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EVALUATION OF COMPONENTS OF WATER BALANCE OF A RIVER REACH

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1987-88

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ABSTRACT

The water balance has been used for computing seasonal and geographic patterns of irrigation demand, soil moisture stress , prediction of streamflow and water table elevations. Although the predictions may be approximate, they are sufficiently accurate to indicate whether a scheme is hydrologically sound or not. The water balance requires that the items of supply balance the items of disposal and storage. In nature a balance is always maintained. This generally manifests itself in the form of floods, water logging, drought, etc. which may be deterimental to human activities.

The water balance of a river reach is closely inter-related with surface water and ground water. It requires quantification of inflow, outflow and storage. Some of these elements could be directly measured while others need to be estimated. There may be noticable difference between the natural inflow and natural outlfow in a river reach over periods of a month, season or an year.

It is useful to classify the river reach into alluvial or non-alluvial so that hydrologic characters can be defined and determined. Bank storage effects have to be properly accounted. Six components (viz. precipitation, inflow from the upstream, storage, diversion, outflow, and ground water component and bank storage) have been identified and dealt with briefly in this technical note. Possible methods of evaluation of each one of these are mentioned.

1.0 INTRODUCTION

The main objective of water balance computation is a more precise evaluation of water resource especially in a poorly gauged watershed. Water balance computations make it possible to reveal and to correct all systematic measurement errors, besides enabling the planning of water resource projects more correctly. The water balance has been used for computing seasonal and geographic patterns of irrigation demand, soil moisture stresses, the prediction of streamflow and water table elevations. It has also been used to find the water flow to lakes and therefrom water level and salinity. Although the predictions may be approximate, they are sufficiently accurate to indicate whether a scheme is hydrologically sound or not.

The term water balance was used in 1944 by a meteorologist, C.Warren Thornthwaite to refer to the balance between the income of water from precipitation & snowmelt & the outflow of water by evapotranspiration, ground water recharge and stream flow. The water balance is defined by a hydrologic equations which is basically a statement of the law of conservation of matter as applied to hydrologic cycle. It states that in a specified period of time all water entering a specified area must go into storage within its boundaries or to go out therefrom. The water balance requires that the items of supply balance the items of disposal. There may be a marked difference between the natural inflow and natural outflow, over a month, season or even an year, in a river reach.

A balance will be maintained naturally. But at times

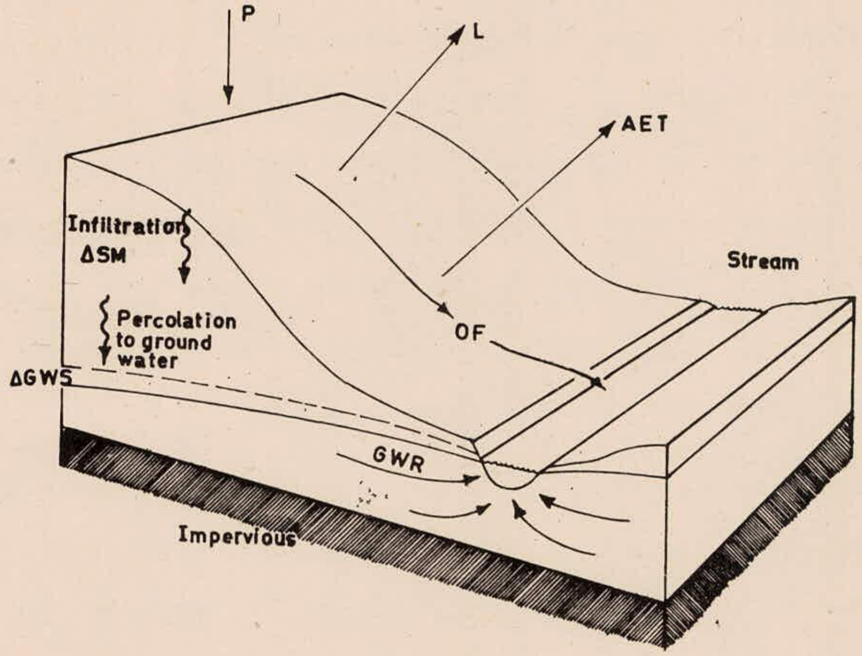

Fig.1 : Components of Water Balance

may not be to the advantage of human beings, as for example floods, water logging droughts, etc. Avoiding such inconveniences or calamity is a prime objective of water management.

The water balance of a river reach is closely interrelated with surface water and ground water. It requires quantification of all items of inflow to and outflow from the river reach, as well as the changes in storage. Few of these items are directly measurable, but some must be estimated It is useful to classify a river reach for example alluvial, mountainous, etc. so that hydrologic characters can be easily determined. Soil characteristics and nature of vegetation have important bearing on the extent of drainage within catchment covered by the river reach.

