitis occur.

The biggest drawback to determination of yield from ungauged catchments, when rainfall data and other data may be available, is that year to year computations can be seriously in error from neglect of the changes in ground water storage.

The Water Balance Approach does provide the best method for determining Annual Yield, even though there may be a phase lag due to delay in ground water storage.

4.3 YIELD FROM STORM RAINFALL

On many drainage basins, particularly in arid and semi-arid regions most of the yield occurs as direct runoff from relatively few, long duration, high intensity storms. On small catchments short duration storms of high intensity are likely to be highly significant with respect to yield. Since such storms frequently have a rather irregular distribution with respect to time, any expression for yield in terms of precipitation and other variables on an annual or monthly average basis is likely to be seriously in error. The approach to the determination of storm yield based on daily and preferably hourly data on precipitation intensity and duration, infiltration capacity and soil moisture storage is essential. However, the variation in time and space of these factors is difficult to express mathematically.

Developments to date have been of an empirical nature for the most part, but deserve attention at this time. The traditional methods adopted for dealing with losses from storm precipitation in connection with unit hydrograph studies will not be discussed here since these are covered in most hydrology text books.

4.3.1 Moisture Balance.

The study of Linsley and Crawford (1960) is typical of this type of approach and has been previously described. The moisture balance was used to indicate infiltration capacity, it being assumed that infiltration capacity varied linearly with soil moisture from 2.50 inches per day at zero moisture to 0.30 inches per day with soil moisture of 13.0 inches.

Absence of hourly rainfall data in this study made it difficult to attain precision with respect to storm yield, probably accounting for some of the error in the estimates of monthly yield.

Linsley and Ackerman (1941) in a study of flood discharges on Valley River, N. Carolina, used a Moisture balance to determine residual for runoff, the infiltration being assumed constant for the storm duration of 0.05 inches per hour.

Total storm runoff was calculated as the difference between precipitation and soil moisture deficit <u>or</u> between precipitation and 0.05t where t = storm duration in hours. The larger of the calculated values was selected as being the storm runoff.

The method would seem to over-estimate values of soil moisture deficit and thus underestimate runoff. However most storms would yield direct runoff as a result of the limiting value placed on infiltration capacity, so that estimates of soil moisture deficit are not critical.

Guthe and Owen (1941) used a somewhat similar procedure for computing storm yield on the Upper Licking Watershed of Ohio. The method for computing soil moisture deficits was very crude.

Transpiration ratios (ratio of, pounds of water required to produce vegetation: ~~pounds of dry matter produced)~~ were used to determine the evaporation from vegetation. Transpiration ratios for any given type of vegetation vary by several hundred per cent depending upon soil fertility, insects, disease, variety, climate etc. Soil evaporation was treated separately in a rather vague manner.

Infiltration capacity was limited to a constant value of 0.40 inches per 30 minutes.

The study is of no particular value in providing a useful approach.

Nash (1960) in a study of an experimental catchment in England found a high correlation between storm runoff and rainfall:

$$Q = 0.80P - 0.125 \text{ inches.}$$

when the rainfall was in excess of the calculated deficit. The deficit was calculated by the method of Penman (1948) with a limit of 2.0 inches to the deficit. The deficit was assumed to be zero when more than 0.05 inches of runoff occurred.

The regression equation given had a coefficient of determination; $R^2 = 0.815$ and a standard error of estimate for Q of 0.065 inches.

The minimum rainfall of this category of storm, required for runoff to start according to the regression was 0.155 inches. However, examination of these data indicated that on several occasions the abstraction from rainfall, that is P - Q was less than this value, indicating that runoff could occur at rainfall amounts as low as 0.07 inches. In fact several storms did produce small flows even though the recorded rainfall was from 0.10 to 0.20 inches.

The significance of the 0.155 inches is probably that this represents the non-capillary porosity of the cultivated layer of soil. A reasonable estimate of non-capillary porosity for the London Clay of the catchment would be 0.02 inches/inch. A cultivated depth of 8 inches would provide the necessary storage. On occasions when runoff occurred with $(P - Q) < 0.155$ in., it appears that non-capillary water had not been drained from the profile from a storm preceding by less than one or two days.

It should be noted that non-capillary storage opportunity is significant in relation to ground water recharge. It would be interesting to compare the base flow in the case of Nash's (1960) investigation with the gross recharge of non-capillary storage in the cultivated layer. There was an average of 24 runoff producing storms per year, and if each contributed 0.155 inches (3.95 mm.) to ground water, then the average base flow rate would be 0.06 cubic meters per second. Nash quotes the base flow at 0.05 to 0.15 cubic metres per second. Unrecorded storms which produced no surface runoff but partially recharging the non-capillary storage capacity would increase the computed mean base flow slightly.

4.3.2 Antecedent Precipitation Index.

The use of an index value for antecedent precipitation is in a sense a moisture balance approach. Kohler and Linsley (1951) introduced the concept as a parameter in storm runoff studies. The index,

$$I = Io\ k^{t}$$

where Io is the index at time, t = o and k is a depletion constant, and t is in days. Corrections are made to the index for basin recharge due to subsequent rainfall.

The value of k depends upon potential evapotranspiration, being smaller in summer and higher in winter, indicating a more rapid reduction in the index during summer months than during winter months.

