DEPARTMENT OF GEOGRAPHY DISCUSSION PAPER SERIES

DRAINAGE BASIN HYDROLOGY: A REVIEW AND SYNTHESIS

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River Studies: Part II

Drainage basin hydrology: A review and synthesis

by

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Preface and Acknowledgements

This paper is the second of a four-part review and discussion of the basic principles and theories of river behaviour. Part I provides an overview of recent developments in the study of drainage-basin morphology and of drainage network characteristics while Parts III and IV respectively provide discussions of the fluid mechanics of open-channel flow and of sediment transport theory in rivers.

This discussion paper is a relatively comprehensive review of drainage basin hydrology, although it is intentionally biased towards hydrologic conditions in Canada. For example, the treatment of snowmelt runoff is more detailed than those found in standard texts. Emphasis has also been given to certain topics which are directly relevant to the discussion in other parts of this River Studies series.

Throughout this work I have attempted to "translate" the often highly technical discussion into a form that a generalist in the earth sciences would find useful. To what extent I have succeeded in this attempt, is a judgement for you, the reader, to make.

Where points and arguments are clearly made in the text, they often reflect that benefit of student criticism of earlier drafts evolved over some 14 years of teaching fluvial geomorphology. I am also grateful for the helpful comments of many colleagues particularly that of Gerald Nanson and Ken Page. Of course I remain completely responsible for the shortcomings of the paper.

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2.1 Introduction

Water that flows over the surface of the land to form rivers and to shape the earth's fluvial landscapes is ultimately derived from atmospheric moisture. The routes followed by water from the atmosphere to the world's rivers, however, often is exceedingly complex. The study of the ways in which water is stored in and transferred among atmosphere, oceans, lakes and land is the science of hydrology. It is a science which has its beginnings in the speculations of Greek and Roman philosophers and its modern form rooted in the recorded observations of Renaissance scholars such as Leonardo da Vinci and Bernard Palissy (see Chow, 1964 for a brief historical review of hydrology).

A fundamental concept of modern hydrology is the hydrologic cycle. This concept treats the movement of water among atmosphere, ocean and land as a continuous circulation or closed system of water transfers. It can be applied at a world scale (see Barry, 1969) or at the level of smaller spatial units (see More, 1969).

For our purposes it will be useful to consider in some detail the operation of the hydrologic cycle at the scale of the drainage basin. The drainage basin, with its bounding watershed, forms a logical unit for hydrological studies; application of the hydrologic cycle at this, rather than at larger, scales is conceptually simpler, and the basin provides a convenient basis for water balance measurements. Furthermore, the basin is a fundamental geomorphic unit in the sense that it defines the limit of fluvial sculpture of the landscape that can be attributed to a given river. It is in basin-scale investigations that the complex links between hydrology and fluvial geomorphology are most likely to be identified.

2.2 The basin hydrologic cycle

The hydrologic cycle, as it applies to a drainage basin, is schemmatically shown in Figure 2.1.

The most direct route of water moving from the atmosphere to the river is represented by precipitation falling directly into the river channel (channel precipitation). Another relatively direct route is represented by water which falls to the land surface and then flows downslope as overland flow to the river. Most water, however, follows a less direct route to the river. For example, some may fall to the ground, infiltrate into the soil to become part of the groundwater zone, and then move relatively slowly under the influence of gravity as subsurface flow until it reaches the river. Some of this throughflow may move into deep groundwater zones and be lost to the local hydrologic system. This latter component, however, is usually relatively small and can be ignored in most cases (More, 1969).

Water is returned to the atmosphere by evaporation from ground, vegetation, lake and river surfaces. Some is returned by evaporation directly from falling raindrops. A large quantity of water also is drawn from the soil by vegetation and returned to the atmosphere by evapotranspiration.

