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STUDY OF SPRINGS AND HYDROLOGIC MODELLING OF SPRING FLOW

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1988-89

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ABSTRACT

Springs are outlet through which the groundwater emerges at the ground surface as concentrated discharge from an aquifer. Springs are part of the groundwater system can be taken as a flowing well with constant head. They could occur in various sizes from small trickle to large streams. Conditions required for having a spring are manifold and these are various combinations of geologic, hydrolgic, hydraulic, pedologic, climatic and biologic controls. A few spring may indicate the existence of thick transmissive aquifers whereas frequent small springs tend to indicate thin aquifers of low transmissivity. Springs not only aid in the evaluation of groundwater potential of the area, it can be used to meet and supplement the different requirements for water of the area. There are various types of spring flow domain depending on aquifer geometry and other physical factors. Discharge rate from a spring depends on the size of the recharge area above it, the rate of precipitation in the area, aquifer geometry, geology and geomorphology of the area, storage coefficient and transmissivity of the aquifer.

There are many springs in the Himalayas, Western Ghats and other places in the country. But, yet there is no systematic study of the spring flow for harnessing them as a dependable source of water. There is enough scope of research in this regard particularly in respect of the mathematical modelling of spring flow.

Jacob Bear (1979) suggested a simple mathematical model to analyze the unsteady flow of a spring with steady state recharge. In the present study, the model suggested by Bear has been improved upon to acccount for time variant recharges that contribute the spring flow with

the help of discrete kernel approach. Variations of spring flow discharge with time in response to variable recharge input for different aquifer parameters have been presented. Another model of spring flow visualising the flow domain as pipe flow as encountered in fractured rocks has been developed.

1.0 INTRODUCTION

Springs are natural outlets through which the groundwater emerges at the ground surface as concentrated discharge from an aquifer and are most conspicuous forms of natural return of groundwater to the surface. Springs are part of the groundwater system and may be treated as a flowing well with constant head. Springs could occur in various sizes from small trickle to large stream. Seepage areas are distinguished from springs because of the slower movement of groundwater from the seepage area unlike the springs. Water in seepage areas may pond and evaporate or flow according to magnitude of the seepage, climate and topography. Likewise excavation, drillings, etc. (wells, boreholes) from which groundwater is drawn, come into the category of artificial discharge centres whereas a spring is the natural groundwater exit.

A few large spring may indicate the existence of thick transmissive aquifers whereas frequent small springs tend to indicate thin aquifers of low transmissivity. So, the springs can aid in the evaluation of groundwater potential of the area. Conditions necessary to produce springs are many and are related to different combination of geologic, hydrologic, hydraulic, pedologic, climatic and even biological controls. Therefore, there are many descriptive terms for springs based on a single or combined controlling factors. But, the most common controlling factors that are used in order to classify springs for different studies are:

1. Character of opening from which water emerges,
2. Character of water bearing formations,
3. The structure and resulting force that brings the water to surface,
4. Quantity of water discharged,
5. Uniformity and periodicity of the rate of discharge,

6. Chemical quality of water discharged,

7. Temperature of water,

8. Deposits and other features produced by springs,

9. Source of water shallow or deep seated, and

10. Direction of movement of water.

Discharge rate from a spring depends on the size of the recharge area above it, the rate of precipitation in the area, aquifer geometry, geology and gemorphology of the area, storage coefficient and transmissivity of the aquifer.

The water from springs are being used from ancient times. Roman empire was supplied with spring water through elaborate aqueducts. In the eastern Mediterranean area, spring water was often used to drive small water turbines before being diverted for irrigation and domestic purposes. In the highlands of Persia insignificant seepages were developed into large artificial springs by excavating long subterranean galleries the so called ghanats. Similar methods of diverting springs flow were universally employed until the present century (Mandel and Shiftan,1981). Now a days, it is more convenient or efficient to divert the water directly through wells from the aquifer feeding the spring. Bear (1979) suggested that by an appropriate groundwater management policy, the water levels in the aquifer in the vicinity of a spring can be maintianed below spring outlet. The water previously emerging from the spring thus be stored in the aquifer and used according to development programme.

Various hydrological aspects of springs and their flows need to be studied in order to harness them appropriately to meet and supplement different water requirement of the nearby area and for better aquifer management.

2.0 FACTORS AFFECTING SPRING FLOW

2.1 Classification of Springs

There are various types of spring flow domain depending on aquifer geometry and other physical factors. Discharge rate from a spring depends on the rate of rainfall/snowfall or other water accretion in the area, the size of the recharge area above it, the geology and geomorphology of the area, aquifer geometry storage coefficient and transmissivity of the aquifer. Spring can occur both under water table and artesian conditions. So it is quite natural that there are a large number of classifications of springs. One of the earliest is the classification of K.Keilhack (1912). The classification was modified slightly by him in 1935 and is given in Table 1. This classification is based on descending (water table) and ascending water (artesian) flow in the spring. The points at which groundwater (descending) and artesian water (rising) emerge can be easily distingushed. Stratal, fissure-vein, and karst water are distinguished according to the nature of the exit and the composition of the rocks containing it.

Table 1. Classification of springs (after Keilhack, 1935)

Decending	Ascending
1. Spring caused by aquifer constriction or	1. Spring caused by hydrostatic pressure a) contact spring b) fissure spring
2. the ending thereof	2. Spring aided by a) steam b) carbon dioxide c) hydrocarbon, methane eruption
3. contact spring	
4. overflowing spring	
5. impounded spring	
6. karstic and fissure spring	
7. fault spring	

Descending type of springs are usually periodically flowing. An example of spring flow from an overflowing groundwater reservoir is given in Fig.1(d).

Discharging points of ascending type of springs are more varied. They form erosion barrier and structural springs (Ovchinnikov, 1968). However, Keilhack's classification is based on the aquifer geometry only.

2.1.1 Classification of springs on the type of water bearing formation

Tolman (1937) classified springs on the basis of type of water bearing formation and the type of opening. He suggested six categories of springs and each one has its own discharge characteristics. These are described briefly herein.

1. Spring issuing from permeable veener formations overlying relatively impermeable bedrock with irregular surface:

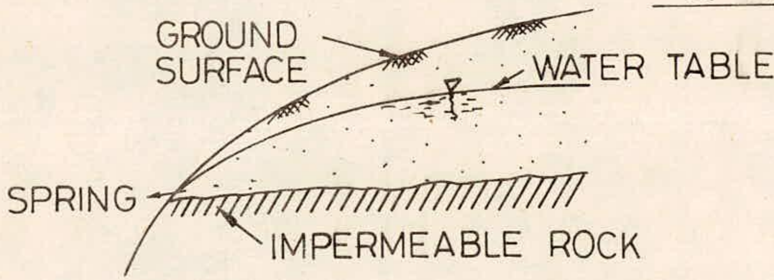
Outcrop of bedrock or nearness of bedrock to surface is the controlling factor. The discharge from such springs are usually small, but some are large. Discharge of spring could vary considerably periodically. Various types of spring that could emerge due to this include contact, gravity, perched, talus and barrier springs (Fig.1a,b,c,d,e).

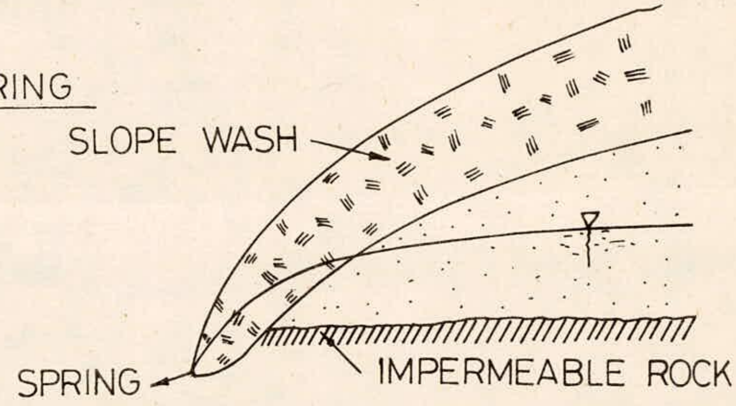
2. Springs issuing from thick permeable formations:

Underlying bedrock is quite deep to have any control over the spring flow. Intersection of sloping water table with land surface is the controlling feature in this type of springs. Discharge of such springs are usually small. All water table springs including channel, valley, alluvial-slope and depression springs come under this category(Fig.2a,b,c).

3. Springs issuing from interstratified permeable and impermeable formations with or without structural deformation of the aquifer:

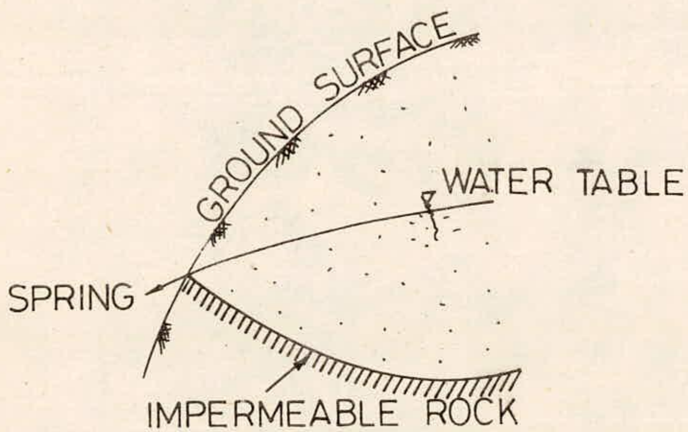
a). GRAVITY SPRING

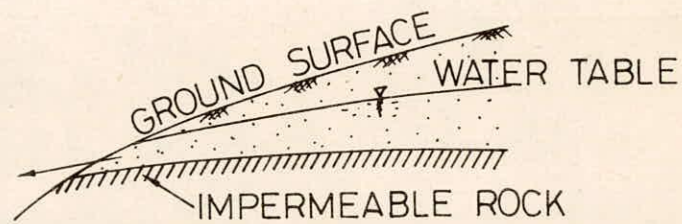




GRAVITY SPRING IN SLOPE WASH

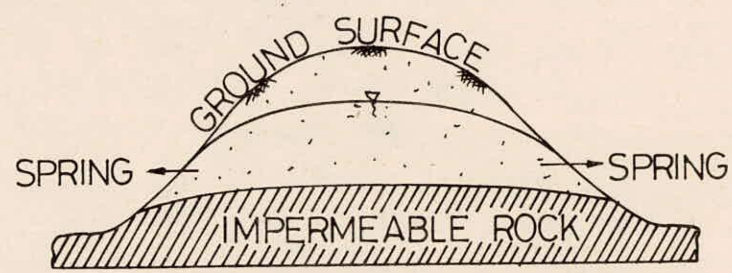
b). PERCHED SPRING



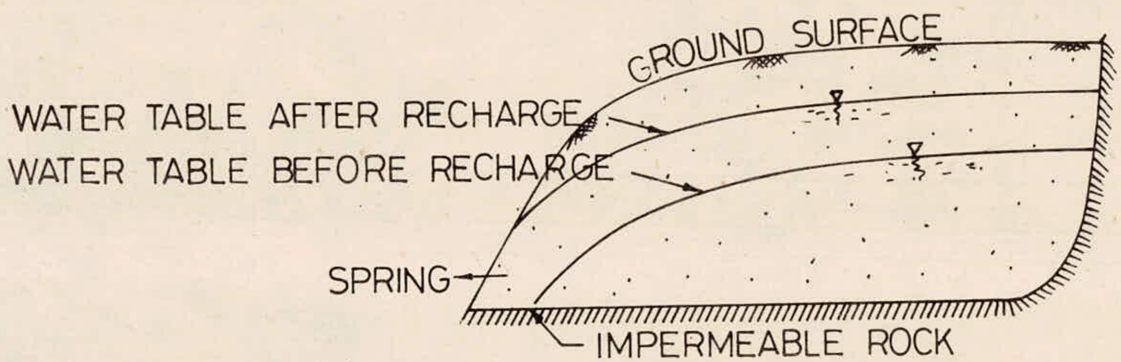


Contd.,

c). CONTACT SPRING



d). OVER FLOWING OR OVER SPILLING SPRING



Contd.,

e). BARRIER SPRINGS

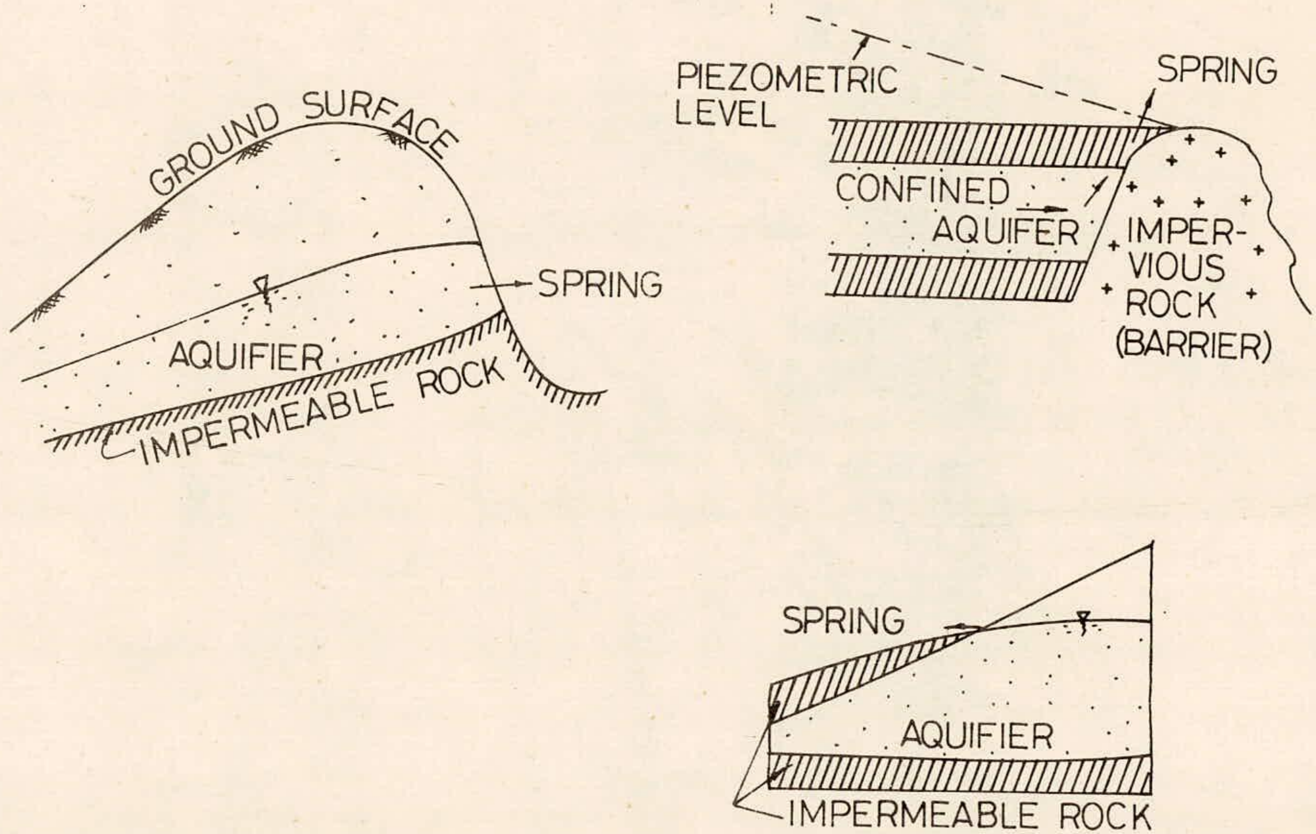


Fig.1 Various types of springs issuing from veener formations overlying relatively impermeable bedrock