Linsley and Kohler (1951) overcame this difficulty by introducing an additional variable, week of the year, into a graphical multiple correlation, including also storm duration, amount of rainfall and finally the dependent variable, basin recharge.

The method is very useful, but requires a record from a large number of storms in order to develop a reliable relationship. It would seem possible to use developed relationships on a regional basis, providing reasonable similarity exists in catchment characteristics.

On small catchments, it would be necessary to adopt a procedure of determining runoff from increments of storm rainfall of an hour duration or even less. This was done by Miller and Paulhus (1957) in a study of Little Falls Branch Watershed in Maryland and Ralston Creek, Iowa. The A.P.Index was adjusted hourly and rainfall in hourly increments was used to determine hourly runoff.

The authors concluded however that a single graphical solution would not apply to both catchments, because of other catchment characteristics.

It may be concluded that this approach once developed for a catchment is useful, but application to other catchments is extremely risky, although there has been little actual study given to this phase of study.

McCutchan (1960)of Australia illustrated the effect of antecedent wetness on storm runoff for several majorstorms on the Dawson River of Queensland. He classified the antecedent catchment condition as Dry, Moderately Dry, Damp and Wet. We have plotted the data for two categories obtaining the following relationships :

$$\text{Wet,} \qquad Q = \frac{5.1^{P}}{15.4 - P} \text{ inches.}$$

Dry to Damp, $Q = \frac{1.8^P}{15.4 - P}$ inches.

For a given precipitation the wet conditions results in nearly three times as much runoff as the dry condition. The data are scarce, so the relationships should not be used as more than a general indication of a trend.

4.3.3 Infiltration Capacity.

No discussion of storm runoff would be complete without a mention of the 'infiltration approach' discussed by Cook (1946) and frequently referred to in hydrology texts. The significance has previously been discussed in connection with rainfall intensity. The infiltration approach has been used directly and implicitly in the previously quoted references discussed under 'Moisture Balance' (4.3.1) and 'Antecedent Precipitation Index' (4.3.2).

The infiltration capacity curve has a characteristic shape :

$$f_c = kt^n$$

where f_c = infiltration capacity at time t

t = time usually in minutes

K = an experimentally determined constant

n = a constant with value from 0 to − 1.

The attempts to determine f_c on a catchment basis, have depended upon simultaneous measurement of rainfall and runoff, with corrections for interception and detention storage. Success in this respect has been limited to very small areas.

However when it comes to applying experimentally determined values of f_c based on infiltrometer studies, to catchment yield from storms, the difficulties have been insurmountable. Soil properties are highly variable both in space and time. Therefore it appears that pending extensive further studies in this direction, it will be necessary to resort to greatly simplified models of infiltration in relation to rainfall intensity.

4.3.4 Summary of Storm Yield.

Success in making estimates of storm yield depend upon accurate information upon the time variation of rainfall intensity and infiltration capacity on a catchment scale.

Some success has been obtained by a combined consideration of the moisture balance and infiltration characteristics, but extrapolation of the empirical relationships to ungauged areas would be extremely hazardous.

This aspect of yield is very important to Australia since a great part of the annual yield is derived from high intensity storms on most of the continent.

4.4 DRY WEATHER FLOW

In this section the nature of flow during periods when ground water recharge is small or non-existent will be discussed. This normally occurs during the season of high evaporation, although it also occurs frequently in areas experiencing a winter freeze-up, while snow remains on the ground. Dry weather sequences of monthly or daily events are important in run-of-the-river projects such as hydro-electric and pollution abatement schemes. Sequences of dry seasons are important considerations in storage investigations.

4.4.1 Periodicity, Patterns and Persistence.

For many years, attempts have been made to prove that meteorological and hydrological sequences of events are repeated in a regular fashion. The term 'weather cycles' has captured and still captures the public imagination. From a short period of record it is quite easy to be misled into drawing a conclusion that extreme events recur at regular intervals. However, modern statistical knowledge has done much to show that recurrence of extreme events is of a random nature.

Even in recent years Girand (1932) and Streiff (1932) indicated their belief that streamflow followed the same cyclic time variation as the Wolf Numbers for sunspots. Girand contended that low flows were associated with maximum sunspots on an eleven year cycle. Streiff on the other hand noted two maximums and two minimums each eleven years for the Muskegon River of Michigan based on a record from 1908–1930 or a mere 22 years. A recent article by Williams (1961) presents additional data which in his opinion support the thesis of long term cyclic variations (a quasi-100 year cycle is suggested). He admits that there is some inconsistency in the pattern from one location to another on the earth. He suggests that there has been a progressive displacement with time of the cyclic behaviour in North America. His method of arriving at the conclusions is unconvincing.

The present state of knowledge in this field should be sufficient to discourage engineers from further attempts to predict future recurrence of extreme events on the basis of cyclic behaviour.

Of a somewhat more tangible nature is the study of tree rings to determine whether at some period in the past a more extreme event has occurred than any within recorded history. Girand (1932), Harding (1935) and Pender, Walsh and Anderson (1956) all mention examples in which tree ring growths have indicated probable extreme dry and wet periods exceeding those within recent recorded times. Caution must be exercised in the interpretation of growth rings since many factors other than moisture can influence the annual growth of a tree. It is suggested that expert foresters should be consulted on such matters.