Water also may be stored at various points in the cycle for varying periods of time. For example, water may remain on the leaves and stems of

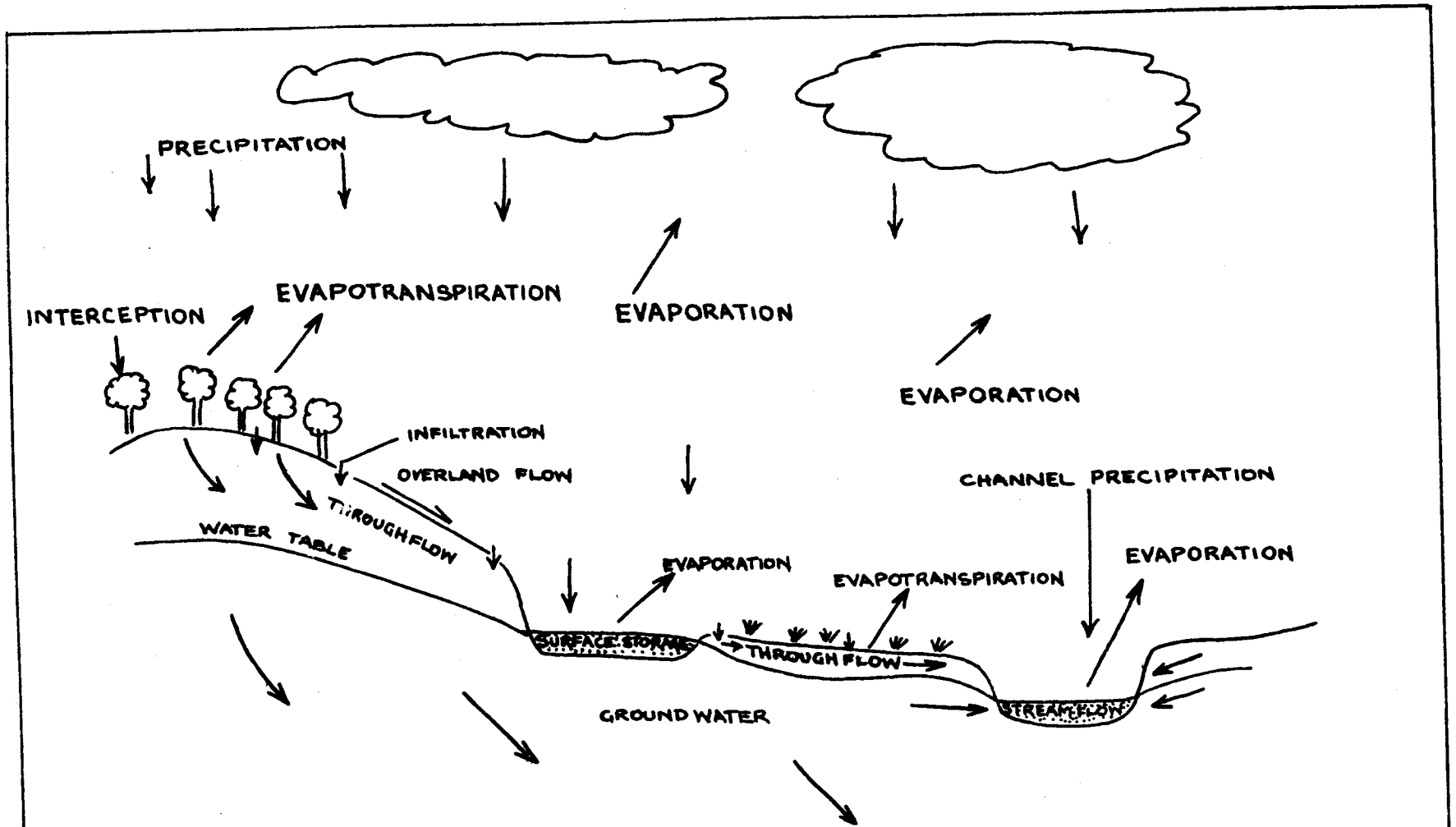


Figure 2.1: Components of the basin hydrologic cycle.

vegetation (interception storage) in small ground depressions, and in lakes (surface storage) or as soil moisture or groundwater.