In general, water balance in a river reach can be categorically studied for monsoon, non-monsoon and for high flow period. The other time periods considered are daily, weekly, 10 daily, monthly. For high flows hourly or even less time periods can be used.

In the monsoon and high flows, water level in the stream channel may be higher than the water table. The stream is an influent. During non-monsoon season the stream will be an effluent. During flood flows, the fast raising water level in the channel cause storage of water temporarily in the banks and depletes at the falling stages. This is more pronounced in the case of alluvial river reaches.

2.0 REVIEW

Water balance expresses the law of conservation of matter as applied to hydrologic cycle. All water entering in a specified period of time and area must go into storage or to go out therefrom.

The water balance of a small drainage basin with underlying impervious rock at depth can be represented by Fig. 1 and expressed as follows:

$$P = I + AET + OF + \Delta SM + \Delta GWS + GWR \qquad \ldots(1)$$

Where,

P = Precipitation

I = Infiltration

AET = Actual evapotranspiration

OF = Overland flow

ΔSM = Change of soil moisture

ΔGWS = Change of ground water storage

GWR = Ground water runoff

All the components are expressed as equivalent depths of water for the same time interval. With reliable independant estimates of evapotranspiration, infiltration, etc., it is possible to abstract much hydrological information, from the law of conservation of mass or the water balance. The water balance for a drainage basin states that the precipitation equals the sum of the runoff, evapotranspiration and change in storage. It has been possible to routinely observe only two of these four quantities. The measurement of changes in storage on a weekly, monthly, seasonal basis is too expensive to practice except in a research situation, because of the hetrogeneity of topography, soil, vegetation in most of the drainage basins. After the completion of International Hydrological Decade i.e. 1974, a great deal of information was generated on the water balance analysis. Franke and McClymond(1972) presented an analysis of effect of urbanisation on water budget of long Island (U.S.A). Al-Khashab (1958) used water balance to define the seasonal and geographical pattern of water supply and irrigation demand at an early stage in the planning for the development of the water resources of the 785,000 square kilometer Tigris-Euphrates basin in Iraq. Boersna (1967) indicated as to how water balance can be used for predicting water table.

Using CRAE (Complimentary Relationship Areal Evapotranspiration) model, Morton (1980), mean seasonal water balance for six North American basins were made.

These seasonal water balances shown in Fig. 2 are based on monthly mean values of observed precipitation, observed runoff & the CRAE model estimates of evapotranspiration for five or more integral years (Morton, 1983). The upper line (Fig. 2) represents a mass curve of precipitation. The second line represents a mass curve of evapotranspiration. The third line is of mass curve of storage changes. This provides a comparison of seasonal water balances for river basins of different environment.

Casenave (1978) studied the water balance of Sanguere basin, and concluded that the unmeasured ground water outflow which could be exploited without depleting the reserves was approximately 15% of precipitation.