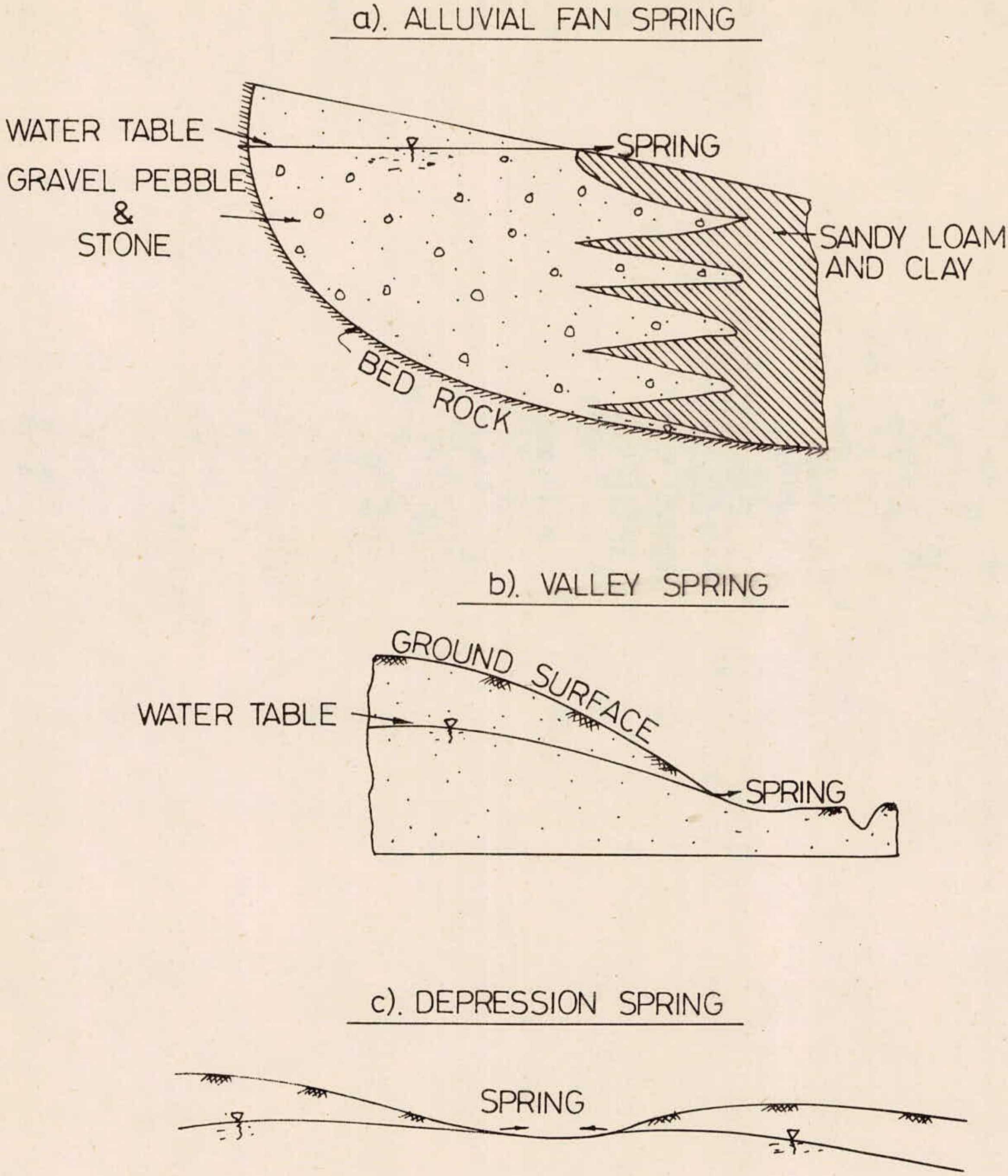


Fig. 2 Springs issuing from thick permeable formation

Springs under this category could draw on confined water (artesian springs) or water may be unconfined. Outcrop of the aquifer controls the springs. Discharge of springs usually small, but some may be large. This category includes contact, monolinal, synclinal, anticlinal and unconformity springs (Fig.3a,b,c).

4. Springs issuing from solution openings formed primarily along fractures and bedding planes in carbonate rocks:

Carbonate rock springs generally represent constricted discharges, with abundant flows at widely separated outlets. The real contributing factor is the presence of numerous and wide cavities through which water circulates. The cavities are genetically developed by water circulation and are formed by the attack of dissolved carbonic acid on the limestone. Discharge of these springs is often large and fluctuate considerably. A spring under this category may either be unconfined groundwater spring or confined groundwater spring depending on the aquifer geometry. A spring formed due to process of dissolving limestone and emerges above or below the sea may be referred as coastal spring, but it can be categorised under this head (Fig.4a,b,c,d). So, it may be noted that a particular spring may be categorised in a number of ways according to different controlling factors.

5. Springs issuing from factures and tubes in lava and from thin interbedded porous strata:

Discharge of such springs is usually large and steady as they are sustained by a large body of groundwater because of numerous narrow openings (Fig.5).

6. Spring issuing from fractures intersecting permeable materials and impermeable materials and fractures supplied in part by waters of deep-seated unknown origin:

Usually the discharge of such springs is small, but it could be large one (Fig.6).

2.1.2 Classification of springs on the basis of magnitude of discharge:

Meinzer (1923) classified the springs according to the magnitude of discharge. This has been given in Table 2. The discharge rate of a spring depends on the size of the recharge area above it, the rate of precipitation and other water accretion, geology and geomorphology of the area, aquifer geometry and aquifer parameters. Meinzer (1927) stated that large magnitude springs occur primarily in volcanic and limestone terranes. Todd (1980) reported the relation of catchment area and annual recharge to average spring discharge. This depicted in Fig.7.

A spring with a few hectares of recharge area can supply the needs of a single family whereas large recharge area with high rainfall is necessary to produce a first magnitude spring. For example, the largest karst spring in USSR the Krasnyi Klyuch spring occurs at the Ufa plateau in the Ufa river valley (Western Ural area). The discharge of this spring time. The later yield could meet the total water demand of Moscow.

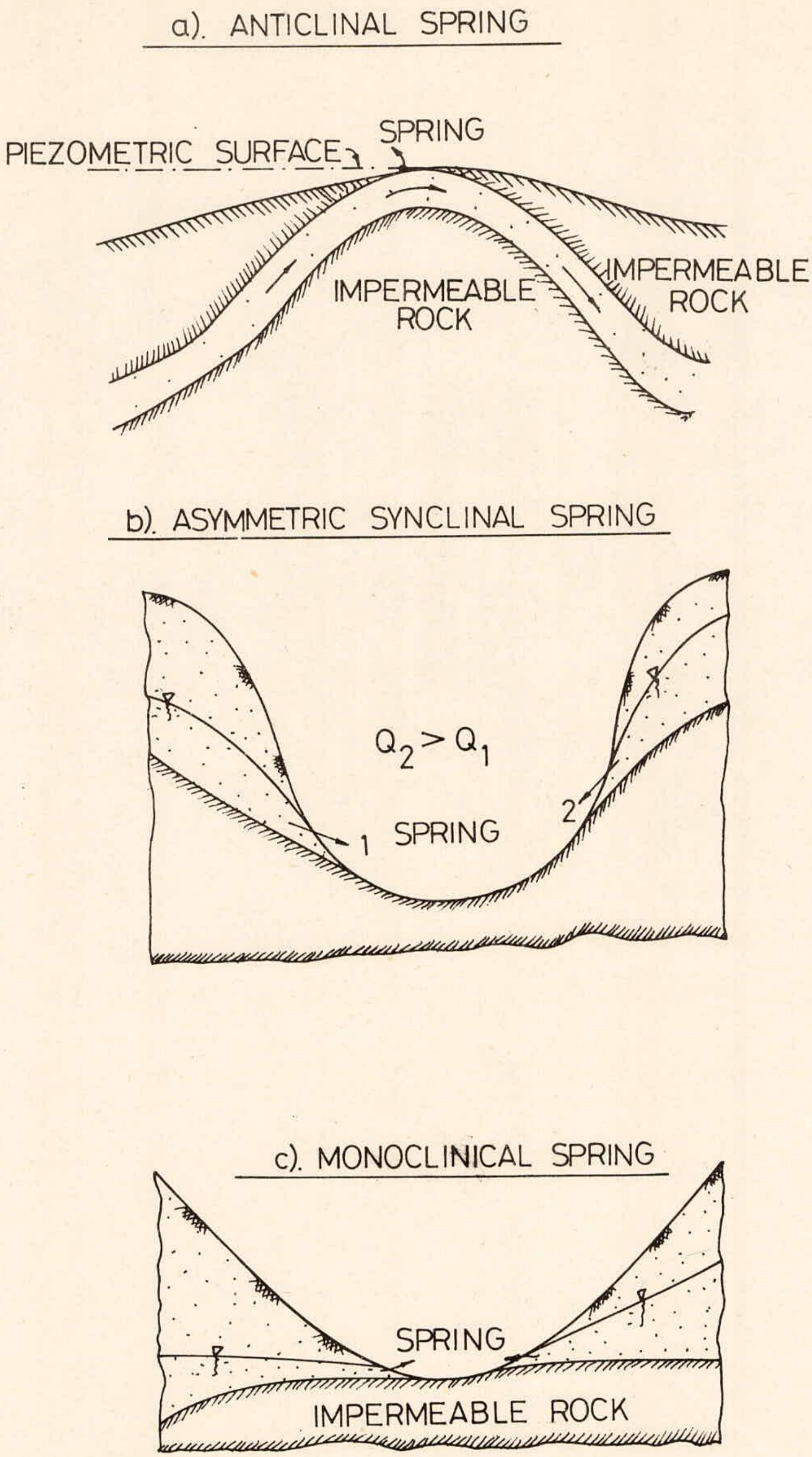
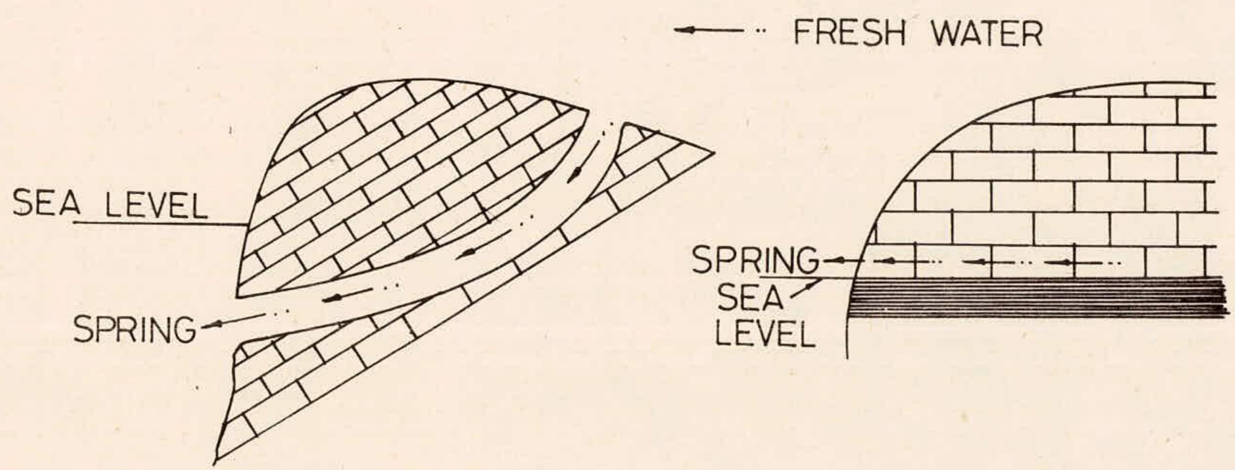