Accepting the fact that annual rainfall events are independent, it still remains that there is equal probability of a given dry year being followed by a dry year as a wet year. Sequences of dry years can and do occur, varying in duration, intensity and frequency, in a random fashion. The influences of such sequences upon streamflow are modified by the storage delay or lag in ground water discharge to the stream or to a lesser extent the discharge from surface storages.

Wahle (1943) describes in a qualitative manner the influence of years of abnormally low discharge on succeeding years for the Columbia River in the United States and Canada. Other things being equal a year of low discharge is likely to be followed by a year of rather low discharge.

McDonald and Langbein (1948) examined this trend more thoroughly for the same region, when a diminution of streamflow over a fifty year period was noted. In discriminating between short term and long term precipitation effects an <u>effective precipitation</u> was calculated as follows:

$$P_e = \frac{P_o + KP_1 + K^2P_2 + K^3P_3 + \ldots\ldots\ K^nP_n}{1 + K + K^2 + K^3 + \ldots\ldots\ K^n}$$

where P_e = effective precipitation

P_o = precipitation for year of streamflow

P_1, P_2, P_3 P_n are annual precipitation

for 1, 2, 3 ...... n years preceding.

By setting values of K of 0, 0.1, 0.2 .... 0.90, 0.95, 0.98 in turn and correlating the computed P_e with annual streamflow the relative short term and long term effects were determined. Low values of K corresponded to short term effects while high values correspond to long term effects.

An example of combined short term and long term influences is presented for the Metolius River of the Columbia Basin in the form of the regression equation of mean annual discharge :

$$Q = 12.8\ P_{0.2} + 89P_{0.98} - 2650 \text{ c.f.s.}$$

$P_{0.2}$ is the effective precipitation based on K = 0.2

$P_{0.98}$ is the effective precipitation based on K = 0.98

The partial coefficients of determination were respectively 34% ($P_{0.2}$) and 37% ($P_{0.98}$) indicating the short and long term influences were almost equally predominant. The geology of the basin is the determining factor in the relative short and long term effects. The Metolius River basin is in basalt with a storage delay time of 3.9

years, which is extremely long.

Hoyt and Langbein (1944) in a broad geographic study of the yield of streams as a measure of climatic fluctuations in the United States and Canada found that there was a slightly greater tendency for persistence of low flows than of high flows, but concluded that it was not more frequent than would be expected on the basis of random occurrence. However, there were certain of the six geographic regions where the persistence was more pronounced. The North Central Region in particular had a higher persistence of both wet years and dry years. The authors suggested that this may be typical of a continental climate. The persistence of annual precipitation at St. Paul, Minnesota, was of the same order as for stream flow of the region.

Kritsky and Merkel (1960) in the U.S.S.R. state that observed sequences of dry years seem to refute the contention that annual flow values may be considered as statistically independent variates. The statement is based upon the derived correlation coefficient for consecutive annual streamflows which averaged, R = + 0.24 for a number of streams. Since the size of sample is not stated it is difficult to assess the significance of a coefficient of such a low order of magnitude. It does appear however that the apparent positive serial correlation is little greater than might be expedted from a random distribution.

Wemelsfelder (1960) discusses the persistence of monthly discharges on the Rhine River at Lobith, near the Netherlands border for the period 1919-1959. He found that there were durations of above average and below average streamflow much longer than that to be expected from straight probability analysis or even on the analysis of precipitation for the same region and time. Precipitation patterns of persistency followed probability theory very closely. On the basis of probability there shouldhave been only 8 periods with six or more months in succession of either submormal or abnormal streamflow. In fact there were 22 periods of six months or more duration of above or below average streamflow for the months concerned.

It is significant also that there was a larger number of low flow persistencies than high flow persistencies. There were 196 months in series of negative (low flow) persistencies and 110 months of positive persistencies. This is to be expected because the depletion rate at high levels of discharge is much greater than at low levels. Also at low levels of discharge, precipitation much above average (with a relatively low probability) is necessary to recharge underground storage and return streamflow to above average values.

The tendency for a given flow to persist or to be reduced still further is utilized in a practical manner in making forecasts of low flow for navigation on the Mississippi River. Mann and Rasmussen (1960) outline a method adopted for making 28 day forecasts based on a co-axial graphical correlation of antecedent precipitation, calendar date, normal and expected rainfall. Weekly, 28-day forecasts of minimum flow based on no precipitation are made from this multi-variable relationship. The relationship is still to be fully tested.

4.4.2 Depletion of Ground Water Storage.

A number of expressions are used to indicate the depletion of ground water or the recession curve of discharge during rainless periods. Barnes (1939) in a discussion of the structure of discharge recession curves proposed :

$$Q_t = Q_o K_r^{\,t}$$

where Q is the discharge at time t

Q_o is the discharge at time t = o

K_r is the recession constant

Wisler and Brater (1959) quote Horton as the source for the following form :

$$Q_t = Q_o \; e^{-ct^n}$$

where c and n are constants.

Obviously the two forms are different since :

$$K_r = e^{-ct^{n-1}}$$

hence being a function of time. This is indeed pointed out by Linsley, Kohler and Paulhus (1958, p.155). Nevertheless the simpler form is preferred because of its ease of derivation.