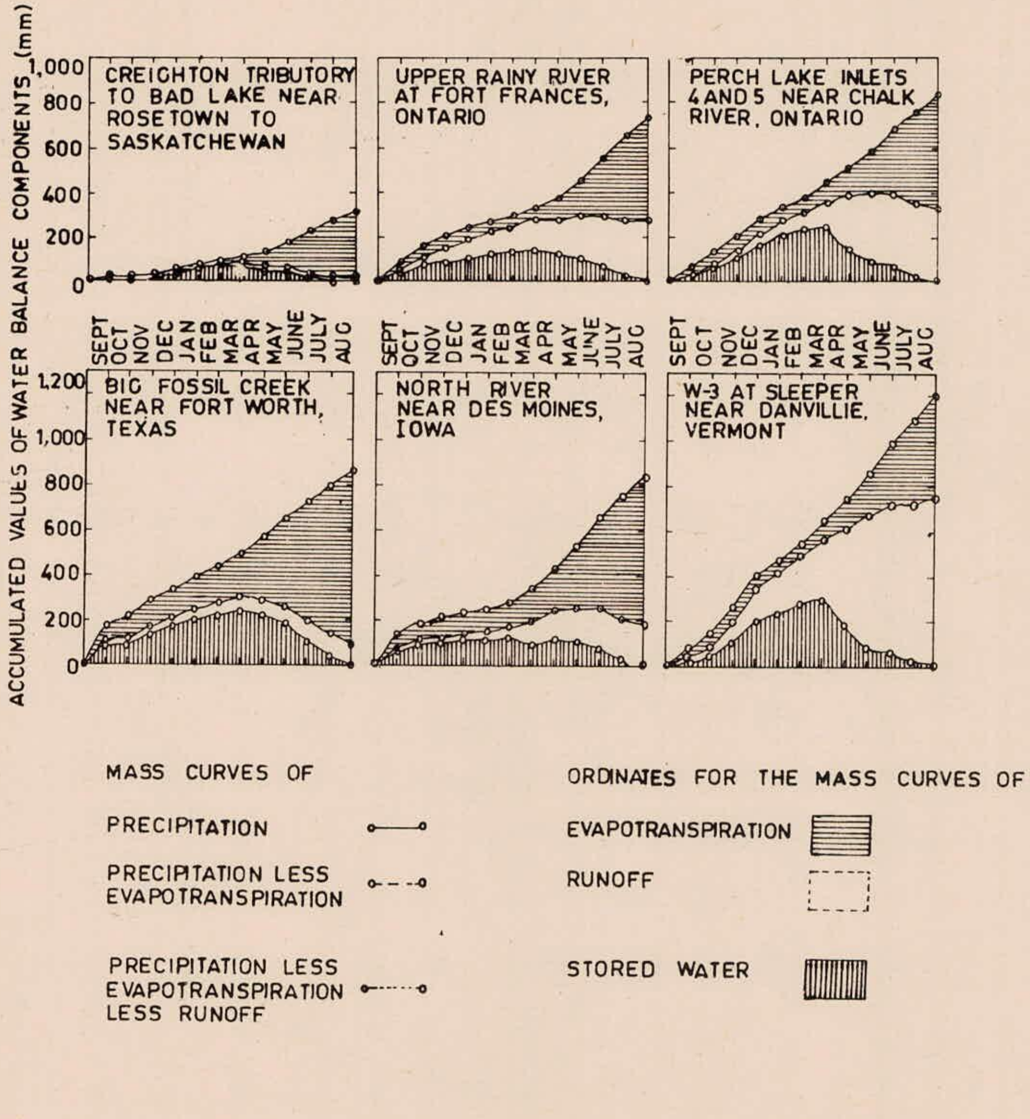

Fig. 2. Seasonal water balance for six basins in North America (Morton, 1983).

Holecek (1982) has reported good results in predicting runoff responses to specific precipitation and snowmelt events for Spring Creek of Canada, using a relationship between effective drainage area and the basin storage. The relationship is highly non-linear and this relates to a major problem that the basin storage includes solid storage in the snow, tension storage in the unsaturated soil and gravity storage in lakes and the saturated ground water zone. Each of these different types of storage has different effects on runoff. Tension storage has an indirect influence on runoff through its effects on infiltration, overland flow and ground water recharge. However changes in tension storage produce no direct or immediate response in the runoff. On the other hand, the gravity storage has a direct effect on the runoff.

Klemes (1983) described a graphical technique for analysing the above effects. Using Turtle river data mass curves of input and output were drawn. The difference in them were the basin storage. Similar analysis was done on Humber river basin Canada by Morton (1983). But the results of Humber basin were complicated because of the presence of snow component. The above studies established the usefulness of the water balance to real time forecasting. Klemes (1983) has provided an example illustrating how easy it is to detect and correct for errors in the input data.

Water balance of a river reach of sufficiently large length needs the approaches used for a basin. This is particularly because of the contribution of the intervening catchment

While considering short reaches, the precipitation, evapotranspiration components may not be of much consequence.

When water balance is carried out for hourly or short duration, the continuity and momentum equations (as per equations, 2, and 3) together provide a solution:

$$\frac{\partial A}{\partial t} + \frac{\partial Q}{\partial x} = q \qquad \ldots(2)$$

$$\frac{\partial Q}{\partial t} + \frac{Q}{A}\frac{\partial Q}{\partial x} + Ag\ (S_o - S_f + S_w) = 0 \qquad \ldots(3)$$

where,

A is area of cross section of flow

Q is discharge m^3/s

S_o is bed slope

S_w is water surface slope

S_f is energy slope

g is acceleration due to gravity

x, t are space and time coordinates

q is lateral flow computed by proper rainfall runoff model

In the case of daily or weekly time period the balance would include all the components identified earlier. For longer time periods it may be possible to assume the changes in storage to be relatively insignificant and the evapotranspiration would be equal to the precipitation less runoff for the intervening catchment.

The following components could be identified for accounting the water in a river reach:

1. Precipitation (rain and snowmelt)

2. Inflow from upstream

3. Storage (in flood plain and channel of intervening catchment)

4. Diversion

5. Outflow

6. Ground water component and bank storage

These components are briefly explained below:

2.1 Precipitation

A part of the precipitation occuring over the stretch of the basin, as shown in Fig.3, joins the river as lateral flow. Lateral inflow would be substantial when the river reach is long and the catchment contributing to the reach is big enough.