Fig. 3 Springs issuing from interstratified permeable and impermeable formations

a). COASTAL SPRINGS



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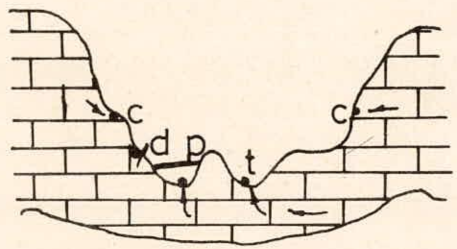
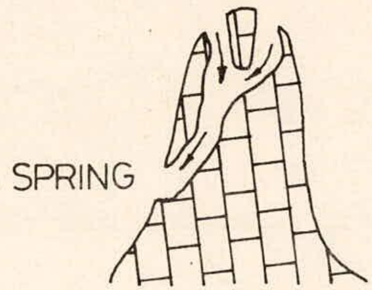
b). EROSION SPRINGS IN CARBONATE ROCK ALONG BEDDING PLANE



c - SPRINGS ON SLOPES, d - ON DOLINE, t - ON A TERRACE, p - IN A RIVER

c). SLOPE SPRING IN CARBONATE ROCK



d). ARTESIAN SPRING IN CARBONATE ROCK

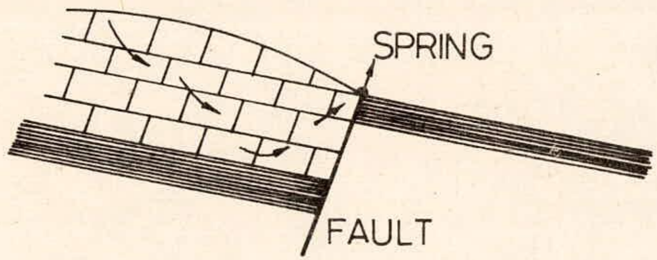


Fig. 4 Springs issuing from solution opening

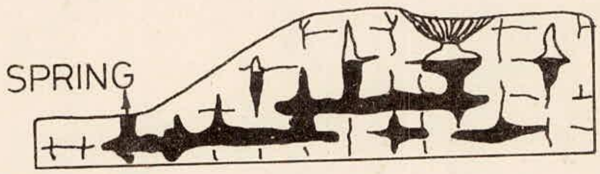


Fig. 5. FRACTURE AND TUBULAR SPRINGS

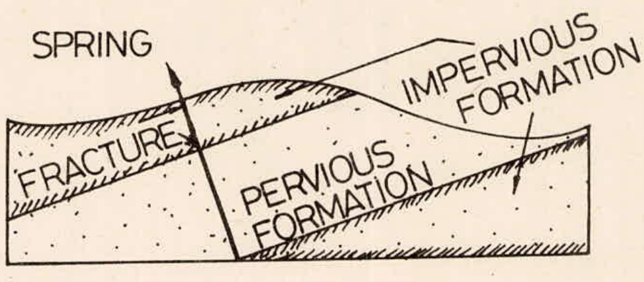


Fig.6. SPRING EMERGES FROM FRACTURE

Table 2 Classification of springs according to magnitude of discharge (after Meinzer, 1923)

Magnitude	Discharge: Cu ft/sec & gpm	Discharge: m^3/day
First	100 cfs or more	0.245×10^6
Second	10 to 100 cfs	0.0245×10^6 to 0.245×10^6
Third	1 to 10 cfs	2450 to 24500
Fourth	100 gpm to 1 cfs (450 gpm)	545 to 2450
Fifth	10 to 100 gpm	54 to 545
Sixth	1 to 10 gpm	5.4 to 54
Seventh	1 pt. to 1 gpm	1.3 to 5.4
Eighth	Less than 1 pt./min.	less than 1.3

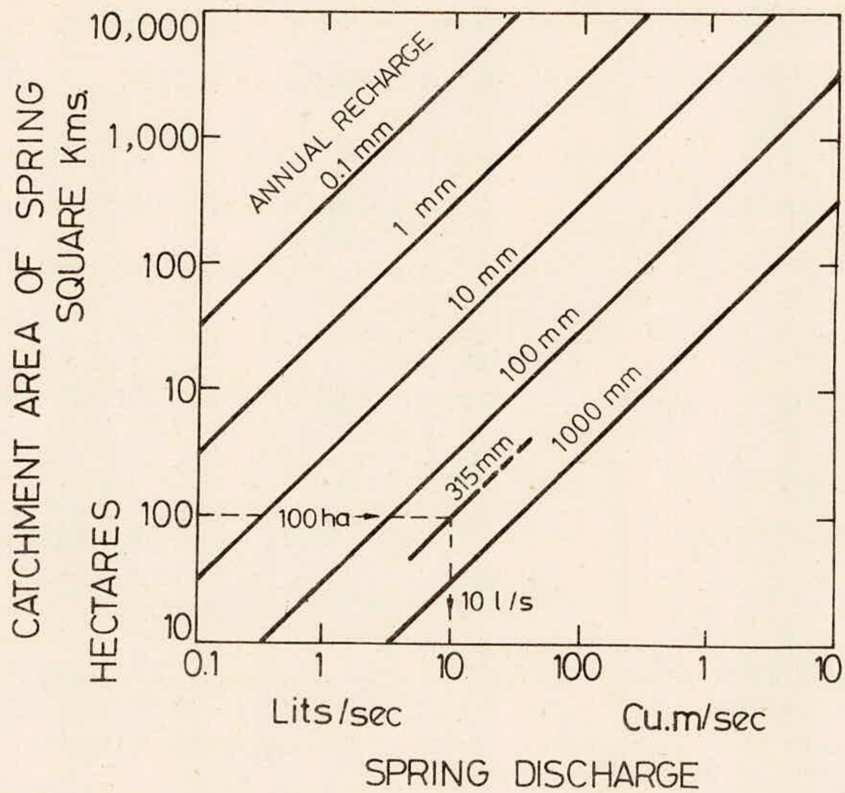


Fig.7 Relation of catchment area and annual recharge to average spring discharge (after Todd, 1980)

2.1.3 Classification of springs according to variability of perennial springs and permanence of discharge (after Meinzer, 1923):

Meinzer (1923) suggested the classification of perennial springs according to variability of discharge. He defined variability as the ratio of the discharge fluctuation to its average discharge within a given period of record.

Thus, $V = \frac{(a-b)}{c} \times 100$

where, v = variability in per cent,

a = the maximum discharge,

b = the minimum discharge, and

c = the average discharge

The classification of perennial springs on this basis are:

a) Constant - springs with a variability of more than 25 per cent

b) Subvariable - springs with a variability of more than 25 per cent but not more than 100 per cent

c) Variable - springs with a variability of more than 100 per cent

2.1.4 Classification of springs according to their exits into another water body:

The natural discharge centres of the groundwater exits in the form of spring (freshwater, carbonated, saline, thermal etc.) pour out at the bottom of rivers, seas or overflow into water bearing layers of higher elevation. Springs, as such, could be classified according to their hidden centres:- 1) sub fluvial 2) sub marine 3) sub surface.

2.1.5 Springs in permafrost region and region of volcanic activity

The condition of spring discharge in permafrost area and those in regions of contemporary volcanic activity are regionally specific.

As a result of freezing of water bearing layers in the active zone above the permafrost in winter, many springs functions only seasonally. Discharging centres of unfrozen water below the permafrost are in active round the year. These form icings round the discharge point.

3.0 CHARACTERISTICS OF SPRING FLOW

Fluctuations of spring discharge are in response to variations in rate of recharge and geologic and hydrologic conditions. Perennial springs drain extensive permeable aquifers and discharge throughout the year, whereas intermittent springs discharge only portion of the year when sufficient groundwater is recharged to maintain flow.

The instantaneous rate of discharge of a spring depends on the difference between the elevations of the water table (or piezometric head) in the aquifer in the vicinity of the spring, and the elevation of spring outlet (called as threshold). During dry season, the spring discharge is derived from water stored in the aquifer. Consequently, the water levels in the aquifer will gradually fall and spring discharge decline. During the precipitation period, the aquifer gets recharged and water table rises and spring discharge increases. This fall and rise of water table in the aquifer go on in a cyclic pattern. The relationship between the rate of decrease and increase of discharge and time (dry and wet seasons) depends on storage characteristics of the aquifer (storativity) and geometry of aquifer (areal extent). Fig.8 shows a typical portion of a spring hydrogram, the recession (or depletion) portion of this hydrogram, corresponding to dry seasons. On a semilog paper (with time on the linear scale), the recession curve usually plots as a straight line.

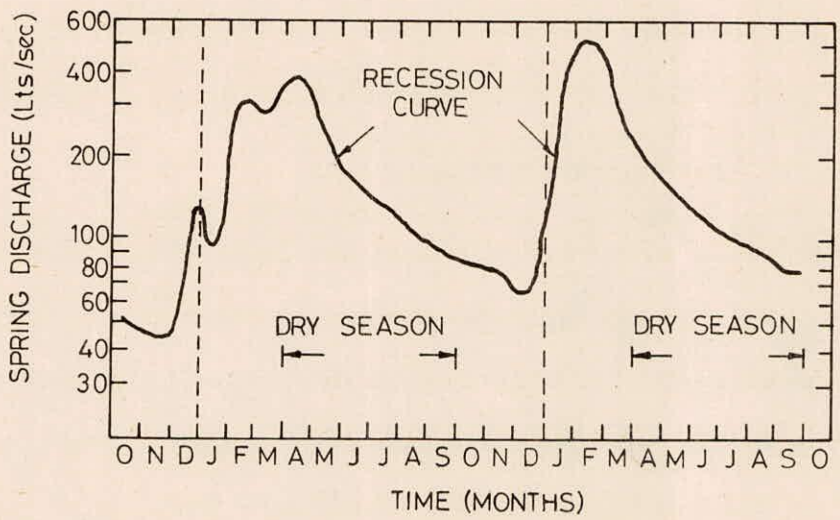


Fig.8 A typical spring hydrogram with seasonal fluctuation (after Bear, 1979)

For the groundwater flow domain analysis on a regional basis, the spring serves as a boundary condition. The elevation of the threshold or outlet of the spring is condidered as fixed head. But, when the water table in the vicinity of the spring drops below the spring outlet at some point of time (which is a priori unknwon), the spring does not act as a boundary of fthe flow domain.

3.1 Existing Models Characterizing the Spring Flow Variability

3.1.1 Model suggested by Mero

Mero (1963) suggested an analytical procedure to find out a solution of problem of forecasting the discharge of the spring during summer months when aquifers are exploited by wells. He attempted to show the hydrological relationship between spring discharge and precipitation and spring discharge and pumpage from the same aquifer. He started from the Darcy's law for the motion of groundwater under dynamic equilibrium.

$$q = -kH \frac{dh}{dx} \tag{1}$$

where,

q = flow of water per unit width of aquifer,

k = hydrualic conductivity of the water bearing formation,

H = total depth of flow $(d+h)$,

h = driving head i.e. head of column of water above line of discharge of zero level, and

dh/dx = gradient of the free water surface, or its equivalent, at a distance x from the point of discharge.

By applying the principle of conservation it follows that the outflow from an unrestricted body of water bearing material equals the amount of inflow plus or minus change of storage. By assuming uniform storativity S and neglecting vertical and cross components of flow, the discharge in case of depleting storage is

$$-\frac{dq}{dx} = S\frac{dh}{dt} \tag{2}$$

By elimination of q between eq.(1) and (2), a differential equation known as the Dupit-Forchheimer partial differential equation, describing unidirectional groundwater flow:

$$\frac{\partial}{\partial x}\left(kH\frac{\partial h}{\partial x}\right) = S\frac{\partial h}{\partial t} \tag{3}$$

This is a nonlinear equation Jacob proposed a particular solution for cases, where the driving head (h) is small as compared with the total depth of flow $H \approx d$. Thus by taking $kH=T$ (termed as transmissivity of the aquifer) as a constant, the eq.(3) could be rewritten as:

$$\frac{\partial^2 h}{\partial x^2} = \frac{S}{T}\frac{\partial h}{\partial t}$$

Now, visualising the flow domain near springs as taken from a finite width and finite depth aquifer as shown in Fig.9, the boundary conditions are $h=0$ at $x=0$ and $x=2L$ for all values of t and $h=h(x)$ at $t=0$. Jacob found the particular solution,

$$h(x,t)= \frac{(WL^2)}{2T}\frac{32}{\pi^3}\sum_{n=1}^{\infty}\frac{1}{n^3}\sin\left(\frac{n\pi x}{2L}\right)\exp\left(-\frac{n^2\pi^2 Tt}{4SL^2}\right) \qquad (5)$$

where W is the average uniform rate of recharge.

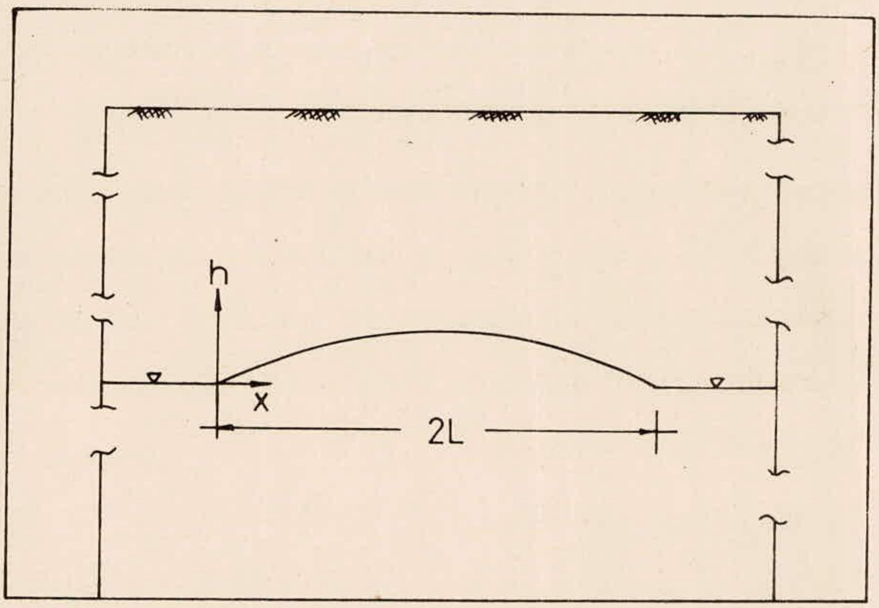


Fig.9 Flow domain in Mero's model

Thus, by substituting (5) in eq.(1) and differentiating with respect to x, an expression for declining flow in the aquifer is obtained.

$$q = -(WL)\ \frac{8}{\pi^2} \sum_{n=1}^{\infty} \frac{1}{2} \cos \frac{n\pi x}{2L} \exp(-\frac{n^2 \pi^2 Tt}{4SL^2}) \qquad (6)$$

As the series converges rapidly, the terms with higher values of n may be neglected. Further, by shifting the point of outflow to x=0, eq.6 may be restated as

$$q(0,t)=q_0 \exp(-\frac{t}{t_0}) \qquad (7)$$

after substituting $-(WL)\ \frac{8}{\pi^2} = q(0,0)$

or the initial flow rate: $\frac{\pi^2 T}{4SL^2} = \frac{1}{t_0}$

where t_0 has the dimension of time, and characterises the depletion of the aquifer. Its physical meaning is the amount of time required to empty the reservoir at the constant initial rate q_0 assuming no replenishment takes place.