That K_r varies with time is probably an indication that depletion rates from various elements of ground water storage are different. Troxell (1953) mentions three types of storage involved in ground water discharge from the San Bernardino and San Gabriel Mountains of Southern California. These are (1) perennial ground water; (2) seasonal ground water; (3) storm ground water.

Storage Ratios (Storage : Mean annual Discharge) were computed as follows:-

	Upper Mountain Areas	Mojave River
Perrennial Storage	1.85	0.43
Seasonal Storage	0.61	0.17

The differences, probably attributable to geological structure are evident. The lower values for storage ratio for seasonal storage are attributable to smaller values of K_r and steeper recession curves. The more rapid drainage of seasonal storage could also be due to higher hydraulic gradients and a more permeable transmission zone near the ground surface.

Values of the recession constant are likely to vary widely for a given catchment from season to season, depending upon the evaporative losses from areas of shallow water table. Likewise areal distribution of rainfall may be such that certain elements of ground water storage on a catchment are at storage levels which provide for discharge rates quite out of proportion to that from the other storage elements. Since the streamflow hydrograph represents the integrated effect of outflow from all storage elements, the more complex the storage pattern the greater the range of recession constants which could be indicated.

Dooge (1960) distinguishes between four main types of storage elements, (1) the deep water table type, and (2) shallow water table element (negative outflow not possible); (3) shallow water table element (negative outflow possible), and (4) composite water table element.

In the case of the deep water table element, recharge at the water table surface is either positive or zero. Lateral recharge from another element in series is also possible.

Shallow water table elements may be recharges positively from above by precipitation or have negative recharge due to evapotranspiration. Where negative outflow is not possible negative recharge continues after outflow has ceased, so that discharge from such an element will not take place again until recharge from precipitation makes up the storage depletion by evapotranspiration to the zero flow storage level.

For a shallow water table element with negative outflow the storage level is maintained at a minimum storage level by backflow from the stream, sufficient to supply evapotranspiration requirements. Positive outflow takes place at the inception of positive recharge.

Composite elements behave like shallow water table elements for a part of the time until the storage drops to a critical stage after which it acts like a deep water table element.

The designations are arbitrary and of course are difficult to distinguish on a catchment basis. They do form a useful basis for formulating a model system for a catchment composed of various types of storage elements in parallel and/or in series.

An interesting low water curve is presented by Indri (1960) for three streams in the Venetian Alps of Italy :

$$q_i = \frac{1}{at_i} + q_o$$

where q_i is the flow in m^3/sec after a duration of t_i days

a = a stream constant

q_o = discharge at an infinite time with no rainfall.

The curve is of the same general shape as the previous recession curve and does not appear to offer any particular advantages, although the values of a and q_o will be given here as a matter of interest.

	a	q_o
Callina River	0.0025	2.00
Boite River	0.0048	2.00
Cismon River	0.0032	1.35

4.4.3 Effect of Rainfall on Recession Flow.

Completely rainless periods of long duration are the exception rather than the rule. It would appear useful therefore to have a means of synthesizing a hydrograph based on superposition of rainfall on a regular recession graph. For it is found that even light rainfalls, insufficient to result in surface runoff or ground water recharge do change the character of the recession graph, probably as a result of the arrest of direct evapotranspiration from shallow water table elements.

Riggs (1953) developed a regression equation for base flow discharge following a storm producing peak :

$$Q = 0.00184 \; Q_o^{0.943} \; Q_p^{0.713}$$

where Q = base flow at time of peak discharge (determined by projecting new recession curve back).

Q_o = base flow preceding the hydrograph rise.

Q_p = increase of peak discharge over base flow projected to time of peak.

The equation is probably unique for the Clark Fork River at St. Regis, Montana. The coefficient of determination was, $R^2 = 0.906$ with a standard error of estimate for the 19 stream rises used in the regression of from 31 per cent above to 24 per cent below the mean value.

Roche (1960) developed a series of equations for monthly discharge from July onwards as a function of discharge on July 1st and subsequent precipitation for the Namorona River Basins of Madagascar.

For example, for the month of August the mean discharge is :

$$Q = 0.49q_o + 0.125P_1 + 0.05P_2 - 1.7 \text{ m}^3/\text{sec.}$$

where Q = Mean August discharge (m^3/sec)

q_o = July 1st discharge

P_1 = August precipitation (mm)

P_2 = July precipitation (mm).

No correlation coefficients are given for these equations, so it is difficult to evaluate them.

One additional feature of Roche's presentation was his method of predicting July 1st discharge :

$$q_o = 0.765x + 0.131$$

where x is the mean discharge for May and June. The coefficient of correlation was 0.997.

Additional regressions are also presented for computing May and June mean discharge from precipitation only for current and three preceding months.

4.4.4 Routing of Ground Water Recharge Elements.

The types of water table elements through which recharge may be routed have previously been discussed. Dooge (1960) and (1961) describes and illustrates a routing procedure for elements of recharge which may be used to estimate low flows.

$$Q_n = C_o R_n + C_1 R_{n-1} + C_2 Q_{n-1}$$

where C_o, C_1, C_2 are routing coefficients

Q_n and Q_{n-1} are respectively the outflow from storage during current and preceding time periods.

R_n, R_{n-1} are respectively the recharge for current and preceding time periods.