The flow may reach the river at tributary junctions as tributary flows or as a flow over land surface which enters the river at numerous points not well defined as tributary channels. The tributary flow component runs almost parellel to the river on the surface of the basin before entering the well defined tributary channel. The other component occurs from plains sloping towards the river reach. This flow occurs at innumerable points that it can be considered as uniformly distributed over the entire stretch. This overland flow is dominant in the case of small basins. In large basins the channel storages suppress the effect of changes in land use etc. (Chow, 1964).

A simple relation between rainfall and the runoff may be obtained to be used for computing this particular component. For example regression analysis of the observed

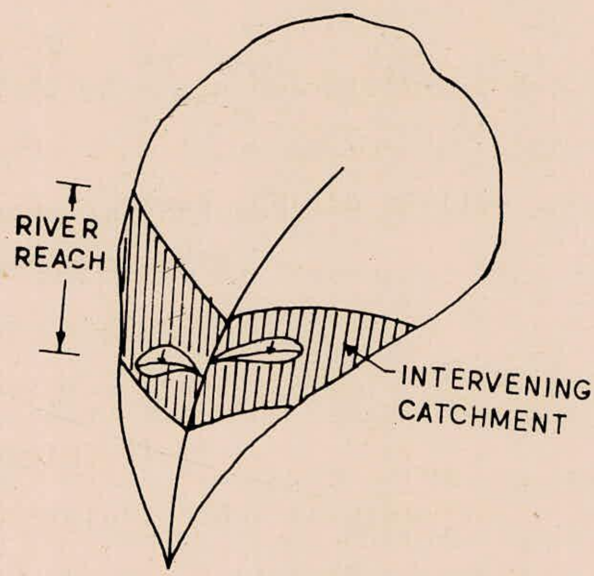

FIG. 3

Fig. 3 : A river reach

rainfall and runoff data could be developed and used.

2.2 Inflow from Upstream

This is obtained by direct measurements when the upstream end of the river reach is gauged. The total water inflow into the river reach can be calculated from the water balance:

$$q_{et} \Delta t = Q_{rn} \Delta t \pm \Delta W \qquad \ldots(4)$$

where, q_{et} is the mean water inflow during Δt into the reach, Q_{rn} is the mean discharge during the time Δt from the reach and Δw is the change in reach storage. The total water inflow into the reach normally fluctuates more widely than the outflow through the downstream end of the reach. The variation of the total water inflow is closely related to the fluctuation in the surface runoff in the upstream catchment. The channel storing capacity therefore does not exert any regulating influence on the magnitude of the inflow. The total water inflow is actually the difference between surface runoff and runoff losses at the upper reaches.

2.3 Channel Storage

As per Apollov et al (1970), the channel storage can be determined from the morphometric characteristics of the channels or from analysis of hydrometric data.

From Morphometric Data:

The drainage area A of a reach bounded by two (upper and lower) gauging sites is determined from large scale maps or from aerial photographs obtained from different water states. Then the relationship $A = f\ (H_u . H_e)$ is derived where H_u and H_e are the water stages at the upper and lower gauging sites.

$$A = f\ (H_u\ ,\ H_e) \qquad \ldots(5)$$

For this purpose a reach can be divided into number of sub reaches. This method is accurate, but its application is limited by insufficient data.

Under natural conditions floods have the character of long waves. Therefore, an approximate correlation exists between water discharges observed simultaneously at different gauging sites. This makes it possible to use the stream cross sectional areas (A) at the different gauging sites as a characteristic for the channel storage. In this case the water storage is calculated from:

$$W_1 = \frac{A_1 + A_2}{2}\ L \qquad \ldots(6)$$

where, A_1, A_2 are the cross-sectional areas of the stream at the gauging sites and L is the length of the reach. The relationship between W_1 and the actual channel storage has the form of $W = a\,W_1 + b$.

From hydrometric Data

The water storage of a channel stretch between two hydrometric sites can be determined from data referring to seperate floods. The water balance equation can be written as:

$$\Delta W = (Q_u + Q_{1t} - Q_e)\,\Delta t \qquad \ldots(7)$$

For the time interval $(t - t_o)$

$$W_t - W_o = \sum_{to}^{t} (Q_u + Q_{1t} + Q_e)\,\Delta t \qquad \ldots(8)$$

where, W_o is the water storage of the channel at the time t_o.

As can be seen from the Fig. 4 the water storage of the channel reach is expressed by abc and a'b'c' respectively from time t_1 and t_2.