The simple equation (7) represents the depletion of an aquifer for non recharge periods (dry weather flow) of spring flow. The constants Q_0, and t_0 can be easily found by plotting observed dry weather discharge (Q) values vs time (t) on a semi logarithm paper, the values of Q being plotted on the log scale.

This model was applied to the Rosh Ha'ayin springs which are the second largest perennial (karstic spring) surface water resources in Israel. These springs are only base flow source of the Yarkon river.

3.1.2 Model suggested by Bear

Bear (1979) suggested a simple model to analyse unsteady flow of a spring (Fig.10).

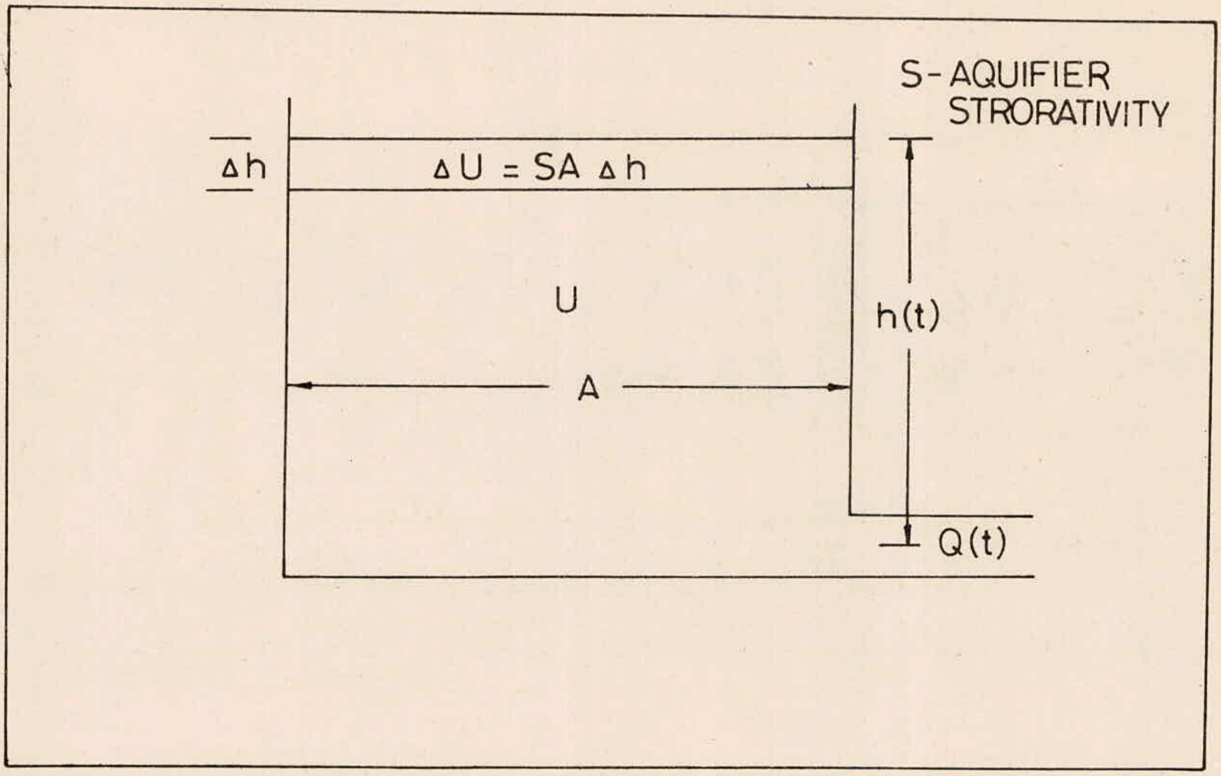


Fig.10 A simple model of a spring(after Bear,1979)

Assuming that the spring drains an unconfined aquifer with discharge $Q = \alpha h$, α = a constant, the discharge from the spring during the recession period (dry period mentioned in the spring hydrogram in Fig.8).

$$Q(t)dt = \alpha h \ dt = -SA\ dh \tag{8}$$

Bear found the solution of the equation (1) with $t = t_o$, $h = h_o$ and $Q = Q_o = \alpha h_o$ as

$$t - t_o(SA/\alpha)\ln(h_o/h) = (SA/\alpha)\ln(Q_o/Q)$$

$$\text{or } Q(t) = Q_o \exp[-\frac{\alpha}{SA}(t-t_o)] \tag{9}$$

The solution satisfies the initial condition $Q = Q_o$ at $t = t_o$. The variation of Q(t) with t will plot as a straight line on a semilog paper (Q on logarithmic scale).

Bear suggested another simple model of a spring draining an unconfined aquifer (Fig.11) with a view to giving an interpretation of α .

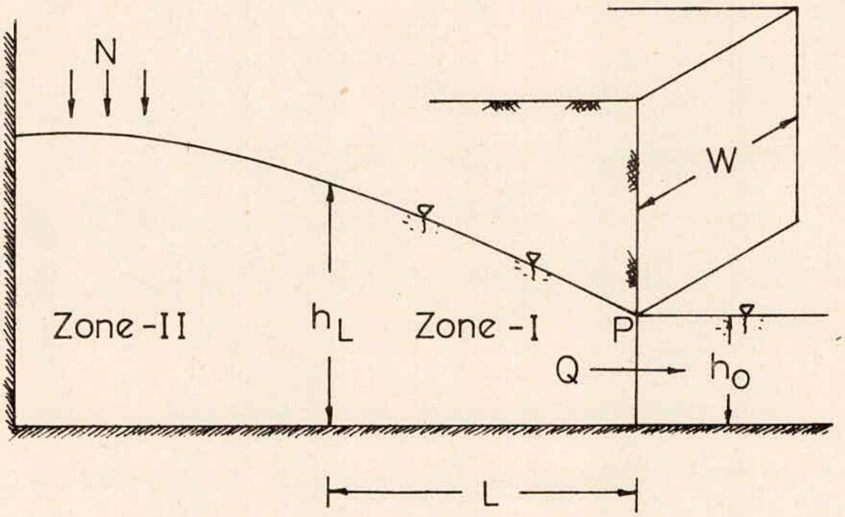


Fig.11 Another simple model of a spring (after Bear,1979)

An unsteady flow can be considered as succession of steady states. The unconfined flow in Zone I has been approximated to follow Dupit's conditions and flow rate at any time has been expressed as

$$Q = WK \frac{h_L^2 - h_o^2}{2L} = WK \frac{h_L + h_o}{2} \cdot \frac{h_L + h_o}{L} = WT(h_L - h_o)/L \tag{10}$$

where,

Q = rate of flow from spring,

K = permeability of the aquifer,

T = average transmissivity of the aquifer,

$(h_L - h_o)$ = Difference of head above spring's threshold point P,

L = Lenth of the transition zone.

As the spring discharge is linearly proportional to the head available, so by comparison the eq.(10) with eq.(1), $\alpha = \frac{WT}{L}$ and after putting the value of α in eq.(9), he obtained

$$\frac{\alpha}{SA} = \frac{WT}{L} \cdot \frac{1}{SA} = \beta$$

So, the equation (9) could be put as

$$Q = Q_o \exp[-\beta (t - t_o)] \tag{11}$$

where β represents the aquifer characteristic.

If the aquifer contributing to the spring flow is made up of several separate subregions, then the each subregion will have its own characteristic coefficient β.

The coefficient β or any other coefficient in one form or other appearing in the expression like equation (11) describing a spring recession curve are related to the aquifer's geometry, transmissivity and storativity. Therefore, it is possible to investigate about these aquifer properties by the analysis of the hydrograph of a spring discharge as an inverse problem. It is assumed in the above mentioned analysis that there is no pumpage or recharge takes place during the analysed depletion period of the spring flow.

3.1.3 Unicell model

This model envisages to interpret the time series spring discharge data to obtain quantitative information on the groundwater system and to predict spring discharge. It is assumed that the spring is perennial with seasonal rainfall and has a well defined outlet (Mandel and Shiftan, 1981).

The groundwater system and the spring is simulated very roughly by the unicell model. The model is visualized as a tank with vertical walls filled with porous material. The tank has a spout at the bottom (Fig.12).

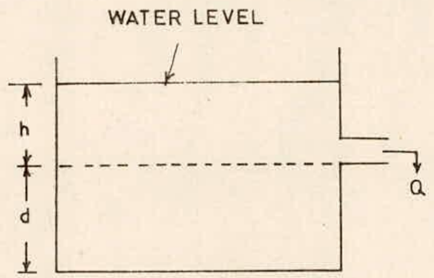


Fig.12 Unicell model for spring flow

The porous medium is saturated with water to a certain depth and discharge coming out of the spout is observed.

During the dry season, the flow of the spring fed from a thick aquifer is computed from the model.

$$\text{If } d \gg h, \quad h+d = b = \text{constant} \tag{12}$$

$$v(t) = ASh(t)$$

$$Q(t) = KbCh(t) \tag{13}$$

$$Q(t) = -\frac{dv}{dt} \tag{14}$$

where, h(t) is the elevation of the water level above the outlet, V the volume of water stored above the outlet, d the aquifer thickness below the outlet, A the base area of the tank, S, K are the storativity and permeability of the aquifer respectively and C is a dimensionless parameter representing the flow pattern.

The elimination h(t) from equation (12) and (13) gives

$$v(t) = \frac{AS}{KbC}\, Q(t) = t_0 Q(t) \tag{15}$$

$$Q(t)+t_0\frac{dQ}{dt} = 0 \tag{16}$$

After solving the equation (16), the solution

$$Q(t) = Q_0 \exp(-\frac{t}{t_0}) \tag{17}$$

Takinglogarithm on both sides of (17),

$$\log Q(t) = \log Q_0 - \frac{1}{2.3}\cdot\frac{t}{t_0} \tag{18}$$

where, t_0 is the depletion time and Q_0 the initial discharge. This solution (18) will plot straight line on a semilogarithmic scale (log Q vs t) (Fig.13).

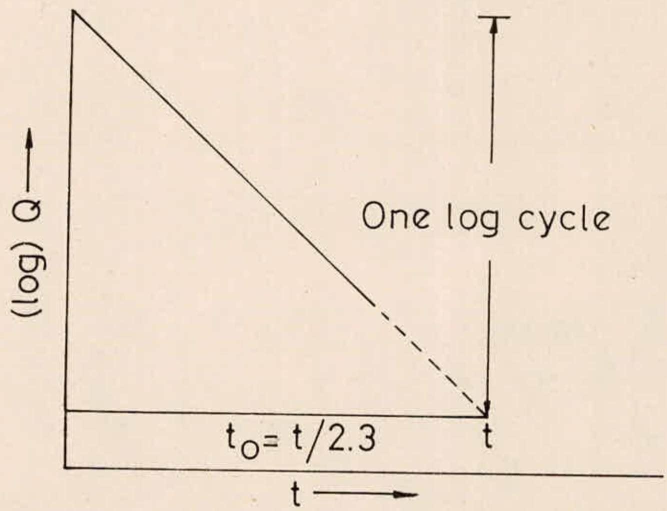


Fig.13 Depletion line analysis

The study of these above mentioned equations and their solutions give the following information.

1. The depletion time t_0 can be read off easily from the semilog plot of dry season discharge.

2. Prediction of the future dry season discharge is possible with

the help of equation (11) if the discharge at any time during the dry season is known a priori.

3. The slope of the depletion line between two given spring discharge Q_1 and Q_2 provides the characteristic of the groundwater system from which the spring emanates. Any change in the slope of the line from year to year is indicative of interference in the groundwater system. A progressive flattening and steepening of the slope indicate the replenishment of the aquifer in the supposedly dry season (probably due to return flow of irrigation/urban effluents or seepage from surface water reservoir) and groundwater abstraction from the aquifer respectively. Occurrences of rate natural catastrophe like earthquake can also have effect on spring discharge and the slope of the line could be flattened or steepened without any recharge or withdrawal from the aquifer feeding the spring.

4. The live reserve of groundwater that maintain spring discharge is equal to $(Q_t . t_0)$ at any given time.

5. Aquifer replenishment between the end of one dry season and the beginning of the next one can be estimated by equation (15) with the aid of principle of continuity

$$AR = Q_2 t_0 - Q_1 t_0 + \int_{t_1}^{t_2} Q dt \tag{19}$$

where R is the replenishment (LT^{-1}), A the replenishment area of the spring (L^2), t_1, t_2 are the instance of time at the end of one dry season and the beginning of the next one, and Q_1, Q_2 the spring discharges at t_1 and t_2 respectively.