The development utilizes the principle of linear storage elements, so that superposition of recharge elements may be effected and the output responses to uniform ground water recharge rates determined.

Linear storage elements have the property :

$$S = Kq$$

where S = storage above or below that at which outflow is zero

q = outflow rate from storage element

K = storage delay time of an element.

The routing coefficients are functions of K and the unit of time selected, and values of the coefficients are listed for various ratios of K/T in Dooge's (1960) publication.

The method presented by Dooge is indeed a much more sophisticated approach than any previously suggested. There are however certain disadvantages with the procedure which may inhibit its practical application.

First a period of record (although a short one of 2 to 5 years) is still required to establish the value of K, in $S = Kq$. Then the value of K is only an equivalent one for the combined storage elements of the catchment.

Failing this, trial and error with various storage elements in series and/or in parallel assuming different values of K, can be used to synthesize a flow record to determine the model most closely representing the catchment.

Having set up an appropriate model, however, it is possible to proceed to synthesize records for other periods providing the necessary data are available on precipitation, evaporation etc.

The assumption of uniform recharge rates is probably not a serious deterrent to determining minimum flows, but could lead to serious errors unless short time periods (one day or less) are chosen. This would increase the computation considerably.

Finally, although the assumption of linearity in storage-discharge relationships is certainly a valuable aid to computation, as has been previously pointed out, (3.5.5) the assumption is only approximate and valid over short segments of the S vs. q curve.

However, since the method is new and untried, criticism should await evaluation on an intensive scale. Its greatest application would probably be, to small catchments with low intensity rainfalls such as occur in the British Isles.

4.4.5 Regional Correlation with base stations.

The use of satellite stations for the purpose of extending stream flow data to ungauged catchments has been proposed by Langbein and Hardison (1955). The difficulty in providing stream gauging facilities for all catchments where water development potential exists is virtually insurmountable. Even the smallest catchments are important for farm and ranch water supplies.

The use of satellite stations operated for a few years (5-10) in the vicinity of base stations with a long period of record, would enable correlations to be established which would in effect result in long period records of reasonable reliability being synthesized. Such an undertaking would no doubt result in great economies and large returns in hydrologic data which could not otherwise be attained. It is unfortunate that in most countries of the world insufficient attention has been given to the establishment of hydrologic networks. Agencies charged with responsibilities have too often been obsessed with the immediate necessities of their particular phase of water control and utilization.

A number of engineering studies have of necessity utilized correlation techniques to establish yield parameters on streams of short period records. Brown (1961) in dealing with streamflow correlations in the Snowy Mountains, found a gross correlation coefficient of $R = 0.979$ between the Crackenback River at Creel (a tributary of the Snowy River) and the Snowy River at Jindabyne, a short distance below the gauging station at Creel. The drainage areas were 97 square miles from Crackenback and 716 square miles from the Snowy River. The two catchments are quite different in many respects, the Snowy being heterogeneous as to elevation and exposure to weather systems while the Crackenback is intermediate in elevation and more homogeneous in character. The high correlation may be to some extent fortuitous in this case.

A similar correlation between the Snowy River at Island Bend (elevation 4,000 ft. plus) and the Snowy at Jindabyne (36 per cent elevation 2,000 ft. - 4,000 ft.) showed a much poorer correlation ($R = 0.907$) between monthly runoff values at the two stations on the same River.

The success of this type of correlation will depend a great deal upon the similarity of the catchments and meteorological homogeneity of the area. The introduction of additional variates into the correlation which can account for any differences will improve the estimates, but care must be exercised in so doing to ensure that the additional variates are truly functionally related to the dependent variable.

It is not possible to consider fully all aspects of Correlation Analysis in this discussion. However it is suggested that the reader could refer to Ezekiel (1941), Langbein and Hardison (1955), Ford (1949), Brown (1961) and Stidd (1953) in addition to any of several Statistical text books.

The use of a tributary as an index of flow in the main stream is beset with the problem of seasonal lag of the main stream behind that of the tributary. This is recognized by Cooperrider, Cassidy and Niederhof (1945), in developing the forecast equation for seasonal flow in the Salt River of Arizona, as a function of the antecedent flow in Parker Creek.

$$Q_s = 1488\, Q_p + 98{,}726 \text{ acre-feet.}$$

where Q_s = March to May inclusive discharge on the Salt River.

Q_p = Oct. - Feb. discharge of Parker Creek.

The value of the correlation coefficient was $R = 0.9756$, with a standard error of estimate of 68,000 acre-feet.

4.4.6 Summary of Dry Weather Flow.

There is greater evidence of persistency in streamflow than in rainfall due to the storage lag of ground water flow.

Persistencies of low flows are more pronounced than high flows. There is scant evidence of periodicity in flow events although efforts to show this have frequently been made.

The ground water depletion curve normally adopted, $Q = Q_o K_r^t$ is in error because it can be shown that K_r is not a constant over the full depletion range. Seasonal variations of evaporation, geology and extent of area with swampy or marshy vegetation are factors influencing the value of K_r. Rainfall during the recession period, even though not recharging ground water may result in the arresting of evaporation from shallow water table elements, thus sustaining streamflow.