With data on the channel flow time 'T' and on the water discharges, the channel storage can be determined from the relationship:

$$\frac{\Delta W}{\Delta Q} = T \qquad ...(9)$$

where, ΔQ is the difference in discharge between the reach. In integral form the storage can be written as:

$$W = \int_0^Q T\, d\, Q \qquad ...(10)$$

For an average flow time 'T' we have the approximate relationship as follows:

$$W = \bar{T}Q \qquad ...(11)$$

Hence the water storage of the reach is equal to the product of water discharge and time of travel.

$$W = \frac{Q_1 + Q_2}{2} T \qquad ...(12)$$

where, Q_1 and Q_2 are discharges observed at upstream and downstream sites respectively. When determining the water storage of a stretch where large tributary flows, the sum of the water discharge of the main river and that of its tributary is considered at the upper gauging site. Somewhat more accurate results of determination of the channel storage may be obtained if the weighted mean is taken instead of the mean water discharge.

Since the flow velocity V_Q of the water discharge is somewhat higher than the stream velocity the relationship (11) gives somewhat low values for the water storage. If T is replaced by L/VQ and $\bar{Q}$ by $\bar{w}\bar{v}$. The equation can be re-written as:

$$W_{cal} = \frac{\bar{V}}{V_Q} L \bar{w} = \frac{\bar{V}}{V_Q} W_{act} \qquad \ldots(12)$$

The quantity $\bar{V} = V_q$ aries from 2/3 for a rectangular to 4/5 for a triangular.

Assuming that the sharpest recession corresponds to periods of draining of the channel storage, when there is no considerable water inflow into the channel, the following method can be adopted for small rivers and upstream reaches of larger rivers.

The amount of water storage of the channel at each moment is calculated from the hydrograph. As shown in the Fig. 5 the water volume determined from the area abc is obviously equal to the water storage W at the time t :

By carrying out such calculations for differennt magnitudes of the water discharge, it is possible to plot the volume (storage) curve W = f(Q). This method, proposed by Bernadskii needs a sharp recession of water discharge and is approximate, since water discharge in a lower and gauging site cannot be taken as characteristic distribution of channel storage in the river system.

2.4 Diversion

Diversion structures are used for diverting the flow from one river to reservoir or another drainage basin. Flow may also be diverted into a channel by pumping where necessary. This may be proportional to irrigation requirement/channel capacity in case of irrigation canals. These quantities can be tabulated for weekly or monthly for use in the

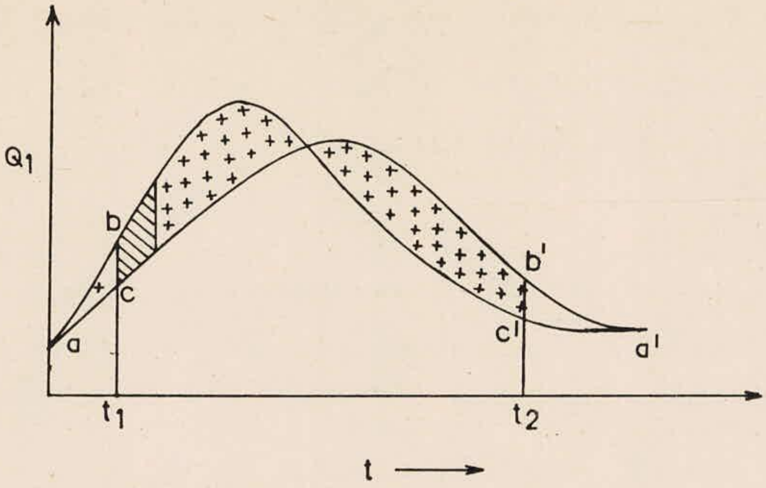

FIG.4.

Fig. 4 : Storage in a channel

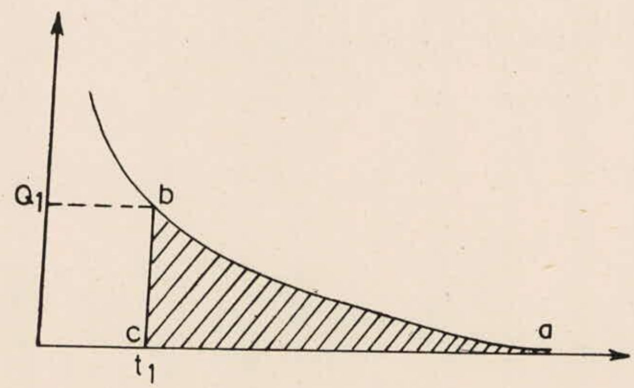

FIG. 5.

Fig.5 : Water storage of a river reach from the recession curve

water balance computation. Diversions as a flood control measures would be generally equal to the channel capacity. In case of diversion for storage purposes will be as per planning. In any case the actual quantity of flow should be considered for the water balance.