6. Estimates of A, S and h derived from hydrogeologic consideration should yield by Eq.(12), a volume of live storage approximately equal to the volume computed by Eq.(15). Estimates of replenishment derived from the assumed replenishment area and from climatic data should

be approximately equal to the figure computed by Eq.(19). A vigorous revaluation of the accepted hydrological concepts of the area is warranted in case of large discrepancies between these estimates.

For a thin aquifer, (permeable veener formation overlying an impermeable bedrock) the spout is assumed at the extreme bottom and therefore, the aquifer thickness is a function of time.

$$\text{So,}\quad d=0,\ h=h(t) \tag{20}$$

$$Q(t)=KbCh^2(t) \tag{21}$$

Combining boundary condition (19) and (20) with equations (12) and (14)

$$Q(t+\Delta t) = \frac{Q(t)}{(1+\Delta t/t_0)2} \tag{22}$$

$$v(t) = Q(t)t_0 \tag{23}$$

The depletion time t_0 is determined by computing the ratio $Q_{(t+\Delta t)}/Q_t$ for successive time intervals Δt and averaging the resulting values. The Eqs. (22) and (23) are essentially applicable for intermittent springs.

Depletion lines may be composed of two linear segments with different slopes (Fig.14). This could happen when

1. There is diversion of water upstream from the spring discharge measuring point,

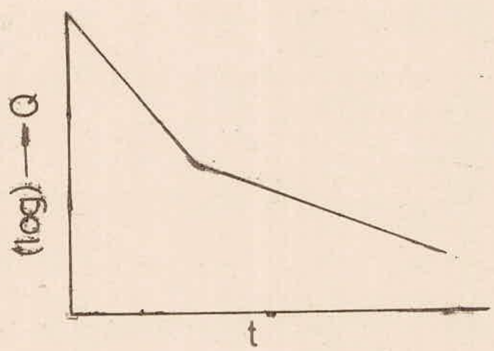


Fig.14 Depletion line with two different slopes

2. Vertical variability of the product A x S which the model assumes to be constant,

3. Delayed runoff(interflow) entering the spring through soil mantle and

4. Two underground reservoir (common in fissured and karstic rocks) contributing to the same spring.

3.1.4 Model depicting influence of fractures

G.Kovacs (1981) showed the influence of different sized fractures on spring flow. The model reported is very simple. It is conceptualised as the depletion of a cylindrical reservoir. The water leaves through a pipe, which is filled with porous material. The length and area of the cross section of the pipe are indicated by l and f respectively and hydraulic conductivity of the material filling the pipe by K. (Fig. 15 a).

At a point of time t the amount of the outflowing water is equal to the change of storage:

$$F.dh = Q.dt = fK \frac{h}{l} dt \qquad (24)$$

$$t = \frac{Fl}{fK} \ln h + C \qquad (25)$$

$$h = h_0 \exp [-a(t-t_0)] \qquad (26)$$

$$Q = Q_0 \exp [-a(t-t_0)] \qquad (27)$$

where initial flow rate at a point of time t_0(i.e $Q_0 = h_0$ Kf/l) depends on the height of the water level(h_0)at the same point of time and the hydraulic transmissibility of the outlet (Kf/l). The constant a = fK/lF is inversely proportional to the storage capacity of the system (horizontal area of the reservoir in the model) and the coefficient of proportionality is the reciprocal value of the parameter characterising transmissibility.

Now, if two reservoir are drained through the same outlet(Fig.15b), their flow rates have to be added to get the instantaneous discharge at a

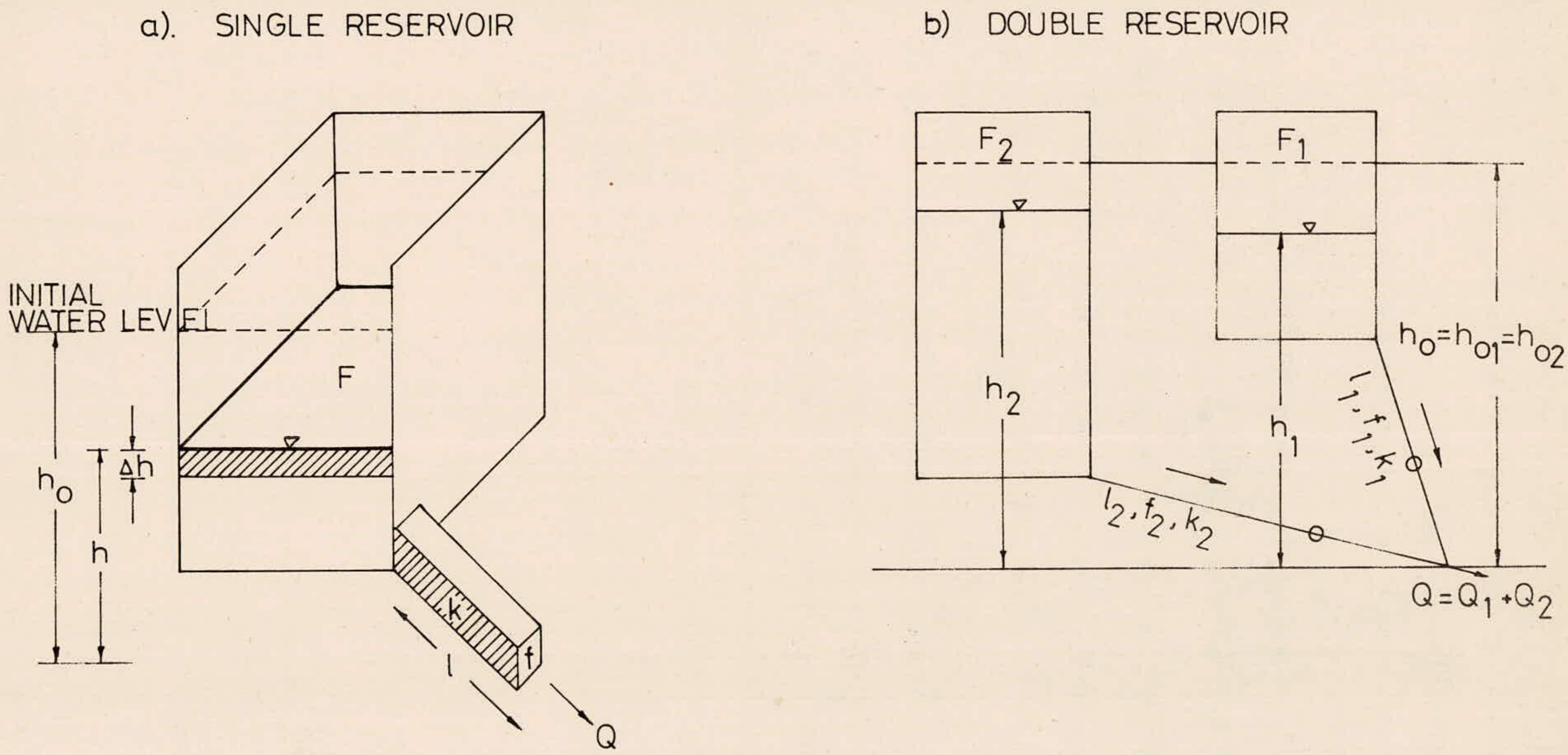


Fig. 15 Model Depecting Influence of Fractures

point of time t:

$$Q=Q_{01}\ \exp[-a_1(t-t_0)]+Q_{02}\ \exp[-a_2(t-t_0)] \tag{28}$$

The comparison of the mathematical model with observed yields of karstic springs proves the accuracy of such approximation of the recession curve and the constants can be calculated from the measured data.

4.0 GEOLOGY OF SPRING FLOW

4.1 General Description

The occurrence of springs is chiefly controlled by the local geology and geomorphology and the drainage system. Springs occur where downgradient parts of aquifer or other water bearing materials with their lower boundary are exposed to the surface like at hill sides, canyons or dissection due to erosion channel. Springs also form where discontinuities like faults or tectonic fractures or dikes present hydraulic barriers and force groundwater to flow upward or where faults cause weak spots in confining layers allowing water to flow upward and reach the surface if the piezometric surface in the aquifer is sufficiently high. The rainwater fills up in the fractured rock fissures which then flows through the same fissures system to form spring at lower points. In some areas, springs occur in alluvial deposits filling river valleys. Such groundwater outcrops are found where river valley has become narrower, thus reducing the cross-sectional area of the groundwater flow. Solution activity and karstification of limestone produce sizeable springs. Springs could occur below surface water, fresh or salty.

Hydrogeologic investigations have demonstrated that gravity springs (occurring at the contact between an aquifer and aquitard) are chiefly found in plains, whereas both gravity and artesian springs occur in the mountains. The large springs are usually associated with permeable aquifers. Geologically speaking, these permeable aquifers are cavernous limestone, porous basalt and sorted gravel. The order of magnitudes of well yields that can be expected from the various geologic materials are given to get an idea of spring flow emerging out of such formations.

1.	Sorted or coarse sand gravels, porous basalts	1000-20,000 m^3/day
2.	Cavernous limestone	500-5000 m^3/day
3.	Sand and gravel mixture, sandstones	100-2000 m^3/day
4.	Fractured and weathered rocks	10-500 m^3/day

4.2 Springs in carbonate rocks

Carbonate rocks are composed of the carbonates of calcium and magnesium in varying proportions, often admixed with clay and siliceous components. The real characteristic of carbonate rock hydrogeology is the presence of numerous and wide cavities through which water circulates. The cavities are genetically developed due to water circulation and are formed by the attack of dissolved carbonic acid on the limestone. Atmosphere provides for only a minute proportion of the carbonic acid, remainder comes from the air contained in the soil.

Carbonate rock springs represent constricted discharges with abundant flows at widely separated outlets. However, there is no emergence of the water table in the sense of a spring extending uninterruptedly along a wide stretch of aquifer front and with a feeble discharge per unit of front. At the start, lines of weakly discharging springs may develope. The largest spring progressively captures the water feeding its neighbours and they become dry. This is especially so in a strongly developed karstic form. The peculiar landscape thus formed is called karst after the name of a locality in Yugoslavia. Gravitational water available in fissures, corridor and cavity systems of karstic rocks (limestone, dolomite, gypsum, halite) is called karstic water. The term karstic rock is used often as a synonym of carbonate rock, although this is not absolutely correct. Karstic indicates basically the development of chemically enlarged openings in a rock mass. This can occur in a non-carbonate sediments as well (gypsum) and there are carbonate rocks in which the developments of dissolved openings is not common (e.g. carbonate marls).

Solution by groundwater proceeds according to the following principles (Mandel and Shiftan, 1981):

(1) Imagine a limestone aquifer having some initial - primary or secondary porosity. The water picks up carbon dioxide in the nonsaturated zone above the water table, dissolves calcite during its flow through the aquifer and emerges from the natural outlets saturated with calcium bicarbonate.

(2) Consider a flow channel leading from the water table to an outlet and assume, for simplicty, that it is not connected to any other channels. Under these conditions the rate of erosion by dissolution is proportional to the discharge of the cannel.

(3) The largest rate of erosion will occur along flow paths that offer the least resistance to the passage of water through the aquifer. These preferred flow paths may be formed by interstices along bedding planes, by faults or by fissures. Openings that carry a relatively large discharge are preferably widened by karstic erosion.

(4) An array of solution channels, even though each one is only one or several milimeters wide, strongly increases the hydraulic conductivity of the aquifer and, consequently, the discharge in a particular direction. Thus an ever-increasing part of the water flowing through the aquifer is concentrated into a system of solution channels.

(5) The water table is lowered and outlets fed from the part of the aquifer gradually cease to function. The pore space between solution channels, where the water becomes semistagnant, may be plugged by calcite deposits. In the vicinity of a natural outlet, where flow paths converge, veritable caverns are created. Eventually one spring - usually the one at the lowest outcrop - grows into giant proportions, leaving older spring outlets "high and dry" (Fig.16a & b).

(6) When karstic erosion runs its full course, the pore space of the aquifer is replaced by a few wide caverns, and eventually the whole limestone massif collapses and is washed away. These spectacular developments are of interest to geomorphologists. The hydrogeologist is concerned mainly with the invisible initial and intermediate stages of the process, which enhance aquiferous properties rather than destroy them.

This conceptual model of karstic erosion by groundwater is amply supported by observational evidence.