The method of Dooge (1960) for routing ground water recharge through linear elements of storage may prove to be a useful technique for areas of low rainfall intensity, although a short prior record of gauging would be desirable before utilizing the technique to synthesize a low flow record.

The use of satellite and base gauging Stations is a valuable method for extending streamflow data over longer periods of time. It should be given serious consideration by authorities charged with stream gauging responsibilities.

4.5 GEOLOGY AND STREAM YIELD

Most of the work relating to the influence of geologic formation on cathcment yield has been qualitative in nature. Mead (1950), pp. 363-410 recognised the influence of geological formation. Jarvis (1931) commented upon the influence of soil particularly upon storm runoff as have several others. The failure to draw quantitative conclusions with broad applicattion may be attributed to the complex interaction of vegetation, land use and soils and the fact that the variations of bydrologic properties within broad geological classifications is often greater than between classes.

A consideration of geology involves two hydrologic aspects, one, the influence of the surface mantle - the soil, and the other the effect of the subsurface rocks. The influence of the first is important relative to storm runoff, while the latter is important in its effect upon ground water discharge and flow during droughts.

4.5.1 Soil Properties.

We have already discussed infiltration capacity, so discussion of soil properties will be limited. Lvovitch (1957) in a discussion on stream-flow formation factors in Russia quotes the following relative discharge coefficients, $C = \frac{Q}{P}$, for storm runoff for some of the great soil groups.

Solonetses and solonchaks.	100
Degraded podsol clay and clayey soils.	80 - 85
Chestnut soils.	65 - 70

Clay and clayey chernozems of high fertility.	40 - 50
Sandy soils.	20 - 35

Krimgold and Beenhouwer (1954) in a paper devoted primarily to infiltration properties for irrigation design tabulates infiltration capacities according to the great soil groups. Other work has been reported by Free, Browning and Musgrave (1940) Sharp, Holtan and Musgrave (1949).

Smith (1955) presents the following figures on yield in connection with small pond design;

Effect of Soil on Catchment Yield.

	Moderately Permeable.		Slowly Permeable.	
Return Period (yrs)	Inches/yr.	Inches/2yr.	Inches/yr.	Inches/2yr.
10	0.71	2.15	1.64	6.10
25	0.43	1.46	0.97	4.30
50	0.32	1.14	0.69	3.43
100	0.24	0.91	0.51	2.80

The depth of soil over rock has an important bearing on storm runoff. Hoover and Hursh (1943) attribute the greater discharge from watersheds at the higher elevations of the Coweeta Experimental Forest of North Carolina to the shallow soils present. Catchments at lower elevations have talus deposits in the ravine slopes which have a high detention capacity.

The non-capillary porosity of a soil has a pronounced effect upon the distribution between surface sunoff, interflow and subsurface flow. Ayers and Wikramanayake (1958), found in plot tests that the mass infiltration as a function of storage capacity in the upper six inches of a loam soil was expressed as:

$$Y = 0.43 + 0.38X \text{ inches.}$$

where Y = mass infiltration in 20 minutes of precipitation excess,

and X is available moisture storage in the upper six inches of soil in inches. Reinhart and Taylor (1954) present a similar equation for mass infiltration for 30 minutes which is:

$$Y = 0.50 + 0.45X$$

4.5.2 Subsurface Deposits.

The unconsolidated deposits of sand and gravel are likely to be most reliable for yielding a sustained stream flow. The continuity and extent of such deposits and the presence of good recharge opportunities of course is of great importance.

Cross (1949) in a study of dry weather flow in Ohio, characterized streams by the discharge per square mile exceeded 90 per cent of the time. Areas of interlobate moraines and buried glacial valleys filled with permeable gravels, in addition to kame terraces, kames, eskers and outwash deposits all had high rates of discharge. Glacial till and unglaciated areas had low rates.

Lane and Lei (1950) had similar findings, pointing out that the variability index I_v was low (0.15 to 0.35) for glacial moraine areas of Wisconsin, compared to clay till areas in the same state with $I_v = 0.53 - 0.71$.

Similar trends were observed in several other States covered in the study embracing 220 stream flow records of the north-eastern United States.

Sedimentary deposits of sandstone and limestone are likely to be the most favourable of the consolidated rocks for sustained stream flow.

Both are likely to be highly variable however. The Bunter and New Red Sandstones of England are reputed by Legget (1939) to yield very well. Clark (1955) reported a good correlation between the dry weather flow of Pond Creek in Oklahoma with ground water levels in the Rush Springs Sandstone of that area.

Tison (1960) attributes disagreement between computed values of discharge coefficient, $C = Q/P$ based on Turc (1954) and Wundt (1937) and the actual value on an annual basis for a number of streams in Central Africa, to the very high permeability of those particular basins. In the case of the highly permeable catchments computed coefficients were in the order of $C = 0.30$, while the actual values were approximately 0.45. Furthermore in general the ratio:

$Q_o/Q_m \simeq 1$ for the very permeable basins and

$Q_o/Q_m < 1$ for less permeable areas where Q_o = flow at beginning of recession period, and Q_m = mean annual discharge.

Amico (1954) drew the same conclusion for the four rivers in Italy.

Limestone terrains can be very unpredictable. Whereas the yield from sandstone depends primarily upon the porosity of the rock, in limestone the presence of solution channels enhances the transmission of water. Sinkholes in "karst" topography may collect considerable amounts of surface drainage to be transmitted through complex underground passages often far beyond the catchment upon which the preciditation falls.