2.5 Outflow

The evaluation of this component is usually based on the following water balance equation:

$$Q \Delta t = W + Q_i \Delta t + Q_{ad} \Delta t \qquad \ldots(13)$$

The channel storage 'w' in the above equation can be determined as explained earlier in sec. 2.3. The determination of the second component - the magnitude of sub-surface inflow is more complicated as explained in sec. 2.6. However, assuming that there is a relation between the channel storage and the magnitude of sub-surface inflow it is possible to establish and use the relationship:

$$Q_i \Delta t = a_1 W + C_1 \qquad \ldots(14)$$

for the determination of sub-surface runoff. Further improvement in the calculation of the magnitude of the sub-surface inflow can be achieved by analysis of the fluctuation of ground water storage and by studying dynamics of runoff from hydrographs.

The third component $Q_{ad}\Delta t$ can be determined from data on basin surface runoff. Precipitation that occured in the intervening catchment of the basin as shown in Fig. 3 will enter as Q_{ad}.

Due to the lack of details on evapotranspiration, etc.

this additional inflow can only determined approximately. Assuming the runoff coefficient is a function of the initial moisture content of the basin stretch which in turn is a function of channel storage the following can be written:

$$Q_{ad} = f\ (w)$$

$$= a_2\ w + c_2 \qquad ..(15)$$

Finally, for simplicity the outflow can be written as a function of storage. Computing for large rivers the outflow for several days of month gives completely satisfactory results, except for the periods of most intensive water inflow into the reach as a result of heavy rain or snow melting.

Kesslitz V and Ogievskii, A.V. showed that it was possible to differntiate between the two genetically different parts of summer runoff : (1) Runoff fed from depletion of the channel storage of ground water flow and calculated by means of established recession curves of the water discharge (2) Surface runoff fed from precipitation (Apollov et al, 1970).

2.6 Ground Water Component and Bank Storage

Sub division of total streamflow into the surface and sub-surface components is necessary in investigations concerning water balance of a river reach. Streamflow is known to consist of three components; surface runoff, interflow and ground water discharge. The first two are collectively described as direct runoff. Surface runoff represents that part of precipitation which flows directly over the land surface into stream channels. Interflow is that part which moves laterally through the soil zone and is discharged to streams relatively quickly after infiltration and with out reaching the zone of saturation. The ground water discharge is that part of precipitation which reaches the zone of saturation and contributes temporarily to ground water storage before being discharged from springs and seepages into a stream system. This component may include 'bank storage' which is the water that enters the banks of a stream when the stream stage rises and leaves the bank as stream level falls. The volume of water comprising bank storage depends upon the maximum height of the stream stage during periods of high flow, length of time, the high stage is maintained, and the intrinsic permeability of the deposits forming the stream banks and contiguous areas.

The components of the streamflow have different recession characteristics and the recession of each can be approximated to a straight line when the logarithm of streamflow is plotted against time, the recession equations being of the general form:

$$Q_t = Q_o^{kt} \qquad \ldots(16)$$

where, Q_t is the discharge after time t.

Q_o is initial discharge and K is a negative recession constant (Barnes, 1939). There is usually a time lag between peak discharge of various components. The peak of interflow occurs after surface flow and that of ground water discharges after interflow. Since each component tends to overlap the previous one in time, the recession curve of streamflow as defined by a semi-logarithmic plot of discharge against time may be curvilinear in form.

Where more than one aquifer occurs in a catchment each makes different contribution to streamflow. This tends to mark the basic simplicity of the recession curve. The hydrological properties of the individual aquifers determine the extent of their individual contributions both in time and volume. The percentage of the total flow of a stream derived from ground water discharge depends upon a variety of factors but principally the distribution and intensity of precipitation and the nature of surface deposits within the catchment. A hydrograph plotted logarithmically is only approximately a straight line in the early stages of recession. The overall curve tends to be curvilinear. This may be due to areal variation in intrinsic permeability of the aquifer contributing to the stream, variations in the head of water in the aquifer and losses due to evaporation from the riparian zone.

Kalinin et al (1957), Riggs (1953), and Snyder(1939) have discussed the methods of differentiating the subsurface components of streamflow. Kunkle(1965) showed that differences between the specific electrical conductivity of surface and ground water could be used to seperate the two components In a similar manner any property for example, temperature alkalinity, beta activity, etc., which distinguishes the two components can be used.

In a river basin where only one aquifer contributes to ground water discharge the analysis of the stream hydrograph to assess the component is relatively simple. When different aquifers are contributing, the study becomes complex to evaluate the component. Theoretical basis for the method has been described by Kudelin (1960). Three categories can be identified for the purpose of stream-aquifer relations. They are, (i) river reach having no hydraulic connection with an aquifer, (ii) reach having constant hydraulic relationship with aquifer, (iii) reach having periodic or intermittent hydraulic relations with aquifer.