The presence and state of a spring are often determined by fractures, faults, or tectonic cracks, and also by joints. On the side of a valley the karst water escapes through a complex of fissures parallel to the valley floor which enable the water to come out some distance from the toe of the slope (Renault , 1967, 1968, 1970 vide UNESCO 1972). Schmidt (1963) stated that the dissolving and mechanical effects of water in limestone are of secondary importance. Caves and headings are developed in the tectonically preformed directions. Generally sinkholes are

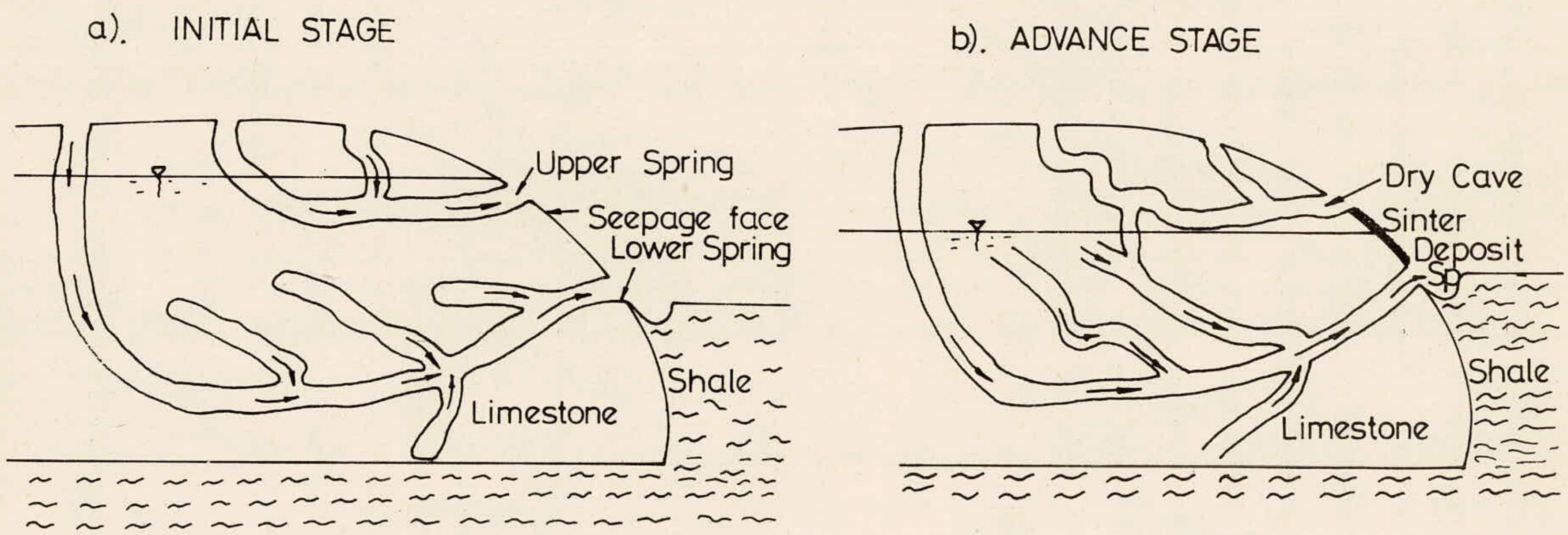
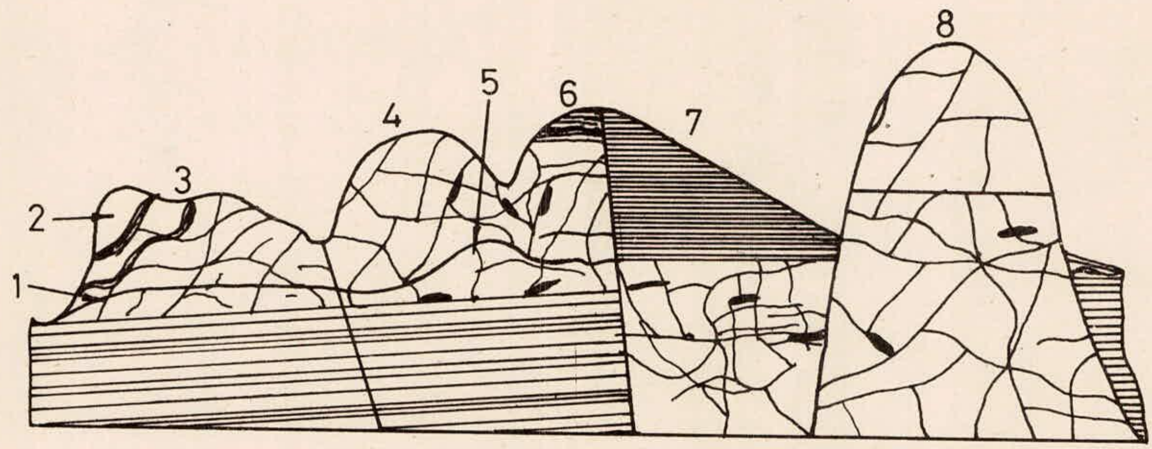


Fig.16 Development of solution channel

developed in thin, plated, loose and disturbed limestone.

Downward moving karstic water is affected by gravity and in general, has a free water surface. Reaching an impermeable stratum, a leaning karstic zone may develope. In broken-fold mountains, leaning karstic waters are connected with each other. Their near horizontal piezometric plane is called the main karstic level. If the main karstic level crosses the ground level, springs may develope. Horusitzky (1953) classified karstic water genetically (Fig.17). Physical, chemical and biologic aspects of karstic waters needed to be examined when used for drinking and industrial purposes.



1. Leaning 2. Descending Karstic Zone 3. Shallow Karst
4. Deep Karst 5. Lifted Karstic Water 6. Free Surface Covered Karst
7. Tied Covered Karst 8. Extruding Karst

Fig.17 Classification of karstic waters (after Horusitzky, 1953)

According to the stability of operation, permanent, periodic and seasonal springs can be distingushed. Seasonal springs operate during a longer period in a year before going dry for a considerable time; - characteristic types are pseudo springs appearing after snow melt or heavy rainfalls and flood spring above a permanent spring (Fig.18).

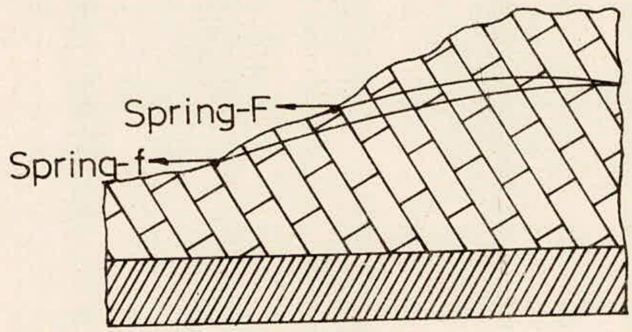


Fig.18 Seasonal karstic spring, flood spring (F) above a permanent spring(f)

Periodic springs stop operation time to time and start working after a well determined time period; - geysers and intermittent karstic springs belong to this category, the latter operating like a siphon.

The Vaucluse spring in France is well known worldwide. It has a recharge area of over 1600 sq.km. and is formed of strongly jointed and karstified Neocomian limestone. This spring is a periodically flowing one and works as a siphon and is termed as siphon spring (Fig.19).

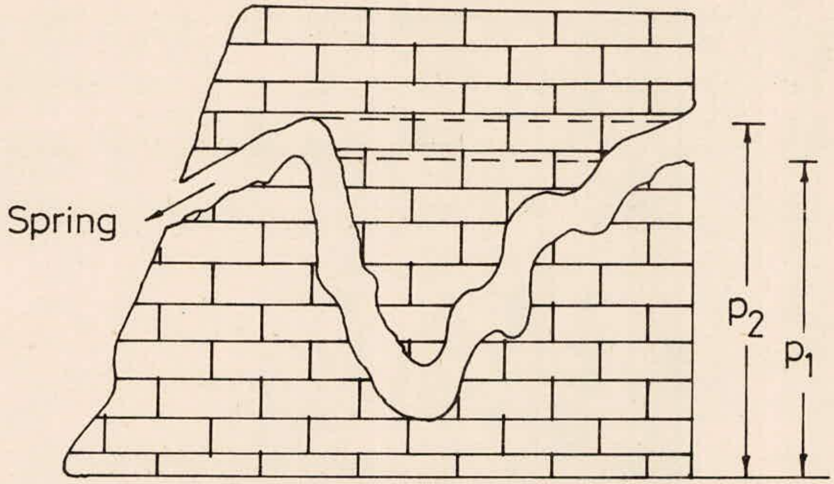


Fig.19 Siphon spring of Vaucluse type

The spring emerges from the huge grotto found in the deep canyon. The annual average yield is 17 m^3/s whereas the maximum recorded yield during spring time is 152 m^3/s. The area receives, on an average 550 mm precipitation annual (Klimentov, 1983).

Incidentally, the largest known spring in the world issues from a limestone at the flow rate approximately of 40 m^3/s at Ras-el-Aim via its tributary, the Khabour (Price, 1985).

It is common to find lines of springs where permeable limestone from high ground which rest on less permeable rocks such as shales or clays (Fig.20).

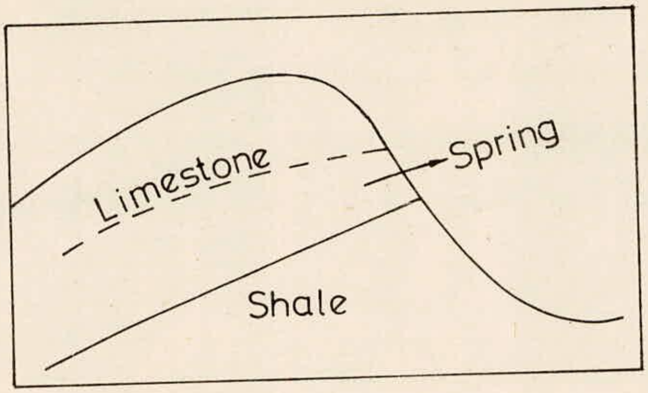


Fig.20 A line of springs emerging from limestone

In the study of Gaula catchment in Nainital district, U.P. the investigators of Kumaun University (1988) observed that few spring emerges from carbonate rocks characterised by cavities and solution channels. The yield of the springs of the karst belt is very high and is in the order of 3763 x 10^3 litres/day. Since the dolomitic limestone has a good networks of joints, there is a good deal of infiltration of water. Selective solutions along these fractures and joints has created network of water courses leading to almost complete lack of perennial streams overground.

4.3 Springs in basalts

Out of all fractured rock formations, basalts are amongst the most productive because of numerous opening they contain. Basalts are spreaded all over the world :- 500,000 sq.km. in India, the trap rocks of the Deccan; 650,000 sq.km. in North America; 900,000 sq.km. in Brazil; 100,000 sq.km. in Ireland and U.K. They represent extensive sources of groundwater. Springs emerging from basalts are among the largest known and compare in importance with those issuing from carbonate rocks. The flow from the basalt springs may be relatively constant where it is sustained by the large waterbody and where the fissures are narrow and ash is present. On the other hand, the flow may be variable if wide fissures are well developed and abundant.

Non-vesicular basalt has a low porosity. Even with an abundance of vesicles, the permeability of rock itself is low. The porosity and permeability values of basalts in USA and Morocco as observed are given in Table 3 and 4.

Table 3. Hydraulic conductivity and porosity of basalts in USA (after Morris & Johnson, 1966)

Parameter	Min.	Max.	Average	No.of samples
Hyd. conductivity	2.1×10^{-9}	4.7×10^{-5}	1.05×10^{-5}	93
Porosity (in %)	3	35	17	94

Table 4. Transmissibility and hydraulic conductivity values of basalts in Oudja region, Morocco(after Mortier, Quang and Sadek, 1967)

Description	K(cm/s)	T(m^2/s)
1. Weathered basalts and fissured compacted ash	1.7×10^{-2} to 2×10^{-1}	1.1×10^{-2} to 3×10^{-2}
2. Basaltic tuffs and slightly fissured compact basalt	1.2×10^{-1} to 7.5×10^{-1}	1.3×10^{-5} to 7.5×10^{-3}

One of the largest series of basalt springs are in Idaho, U.S.A. The springs above the canyon walls between Milner and King Davis discharge about 110 m^3/s, each of the springs yielding 4 to 16 m^3/s. The famous thousand springs, which rise between the basalt flows yield 15 to 20 m^3/s (Meinzer, 1927).

One group of basalt springs in California and Oregon, USA has a combined discharge of 40 m^3/s of which Datta spring alone supplies 1.4 to 3 m^3/s. The springs at Oahu and Kaluaoopu islands in Pearl Harbor, the spring discharge vary between 0.4 to 0.7 m^3/sec(UNESCO, 1972).

As reported in the report published by Kumaun University (1988), the Bhimtal volcanics in Nainital district of Uttar Pradesh, there are vesicles in basalts and foliation in tuffite, in addition to vertical and horizontal joints and shear planes. The depth of weathering varies from 60 cm to 2.5 m. About 9% of the springs in the Gaula river catchment are located in Bhimtal volcanics.

4.4 Classification of springs on the basis of structural tectonics in the Gaula river catchment, Nainital district

Kumaun University conducted geohydrological investigations of the Gaula catchment (1988). The investigators studied the formation of the springs in the catchment. Based on the genesis, nature of water bearing formations and conditions governing the formation of the springs, the springs of the Gaula river catchment are classified into 7 categories. These are stated briefly.

1. Fracture - joint related springs

These are formed in fractures and joints in hard rocks and emerge either along the hillslopes or in stream beds wherever the local water table is intersected by fractures or joints or by the ground surface. The average seasonal discharge of such spring in the Gaula river catchment varies from 0.9 to 145 ($x10^3$) litres/day.

2. Lineament - Fault related springs

The groundwater rises through pathway available due to fault plane and lineaments extending deep underground.

The discharge of lineament-fault related springs in the area varies from 25 to 326($x10^3$) litres/day.

3. Colluvial springs

Usually, the low upslopes (8^o to 12^o) of the recharge area provide a good hydrualic gradient. Locally, there could be an overlapping conditions of colluvial and fracture related springs. In such cases, the spring discharge of colluvial springs is augmented by down flowing groundwater along the fracture. The discharge varies from 3.7 to 57($x10^3$)litres/day.

4. Springs related to fluvial deposits

Here, the spring issues from the contact of alluvial fans and terrace deposit on terrace scarps. The discharge of these springs is high but vary widely from 2 to 2505($x10^3$) litres/day in the study area.

5. Bedding plane related springs

The bedding and/or schistosity plane serves as the path for emergence of groundwater as spring. The discharge of such spring varies from 1.8 to 168($x10^3$) litres/day.

6. Dyke related springs

Springs occur due to dyke serving as a barrier to groundwater movement in the country rock.

7. Karst springs

Water emanates through the cavities and channels in dolomites and limestones and give rise to karst springs. Such springs are there in Nainital area. The diameter of underground channels and cavities varies from 0.50 to 0.75 cm.

5.0 DEVELOPMENT OF THE PROPOSED MODEL

A mathematical model is proposed herein which will assess the strenght of a spring emerging from an unconfined aquifer.