Thus the gross operation of the basin hydrologic cycle may be expressed by a simple continuity equation as follows:

precipitation = runoff and evapotranspiration + changes in storage

$$\text{or} \quad p = q_b + E_{vt} + \Delta(I_s, L_s, M_s, G_s) \tag{2.1}$$

where p is precipitation, q_b is total basin runoff, E_{vt} is evapotranspiration and I_s, L_s, M_s, and G_s are respectively interception, surface, soil moisture, and groundwater storage components.

To the fluvial geomorphologist, total basin runoff is the term of direct concern because it in part determines the size of rivers draining the basin. However, although equation 2.1 provides a useful framework for considering the water budget of a drainage basin, it is of limited use in predicting runoff in its present simple form. Each term of the equation represents a complex set of relationships which warrant further discussion. Such a discussion, however, must be placed in a specific temporal context. For example, we could consider the water budget of a drainage basin for an individual storm event, or for a period of a month, a year, or a number of years. For the sake of discussion we will first examine the factors influencing the operation of the hydrologic cycle over the period of an individual storm event. It is at this time scale that the basin hydrology is the most complex; for longer periods the effects of averaging somewhat simplify the hydrologic responses.

2.3 Precipitation

2.3 (a) The nature and causes of precipitation

Precipitation occurs when air containing water vapour cools sufficiently to produce condensation and growth of cloud droplets or ice crystals (see Gilman, 1964, for a detailed discussion of this process). The only known mechanism capable of producing a large enough lowering of temperature to account for observed precipitation rates is the pressure reduction associated with ascending parcels of air. As air is lifted from near the earth's surface to upper levels in the atmosphere the decreasing atmospheric pressure allows it to expand and adiabatically cool. Precipitation can be genetically classified according to the lifting mechanism responsible for the adiabatic cooling (see Bruce and Clark, 1966) or it can be simply classified according to its form (see Barry, 1969).

Four major types of lifting mechanisms are commonly distinguished: horizontal convergence, orography, convection, and weather fronts. Horizontal convergence is the process by which wind fields concentrate the inflow of air into a particular area such as a low pressure zone thereby forcing the air to rise. Orographic lifting occurs when flowing air encounters and is forced to rise over a topographic barrier such as a mountain range. Precipitation resulting from this process is common along the slopes of the Rocky Mountains. Convection occurs when differential heating or advection results in air becoming more bouyant than its surroundings. The air consequently rises and is often

associated with intense but very local thunderstorm activity. Weather fronts are the zones which separate large air masses which have markedly different physical properties. Most low pressure areas in temperate and polar regions have frontal systems associated with them. Fronts can be regarded as discontinuities in density within the atmosphere, forming boundaries between relatively light warm air and a wedge of heavier underlying cold air. It is the process of the lighter air mass riding up over the wedge of denser air that may cause cooling and consequent widespread precipitation.

Although these different types of precipitation have quite different physical bases, they often occur together; for example, flow convergence is often associated with frontal activity.

2.3 (b) The measurement of precipitation

Precipitation can occur in a variety of forms but for our purposes we need recognise only two basic types: rain and snow. Rainfall is measured by a variety of instruments, most designed for daily observations, but also others for continuous recording over longer periods. Specific designs vary from one country to another (see Gilman, 1964; World Meteorological Organisation, 1965; Rodda, 1969; McKay, 1970; Gregory and Walling, 1973) but each attempts to meet the requirements for accurate rainfall assessment. The standard precipitation gauge such as those used in the United States, Canada and Britain, is simply a cylinder set on the ground and topped with an orifice or collector formed by a sharp rim bevilled to the outside (see Figure 2.2 A). Rain or snow falling into the collector is funnelled to a graduated receiver in which it can be measured. The collector is deep and the funnel has at least a 45° slope to prevent splash loss. The receiver has a narrow neck and is protected from radiation in order to minimise evaporation loss.