When the stream and aquifer are not hydraulically connected, the ground water discharge to stream will be very similar to surface runoff except that the peak of ground water discharge is less pronounced and lag behind the peak of the stream as shown in Fig. 6. An increment of ground water discharge is added to the stream in the flood period.

The peak of ground water discharge corresponds to that of minimum surface runoff, for the case where stream and aquifer are hydraulically connected. During a flood when the stream rises above the ground water level, surface water enters into bank storage. As the stream stage falls below this level the water returns back as shown in Fig.7.

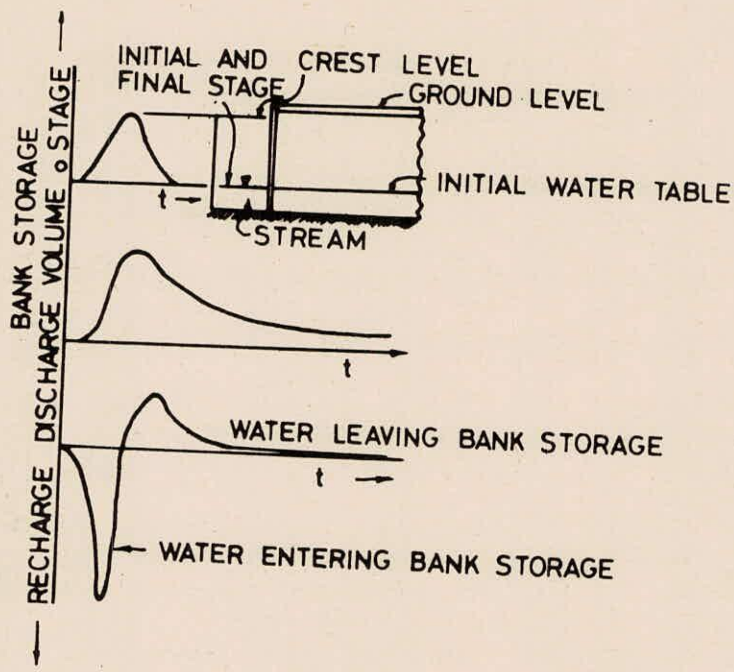

Fig. 6 : Flow to and from banks

The net contribution to stream flow from an aquifer connected hydraulically with a stream is relatively small during a flood period.

When the stream has intermittent connection with aquifer, it is classifed as 'mixed'. At low stream stages the stream is not hydraulically connected. The size of the drainage basin and the areal distribution of ground water discharge must be considered in the analysis of a stream hydrograph. In comparatively long river reach the movement of a flood wave down the river may affect the volume of ground water discharge to differing degrees in different reaches of the river (Unconfined) Ground water discharges may not have appeared in the stream flow measured at a gauging station at the downstream site although ground water discharge may have occurred already in the upper reach of the basin and may be in transit down the stream. In this event the analysis of the stream hydrograph requires data relating to the beginning and the end of the flood at the upstream site and the travel time of the same to reach the downstream site.

Makarenko (1948) has devised a method for computing the annual ground water discharge from aquifers not hydraulically connected to stream. An estimation of ground water discharge can be made by establishing a relationship between ground water levels in an aquifer and group water discharge to a stream (Rasnussen et al, 1959, and Schicht, et al, 1961).

The coefficient of baseflow or ground water discharge can be obtained without specific separation of the stream hydrographs, if it is assumed that the minimum mean monthly flow of the stream is entirely ground water discharge.

An excellent review by Hall (1968) traces the development of mathematical modelling of streamflow recession curves and provides an extensive bibliography on the subject.

Bank Storage

During a flood period ground water levels may be temporarily raised near a channel by inflow from the stream. This water is known as bank storage (Chow, 1964, p-13-34). Todd (1955) indicated the volume and flows to and from bank storage as function of time as shown in Fig. 6. He determined this using a laboratory model.

The flow can be studied using one dimensional, unsteady flow equation as given below (McWhorter et al, 1977):

$$\alpha \frac{\partial^2 s}{\partial x^2} = \frac{\partial s}{\partial t} \quad \text{...(17)}$$

where,

s is draw down

α is parameter known as hydraulic diffusivity (equal T/s where T-Transmissivity)

s is storage coefficient or specific yield).