5.1 Model Configuration:

Configuration of the proposed model has been designed by combining the two simple models of Bear (Figs.10 & 11). Model described in Fig.10 is the zone II of the proposed model (Fig.21).

5.2 Statement of the Problem:

h_0 is the initial level of groundwater table before the onset of recharge in zone II at which the spring is inactive. A variable recharge occurs in the recharge area of the spring (zone II) from t=0 to t=t and groundwater table rises to h(t). The time parameter has been discretised into uniform time steps. During a time step the recharge rate is assumed to be constant, but the recharge rate varies time step to time step.

In the recharge zone (zone-II) which has a surfacial area 'A' per unit length the direction of groundwater flow is vertical and as such the flow in zone-II is one dimensional. P is the threshold point of the spring outlet where the boundary head is h_o. The storage in the gronundwater due to variable recharge over time t, N(t) which will subsequently be discharged through the spring needed to be assessed. In other words, the dissipation or variation of this recharge at spring outlet as Q(t) is to be determined. T and ϕ are the average transmissivity and storativity of the aquifer.

5.3 Assumptions:

i) Assumed flow pattern in the model is two dimensional in the vertical plane,

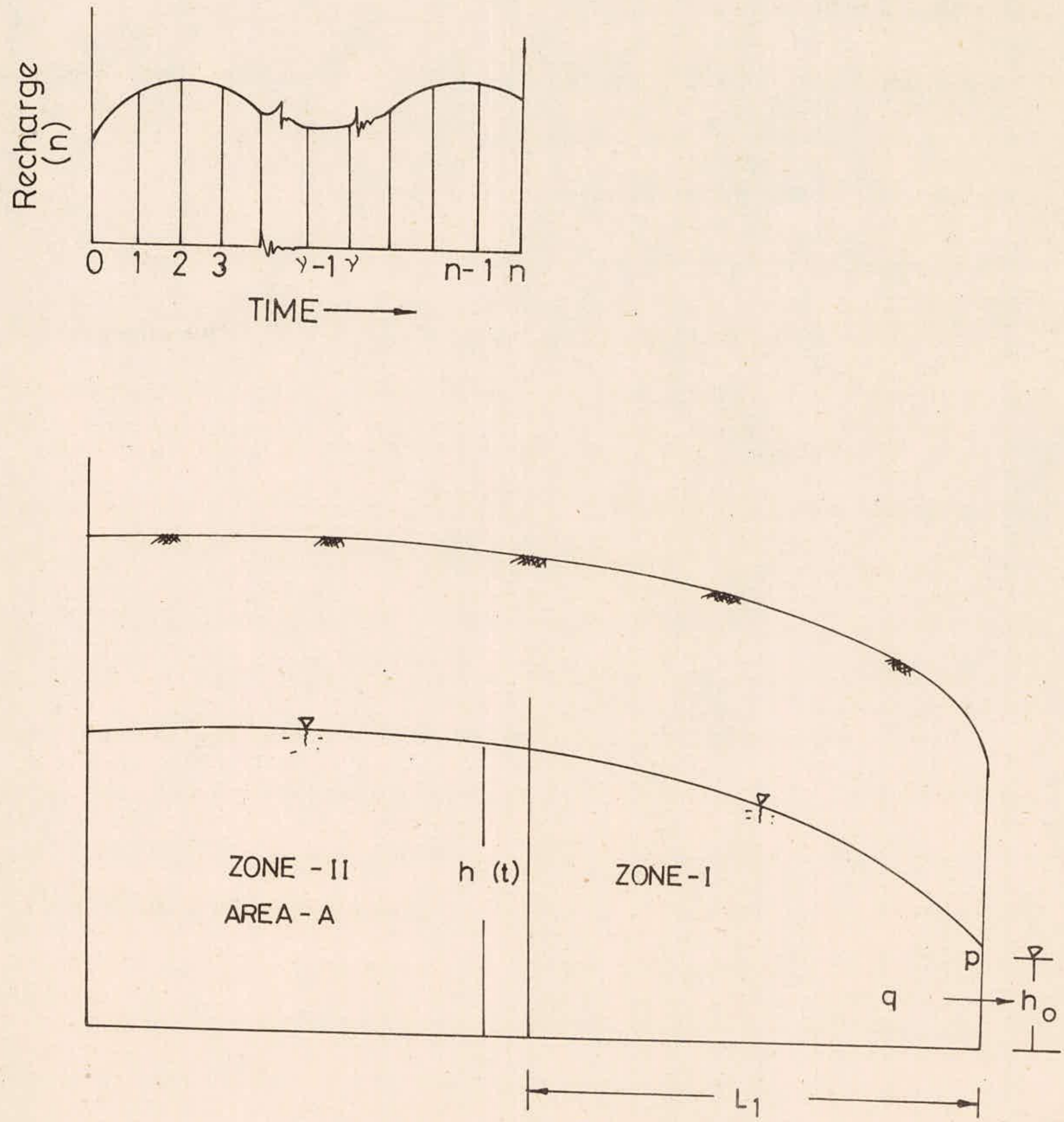


Fig. 21 Proposed model configuration

ii) Flow in zone-I follows Dupit-Forchheimer assumption and the flow is in horizontal direction.

iii) Unsteady state problem has been assumed to be succession of the steady state condition.

iv) Flow leaving zone-I through the spring threshold 'P' is given by the linear equation

$$Q(t) = \frac{T[h(t)-h_o]}{L_1} \tag{29}$$

5.4 Methdology

Let, $\Delta h(t)$ be the fall in water level in zone-II during time Δt due to spring flow. Hence,

$$Q(t) \quad t = - h(t) \phi A \tag{30}$$

Replacing Q(t) by Eqn.(29) and rearranging, Eqn.(30) reduces to

$$\frac{dh(t)}{h(t)-h_o} = \frac{T.dt}{L_1 A \phi} \tag{31}$$

Integrating

$$\log[h(t)-h_o] = - \frac{Tt}{L_1 A \phi} + C \tag{32}$$

Assuming that a recharge of N occurs on zone-II at time t=0, the head at t=0 is given by

$$h(o) = h_o + \frac{N}{\phi}$$

The constant appearing in Eqn.(32) is given by

$$C = \log \frac{N}{\phi}$$

Sub tituting the value of C in Eqn.(32) and simplifying, the following expression for head in zone-II is obtained after putting $\alpha = T/L_1$.

$$h(t) = h_o + \frac{N}{\phi} e^{-\frac{\alpha t}{A\phi}} \tag{33}$$

The spring flow Q(t), is evaluated by substituting the expression of h(t) in Eqn.(29) and spring flow in given by

$$Q(t) = \frac{T}{L_1}[\frac{N}{\phi} e^{-\frac{\alpha t}{A\phi}}] \tag{34}$$

5.4.1 Development of the model for time variant recharge:

In practice, the recharge to aquifer would occur over a span of time depending upon precipitation values. A typical time variant discontinuous recharge pattern is shown in Fig.(21).

$N(\nu)$ is the average recharge rate during the time step All the recharge rates are assumed to be independent and constant in the time interval $(0,1),(1,2),\ldots\ldots\ldots\ldots,(\nu-1,\nu),\ldots\ldots\ldots\ldots,(n-1,n)$. For N=1, Eqn. (34) gives the response of the system to a unit impulse recharge. Let the response of the system to a unit impulse excitation be designated by K(t).

K(t) is given by

$$K(t) = \frac{\alpha}{\phi} e^{-\frac{\alpha t}{A\phi}} \tag{35}$$

For variable recharge, the spring outflow is given by

$$q(n) = \int_o^1 K(n-\tau)N(1)d\tau + \int_1^2 K(n-\tau)N(2)d\tau$$

$$+ .. + \int_{\nu-1}^{\nu} K(n-\tau)N(\nu)d\tau + .. + \int_{n-1}^{n} K(n-\tau)N(n)d\tau$$

$$= \sum_{\nu=1}^{n} N(\Upsilon) \int_{\nu-1}^{\nu} K(n-\tau)d\tau \tag{36}$$

The integral in Eqn.(36) can be put in a more standard format by using the discrete kernel, $\delta(n)$. A change of variable of integration in Eqn.(36) is the discrete kernel (Morel Seytoux and Daly, 1975) with argument $(n-\gamma+1)$ and Eqn.(36) takes the form

$$q(n) = \sum_{n=1}^{\nu} \delta(n-\gamma+1)N(\tau) \tag{37}$$

If unit recharge takes place in first unit time period and no recharge occurs thereafter, the spring outflow corresponding to this unit pulse excitation can be obtained by starting from the response of the spring to a unit impulse excitation given by Eqn.(34) and is given by $\delta(n)$.

$$\delta(n) = A\, e^{-\frac{\alpha n}{A\phi}}[e^{-\frac{\alpha}{A\phi}} -1] \tag{38}$$

The response to a unit pulse excitation at time step 'n' designated as $\delta(n)$ is given by

$$\delta(n) = \int_{o}^{1} \frac{\alpha}{\phi}\, e^{-\frac{(n-\tau)}{A\phi}}\, d\tau \tag{39}$$

$\delta(.)$ is the response due to the unit pulse excitation of linear system which was initially at rest before the onset of recharge. So, once the $\delta(.)$ are generated by the model and saved, these could be used for any set of varying recharge over time as $\delta(.)$ are the properties of the aquifer and are independent of the excitation.

5.4.2 Utility and testing of the model

In the field, we can measure spring flow discharge with time and the rainfall distribution in the area is usually available. But, the transmissivity, storativity of the aquifer and the percentage of rainfall recharge to the groundwater are not known.

This model could be used to solve the inverse problem of finding out the fraction of rainfall going to aquifer as recharge and aquifer parameters by making use of the rainfall and spring flow values. Once the aquifer parameters are calibrated and rainfall in the area is known, it is possible to predict the future spring flow.

The developed model has been tested with an assumed time variant (daily) rainfall distributed over 90 days (monsoon period) on the recharge area. The transmissivity of the aquifer has been assigned low value since it is intended to model spring of small strength. Because the infiltrated water from rainfall will first satisfy the soil moisture deficiency in the unsaturated zone before replenishing groundwater, the groundwater recharge from rainfall is taken as 10% of the rainfall for the first 10 days of rainfall and it is 20% for rest of the rainy season. A lag of two days has been assumed from the day of rainfall to the day of groundwater recharge. Aquifer flow domain is taken to be homogeneous. The model is used to simulate the spring flow for 365 days after the first day rainfall (i.e. after the onset of rainfall). The two nondimensional parameters representing discharge from the spring and the time are plotted in the logarithmic scales. This plot of spring flow for variable recharge is shown in Fig.22 along with the assumed daily rainfall pattern.

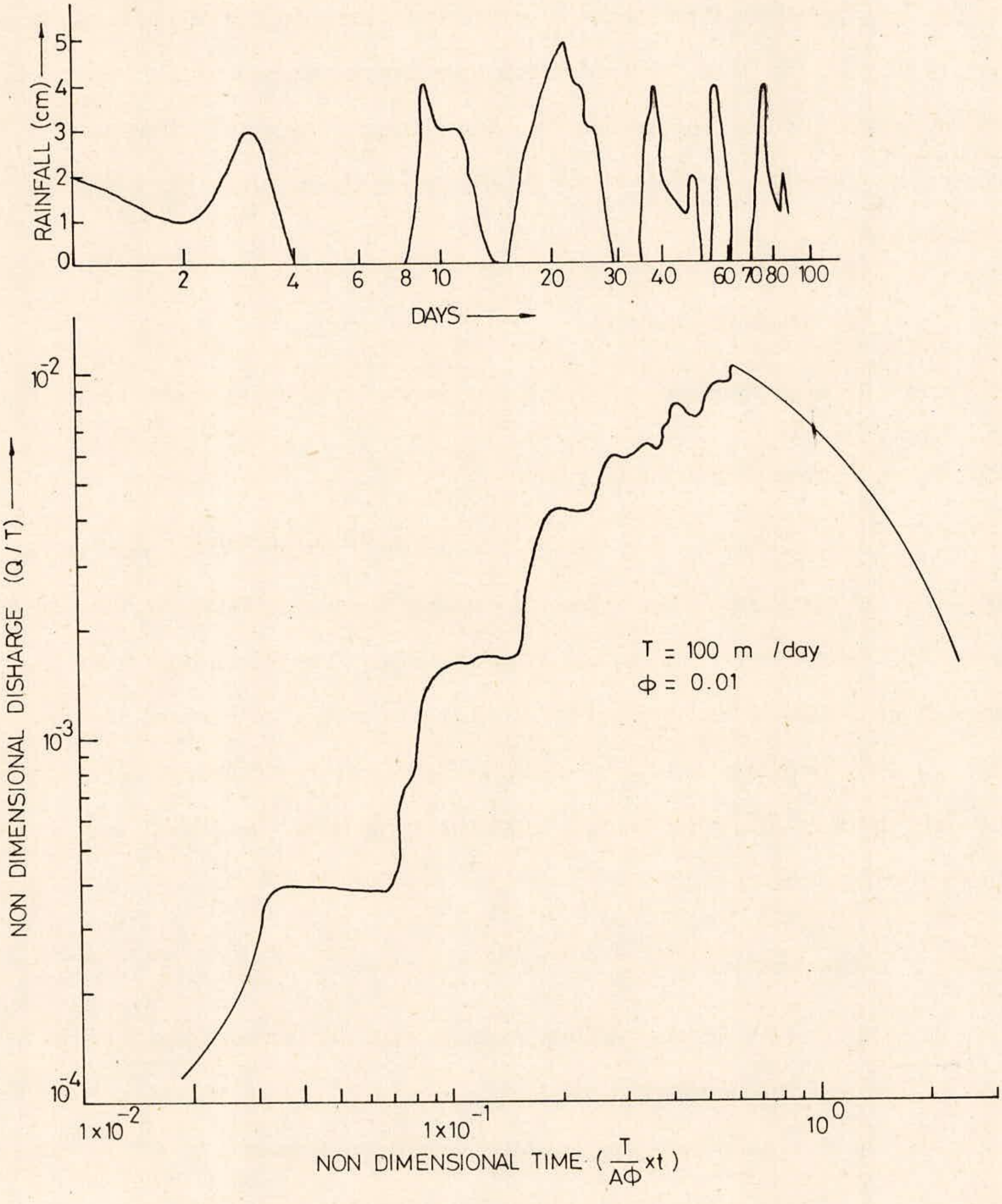


Fig.22 Non-dimensional plot of spring flow under variable recharge

6.0 DEVELOPMENT OF A SPRING FLOW MODEL AS FLOW FROM A RESERVOIR THROUGH A PIPE

A mathematical model is proposed conceptualising the flow from the spring as the flow from depleting reservoir through pipe. A solution channel in limestone aquifer can be considered as a pipe. The spring flow reservoir with its out-crop through which recharge takes place can be considered as a tank.