The continuously recording instruments most commonly used are the tipping bucket, weighing, and float gauges. The first consists of two balanced buckets which tip back and forth as they are filled in turn by rainfall directed to them by a collecting funnel (see Figure 2.02B). As the balance swings about its pivot it opens and closes an electrical contact which is linked with a recorder. This type of gauge is well suited to remote (telemetering) recording stations where there is no snowfall. The weighing and float gauges simply record respectively the mass and height of the water in a calibrated gauge receiver. Only the weighing recording gauge is suited to the measurement of both rain and snow.

Precipitation gauges designed to collect and store snow have relatively large orifices (12-30 cm) to prevent snow capping and usually contain a charge of an antifreeze solution such as ethylene glycol to convert the collected snow to a liquid. Oil is also added to the receiver to suppress evaporation from the large orifice. Snow gauges usually have some type of shielding device around the collector in order to minimise wind flow effects at the orifice; the Nipher shielded snow gauges used by the Meteorological Service of Canada are of this type (see Figure 2.02C).

The conventional method of measuring snowfall used in most

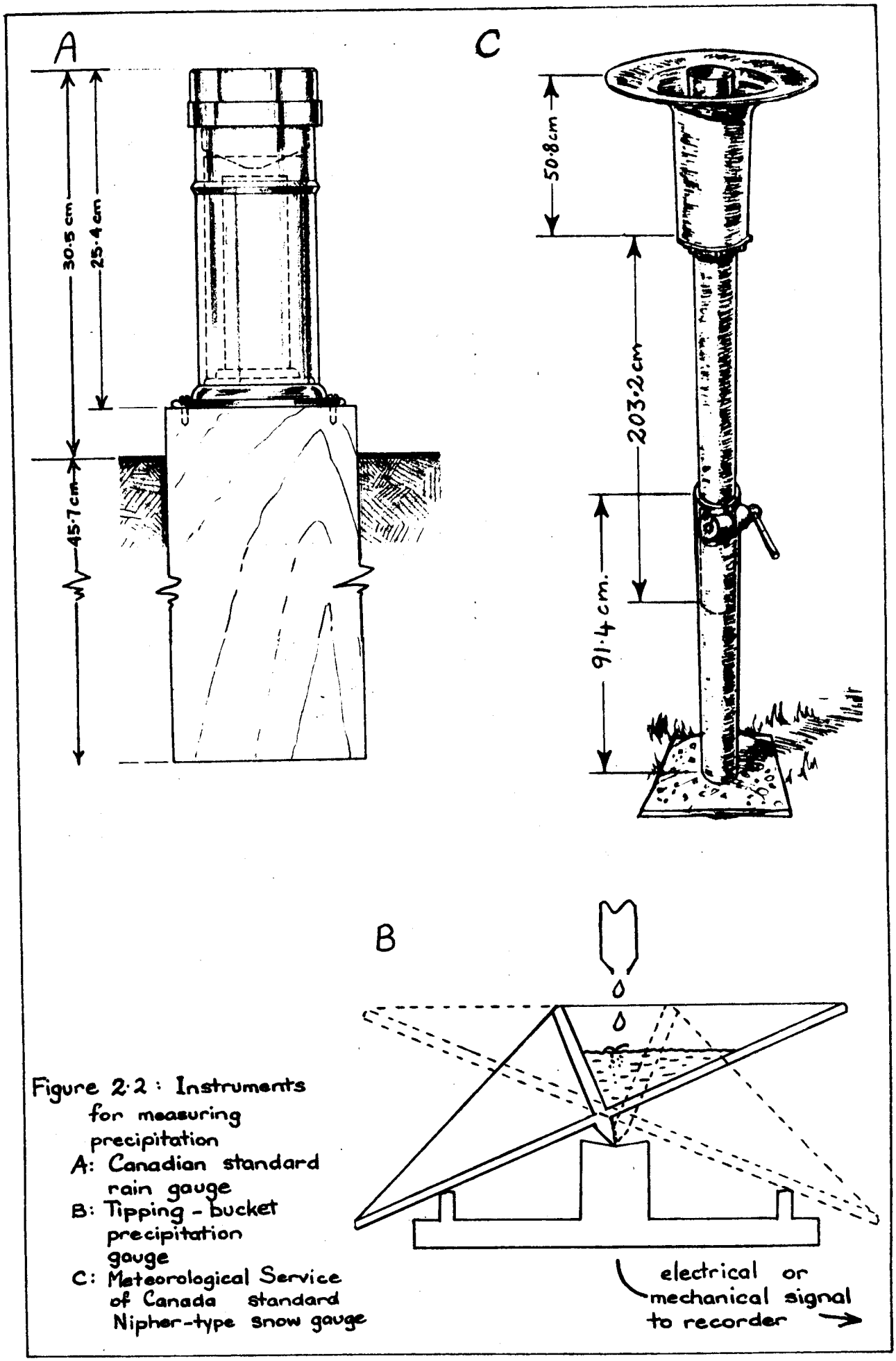


Figure 2·2 : Instruments for measuring precipitation
A: Canadian standard rain gauge
B: Tipping - bucket precipitation gauge
C: Meteorological Service of Canada standard Nipher-type snow gauge

countries simply involves the measurement of accumulated snow on the ground with a ruler. The average of several points selected so that they are free of snow drifting effects and other local disturbances is used as a representative snow depth. The average density of snow is such that water equivalents are assumed to be one tenth of the snow depth. For large areas such as a drainage basin, snow survey courses are laid out as a basis for measuring the depth of the snowpack. At this scale the principles of sampling are the same as for single station measurements: a number of representative locations are established as part of a permanent network (see Garstka, 1964).

More recently, snow pillows or pressure pillows have been used to measure depths of snow, rate of snowfall and rate of snowmelt (see Kerr, 1976). The pressure pillow is essentially an air mattress filled with antifreeze solution. It is placed at a representative site at the beginning of the snow season and monitors snow accumulation (or loss) by pressure changes in the mattress. Because these devices are inexpensive and easy to operate, and measurements are easily recorded or telemetered, and because they integrate a relatively large area (usually about $10m^2$) and have a very sensitive and almost instantaneous measurement response, they offer considerable promise in improving the accuracy of snowfall measurements, particularly in remote locations (see McKay, 1970).