The flow can be computed by solving the above equation subjected to the following conditions. Initial and boundary conditions (referring to 7):

$$S(x, o) = 0 \quad \text{...(18)}$$

$$S(\alpha, t) = 0 \quad \text{...(19)}$$

$$S(o, t) = 0 \quad \text{...(20)}$$

The partial differential equation can be transferred to ordinary differential equation by introducing a new variable:

$$y = \frac{x}{\sqrt{4 \alpha t}} \quad \text{...(21)}$$

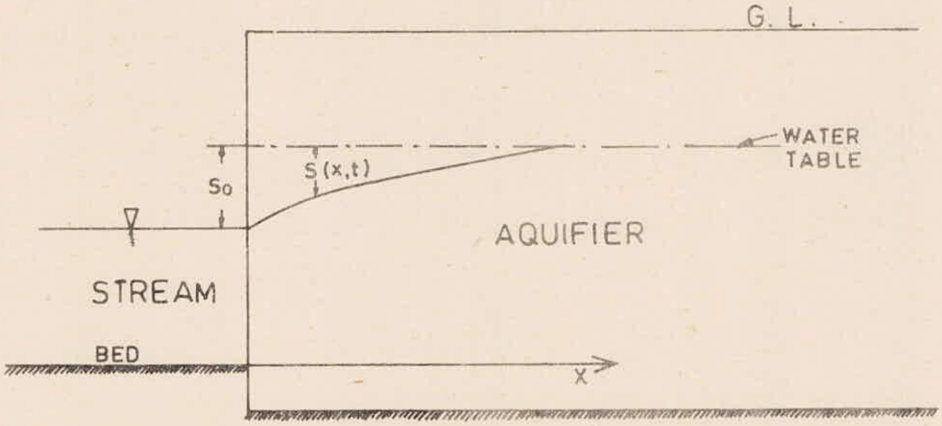

FIG.7 FLOW FROM BANK STORAGE

The solution finally turns out to be :

$$S = S_o \left(1 - \text{erf} \left(\frac{x}{4\alpha t} \right) \right) \qquad \ldots(22)$$

This solution is true for an aquifer that extends to infinity but also provides a good approximation for finite aquifer in the vicinity of the stream until the effective extent of influence reaches the aquifer boundary.

The discharge from the aquifer to the stream is :

$$Q = \frac{S_o \;\; T}{\sqrt{\pi\alpha} \;\; t} \qquad \ldots(23)$$

The discharge must be multiplied by two to include both the sides of the stream.

Hall et al (1972) regarded the quantity (1-erf x/ $4\alpha t$) as a response function of the aquifer to a unit step change of stage.

3.0 CONCLUDING REMARKS

A river reach bounded by drainage divided from a logical unit for hydrologic studies. Within this framework one can conveniently draw up water balance and assess water resources, generate hydrological information to manage the water resource. A river reach is a proper unit for studies of relationships between input, storages, transfers and output, leadiong to knowledge as to when and in what ways it is advantageous to intervene locally in the natural balance.

The following components were identified:

a) Precipitation

b) Inflow from the upstream

c) Storage

d) Diversion

e) Out flow

f) Ground water component and bank storage

The precipitation occurring in the intervening catchment contributes to the flow. The resulting flows may join the river as tributatry flow or as overland flow distributed over a length. Conversion of the precipitation value (daily, 10 daily monthly or annual) need to be converted to runoff. A relationship may have to be established for each one of the time period and used.

The inflow is usually gauged. Current meters are employed to measure the discharges. In some cases stages are measured on a regular basis and discharges are computed using established rating curves.

The storage can be calculated assuming a linear variation of water level and cross-sectional area. Use of toposheets for determining the variation in width of river may be made. The storage in the river channel and plains may be important only when daily, weekly or ten daily durations are used. Since for the other longer durations the excessive storage on plains would have drained and only small storage component would be present.

Diversions, if any, are to be obtained from records. The outflow is also a measured quantity. In case if there is no gauging site the quantity can be estimated balancing the other 5 components.

Groundwater component is a most important aspect to be considered. Estimation of the same is quite complex. Certain methods have been discussed in the text. The bank storage component will come into picture when water level in the channel is raising or falling during floods. The component need to be accounted for short duration balance and for alluvial reaches.

Major difficulties involved in evaluating the water balance component are : (i) transmissivity, (ii) storage coefficient of aquifer, and (iii) cross sectional details of the reach at close intervals. It can be noted that these are field data which are difficult to obtain. For example, the transmissivity and storage coefficient vary largely along the reach and they need to be obtained for short interval of length. Getting cross-section survey on the river for necessary number of cross-section could be very costly and time consuming. Possible use of remote sensing data would reduce the cost of acquiring the necessary details. Further, a long range of data on precipitation and runoff would suffice the establishment of the balances.

However, it needs to be studied (i) what sort of relationship is needed between rainfall and runoff when different durations are used; (ii) what kind of averaging procedures can be used for determining ground water components and the sensitivity of these parameters for the required water balance component with particular reference to different regions.

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