Igneous rocks are usually a poor source of water for streamflow. There are however notable exceptions. Stafford and Troxell (1944) attribute the higher sustained yield of the San Antonio

Creek and Lytle Creek in the San Bernardino and San Gabriel Mountains of California to the deeply fractured granite on these watersheds as compared to Strawberry Creek, Deep Creek and the Mojave River in the same locality.

A similar experience is reported in the United Kingdom by Thomson (1923) in comparing the dry weather flow for 30 streams. The River Avon of Scotland had a dry weather flow of 1.28 c.f.s. per square mile while the river Alwen of North Wales had a low flow of 0.035 c.f.s. per square mile. The catchments were reported to be similar climatically, topographically and in size. The Avon was wholly granite, deeply fractured, while the Alwen consisted of Silarian shales and grits covered with boulder clay.

Clifford in the discussion of a paper by Justin (1914) cited the very high percentage of discharge from the volcanic lava beds of the Pacific Coast in North America. The percentage discharge of the North Fork of Feather River was C = 61% (P = 59 inches) and C = 82% (P = 23 inches) in 1910-11, and 1911-12 respectively.

Basalt of the Metolius River of the Columbia Basin is quoted as having a sufficient ground water storage that discharge would only drop from 1800 c.f.s. to 750 c.f.s. in four years in the absence of recharge according to McDonald and Langbein (1948).

Serra (1954b) after removing the effect due to slope on five rivers in Southern France determined the following relative coefficients $C = Q/P$:

Basalt,	C = 0.81
Granite,	C = 0.635
Moraine,	C = 0.17

4.5.3 Summary of the Influence of Geology.

Surface mantle of soil is important with respect to storm runoff, but soil properties are so complex and dynamic that a simple cause and effect relationship is difficult to develop.

Unconsolidated deposits of sand and gravel provide high volumes of storage and if extensive are highly effective in assuring a uniform flow.

Sandstone and limestone are frequently effective but are highly variable. Well fractured granites on occasion transmit considerable quantities of water to streams. Most igneous rocks in addition to sedimentary deposits of shales have low storage and transmission characteristics. Basalt deposits often have a relatively high storage capacity and assist in maintaining uniform stream flow during dry periods.

4.6 LAND-USE EFFECTS ON YIELD.

Much has been written about the effect of land use on catchment yield, but unfortunately a great deal of it has been conjecture unsupported by scientific fact. The controversy has related largely to (1) the influence of forests and (2) the influence of intense farm cultivation.

4.6.1 Agricultural practices.

Lvovitch (1957) of U.S.S.R. reports that autumn plowing decreased runoff to one-fifth in the Transvolga Region. Apparently this would be in the region of winter freeze-up and major runoff from spring snow melt. Similar findings have come from investigations reported by Hays (1955) at Wisconsin and have been observed by the writer in Southern Ontario in Canada. Apparently plowed land is more receptive to water than sod during the melt period, since hay or meadow land remains frozen for a longer period of time. Hays (1955) reports that highest storm runoff was from spring planted grain followed by first year hay land and corn land in that order. Older hay stands (2 and 3 years) yielded very little runoff from summer rains compared to corn land and spring grain. Soil vegetative cover is apparently the governing factor in maintaining high infiltration capacities on hay land. Young (1948) quoted tentative relative water yields in descending order for the following cultural practices in the Ozarks of Arkansas on freely permeable soils: cultivated crops, terraced meadow, strip cropping, pasture and wooded lands. This would be south of the winter freeze-up zone.

Sharp et al (1960) however tested by regression analysis the significance of several factors on the annual yield from the Delaware River Basin of Kansas and found none of the following to be significant factors: percentage of row crops, miles of terraces per square mile, average percentage of normal pasture at start of month, pasture condition.

4.6.2 Forest Influences.

Numerous investigations have been made in the Coweeta Experimental Forest Area of North Carolina and in the Western United States.

Hoover (1944) reported that stream yields were increased by 16.74 inches and 10.68 inches respectively in 1941-42 and 1942-43 following the complete cutting of forest vegetation from watersheds at Coweeta. The watersheds had been compared prior to cutting, so the increases are based on the probable yield of the treated watershed, had it remained in forest cover. The area is one classified as super-humid, according to the Thornthwaite (1948) system. Only about two inches of the total of 70 inches of precipitation occurs as snow. The increased streamflow is attributed to the reduction in evaporation as a result of the removal of deciduous trees. Lieberman and Hoover (1951) reporting on the flow distribution from treated catchments at Coweeta found that the median value in the frequency distribution was increased by a factor of two when vegetation was cut and regrowth was prevented. This partially dispels any concern that increased yields result from flood flows at the expense of lower flows. The experimental evidence regarding the stream flow frequency-distribution when forest areas are denuded is meagre however, so that generalizations are not in order.

Lvovitch (1957) states that an increase in forest cover from 45% to 75% had decreased the annual discharge from 65 per cent of annual rainfall to 35 per cent of the annual.

Croft and Monninger (1953) report similar results from the Inter-mountain Forest Experiment Station of Ogden, Utah. However, the experiments were carried out on 1/10 acre plots and do not appear to have been properly replicated. Increased stream yields were assumed to have resulted from deep seepage as estimated from precipitation and soil moisture measurements.