6.1 Model Configuration:

Configuration of the proposed model has been shown in Fig.23.

6.2 Statement of the Problem:

Let h_0 be impulse rise in the level of groundwater due to the recharge 'R' per unit area. Due to spring discharge through threshold point 'p', the water level in the aquifer falls. The discharge from the spring with time is to be predicted. Let the recharge area of the aquifer be 'A'; the coefficient of permeability of the material filled in the pipe be 'K'; the area of the pipe be 'a'. Let the pipe be of length 'l' and ϕ be the storativity of the aquifer.

6.3 Assumptions:

1. Flow in the spring reservoir is one dimensional and is in the vertical direction.
2. Unsteady state problem has been assumed to be the succession of the steady state condition.

6.4 Methodology:

The spring flow discharge is given by

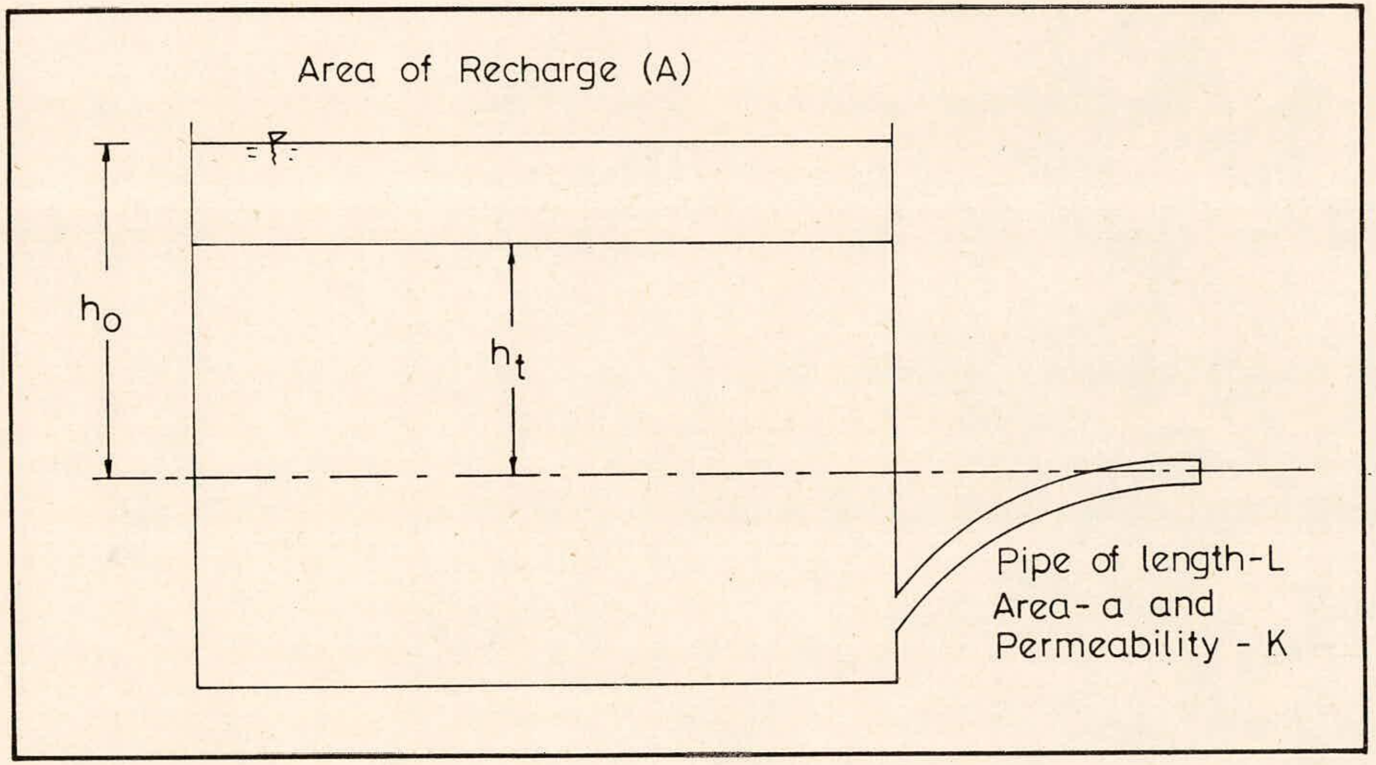


Fig.23 Conceptualised tank pipe model for spring flow

$$q(t) = Ka \frac{h(t)}{L} \qquad \ldots(37)$$

Groundwater level in the spring reservoir due to spring flow is

$$h(t) = h_0 - \frac{1}{A\phi} \int_0^t q(\tau)d\tau \qquad \ldots(38)$$

Differentiating,

$$\frac{dh}{dt} = - \frac{q(t)}{A\phi} \qquad \ldots(39)$$

$$\frac{dq(t)}{dt} = \frac{Ka}{L} \frac{dh}{dt} = - \frac{Ka}{L} \frac{q(t)}{A\phi} \qquad \ldots(40)$$

$$\frac{dq(t)}{q(t)} = - \frac{Ka}{L} \cdot \frac{1}{A\phi} \cdot dt$$

Integrating,

$$\log_e q = - \frac{Ka}{LA\phi} t + C \qquad \ldots(41)$$

At $t = 0$, $q = q_0$

So, $C = \log_e q_0$

Putting the value of constant of integration in Eqn.(41),

$$\log_e \frac{q}{q_0} = - \frac{Kat}{LA\phi}$$

So, $$q = q_0 e^{-\frac{Kat}{LA\phi}} \qquad \ldots(42)$$

Substituting the value of q_0 in Eqn.(42)

$$q = \frac{Ka\ h_0(t)}{L} e^{-\frac{Kat}{LA\phi}}$$

$$= Ka \frac{\frac{R}{\phi}}{L} \cdot e^{-\frac{Kat}{LA\phi}}$$

$$= \frac{KaR}{\phi L} e^{-\frac{Kat}{LA\phi}} \qquad \ldots(43)$$

In practice, the recharge to aquifer would occur over a span of time depending upon precipitation values.

$R(\gamma)$ is the average recharge rate during the time step γ. All the recharge rates are assumed to be independent and constant in the time interval $(0,1),(1,2),\ldots..(\gamma-1,\gamma),\ldots..(n-1,n)$. For R=1, eqn.(43) gives the response of the system to a unit impulse recharge. Let the response of the system to a unit impulse excitation be designated by k(t). k(t) is given by

$$k(t) = \frac{Ka}{\phi L} e^{-\frac{Kat}{LA\phi}} \qquad \ldots(44)$$

For variable recharge, the spring outflow is given by

$$q(n) = \int_0^1 k(n-\tau)R(1)d\tau + \int_1^2 k(n-\tau)R(2)d\tau ,+\cdots+ \int_{\gamma-1}^{\gamma} K(n-\tau) R(\gamma)d\tau +\cdots+ \int_{n-1}^{n} k(n-\tau)R(n)d\tau$$

$$= \sum_{\gamma=1}^{n} R(\gamma) \int_{\gamma-1}^{\gamma} k(n-\tau)d\tau \qquad \ldots(45)$$

The integral in eqn.(45) can be put in a more standard format by using the discrete kernel, $\delta(n)$. With the change of variable $\tau-\gamma+1=v$, eqn.(45) reduces to, $q(n) = \sum_{n=1}^{\gamma} \delta(n-\gamma+1)R(\tau)$ in which,

$$\delta(m) = \int_0^1 K(m-\tau)d\tau$$

$$= \int_0^1 \frac{Ka}{\phi L} e^{-\frac{Ka(m-\tau)}{LA\phi}} d\tau$$

$$= Ae^{-\frac{Kam}{LA\phi}} [e^{\frac{Ka}{LA\phi}}-1]$$

$\delta(m)$ is the response due to unit pulse excitation of the linear system which was initially at rest before the onset of recharge. So, once the $\delta(m)$ are generated by the model and saved, these could be used for any set of varying recharge over time as $\delta(\cdot)$ are the properties of the aquifer and independent of the excitation.

6.5 Utility and Testing of the Model:

This model can be used to analyse the springs flow emanating from the hard rock area where the spring flow emerges from a solution cavity. Like the previous model, this model could be used to solve the inverse problem of finding out the fraction of rainfall going to aquifer as recharge and aquifer parameters by knowing the rainfall and spring flow. After rainfall in the area, it is possible to predict the future spring flow.

The model has been tested with an assumed time variant (daily) rainfall distributed over 90 days (monsoon period) on the recharge area. The infiltrated water from rainfall will first satisfy the soil moisture deficiency in the unsaturated zone before replenishing spring reservoir. The recharge due to rainfall is taken as 10% of the rainfall for the first 10 days of rainfall and it is 20% for rest of rainy season. A lag of two days has been assumed from the day of rainfall to the day of ground water recharge. Aquifer flow domain is taken to be homogeneous. The model is used to simulate the spring flow for 365 days after the first day rainfall. The discharge from the spring through the conceptualised pipe output has been plotted with respect to time. This plot of spring flow for variable recharge is shown in Fig.24 alongwith assumed daily rainfall pattern.

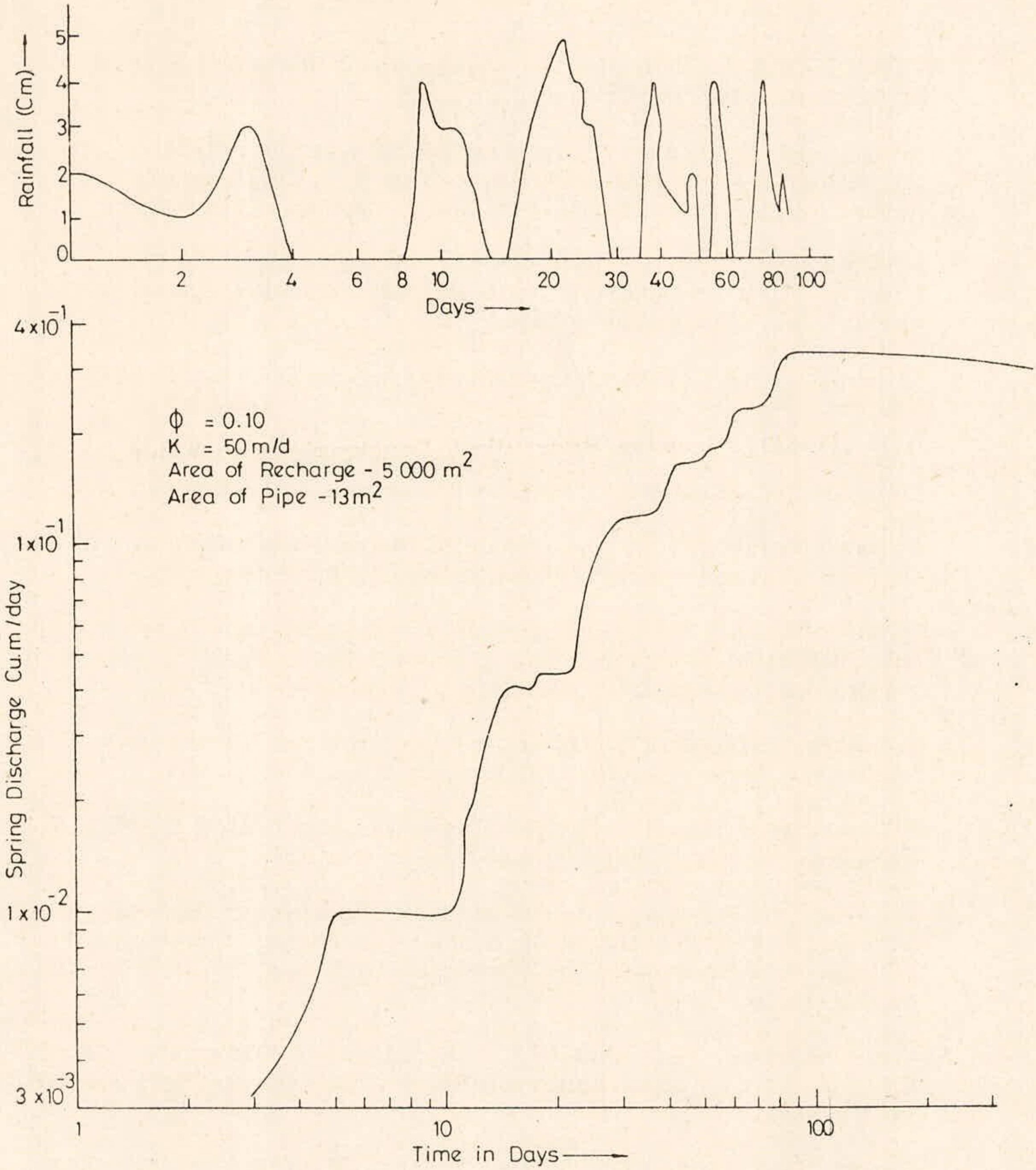


Fig.24 Plot of predicted spring flow with time

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