Rennie (1957) in England presents data which tends to refute the work in the United States. He found that seasonal soil moisture deficits in an afforested Calluna moor in Yorkshire were 5-9 centimeters of water less than calculated. From this he concluded that stream flow would not be affected to the extent anticipated due to the lower level of evapotranspiration experienced. His conclusions are not valid however because of a shallow depth of moisture sampling and the absence of a suitable unforested check area.

Johnson and Meginnis (1960) confirm the effects reported by Lieberman and Hoover (1951). Median values of discharge were higher for cut-over areas. The gain in summer flow was 0.55 inches for the cleared area and the discharge rates were consistently higher through the recession period.

Love (1955) reported that the killing of a stand of pine and spruce by Engelmann spruce beetle on the White River in the Western United States, resulted in annul discharge being increased by 2.31 inches, over the period prior to the 1941-46 infestation. The period of gauging prior to the outbreak was only 4 years, but the author was able to correlate the discharge with that of nearby Elk River whose catchment was not affected by the outbreak.

4.6.3 Effect of Forest Fires and Erosion.

Most of the work cited has been carried out experimentally where the forest floor consisting of decayed and partially decayed forest residue was not disturbed. Anderson (1955) has shown that serious wildfires on the Santa Ynez watershed of California resulted in increased peak discharges and sediment yield with little change in annual discharge. The fires probably destroyed a considerable portion of the organic residue, thus exposing the soil to the erosive influences of rain and degradation of soil structure. Infiltration capacities were probably reduced so that larger proportions of surface runoff occurred at the expense of ground-water recharge.

Similar increases in flood peaks were noted by Hertzler (1939) in North Carolina from catchments on abandoned farm lands, (eroded), overgrazed pasture and denuded peaks when compared with forest areas in the same region.

4.6.4 Forest Management.

Since there is considerable evidence to show that reduction in forest vegetation results in increased stream flow, it is natural that investigations should be carried out on controlled cutting to optimize water yields and timber yields.

Anderson and Gleason (1960) report additional storage of moisture in the soil where strips of a width twice the tree heights were cleared in the Cascade and Sierra Nevada Mountains. The soil moisture saving indicated was eight inches due to a higher snow water equivalent in the cleared areas. In a selectively logged area where all trees of less than 18 inches diameter were retained the moisture retention was three inches greater than in the natural stand. However it is difficult to interpret the findings on a catchment basis.

Bailey and Copeland (1960) compared the yield of two catchments, the Parrish and Centreville in the mountainous areas of Utah. The former prior to 1937 was badly eroded following destructive fires. Mechanical and cultural treatments applied to the Parrish catchment were as follows:- (1) fire control; (2) elimination of livestock grazing; (3) seeding of grasses on depleted but not seriously eroded areas; (4) construction of contour trenches on the most barren and gullied area with sufficient storage for these areas only, for 1.5 area-inches of precipitation. The treatments have resulted in a total decrease of yield over 22 years of 2.70 inches (mostly during the first eleven years).

Peak discharges have also been reduced but curiously enough the June discharges have increased slightly, although probably not significantly. The Centreville catchment had no history of vegetation destruction or erosion and showed no time trend in the stream yield per unit of snow-water content as determined by survey.

4.6.5. Summary of Land Use Effects.

While much work has been done on this phase of yield, a considerable amount of the literature is coloured by opinion based on meagre observations.

The more vigorous the plant growth the greater the evaporative loss and generally lower the water yield.

The optimum condition for water yield appears to be denudation of catchment of vegetation, while protecting the soil from erosion. This is obviously impractical in many cases, so there is an obvious challenge to optimize water yield with forest and farm production.

5. CONCLUSION.

The major approaches and principles have been summarized at the conclusion of relevant sections of the report. The nature of the subject precludes presenting a simple concluding statement from the survey, unless it is the timeworn phrase "more research is required".

The content of the Survey is notable for the apparent weaknesses and shortcomings in our knowledge of the hydrologic principles of yield determinations. If these are recognized and appropriate well planned investigations are instigated, one major purpose of the survey will be served. In this aspect of hydrology, more than any other, the intimate association of physical and biological factors is evident. While remarkable advances have been made in particular phases of the subject, there is considerable scope for investigations embodying the application of physical and biological principles to the catchment as a unit. It would seem essential that more emphasis on a unified approach based on the proper interpretation of the physical processes will in the end result in substantial progress.

For those concerned with immediate problems involving yield from ungauged catchments or where records are scarce, the Survey will provide no simple solutions. The nature of the project, extent and reliability of pertinent data on site or from neighbouring catchments will dictate the best approach. As in most hydrologic studies alternative approaches are often available. These alternatives should each be applied to the greatest extent possible. The Survey outlines approaches used by other investigators and attempts to point out the limitations and weaknesses. The balance is left to the judgement of the hydrologist or engineer.

While the impression may be gained that the author has sought a substitute for stream gauging measurements, this has not been the intention. On the contrary, expansion of stream gauging networks, strategically planned with greater foresight and coordinated with meteorological networks should ever be the aim of those charged with this responsibility. Financial and human resources must be utilized in this important task with the economic return clearly in sight.

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