Aerial photography is commonly used to determine the snowpack extent early in the melt season and photogrammetry is increasingly being used to estimate snowpack depths (see recent developments in Rango, 1977).

An excellent source of information regarding the performance characteristics of methods and instruments of snow measurement is the respectively annual and occasional Proceedings of the Western and Eastern Snow Conferences in the United States.

2.3 (c) The character of rainfall

Three important characteristics of storm rainfall are intensity, duration, and distribution over the basin. For longer term hydrologic studies it is also important to know the frequency of storm events.

Rainfall intensity is simply the rate of rainfall, usually expressed in mm/hour. Rainfall intensity can vary from a fraction of one mm/hour in light showers to several hundred mm/hour in tropical downpours (see Jennings, 1950; Hersfield, 1961). We shall soon see (Section 2.8) that the character of river floods is greatly influenced by the intensity of rainfall over the basin.

Rainfall duration is the elapsed time from the beginning until the end of the rainstorm. The total depth of water yielded by a storm is the product the intensity and duration of the rainfall.

The distribution of rainfall over a drainage basin is important both to the understanding of river flood behaviour caused by storms, and to the calculation of the amount of rain fallen. Measurement of the total volume of rain falling in a drainage basin during a storm (or during a longer time period) is complicated by the fact that the

rainfall and the measurement stations are rarely ever uniformly distributed in space. It is therefore necessary to use some spatial averaging technique in order to provide realistic estimates of precipitation over the entire drainage basin. Ideally, the measurement stations should be uniformly and densely distributed over the drainage basin (see Figure 2.3A) so that a simple average of precipitation at all stations would provide a meaningful basin average. However, although it is feasible to maintain such a measurement grid in laboratory work or in specialised field studies, geomorphologists are often interested in natural basins in which rainfall is monitored by only a few scattered meteorological stations. In this latter case it is common practice to base precipitation estimates on one of two basic methods of spatial averaging: the Thiessen polygon method and the isohyetal method.

The first method is well suited to areas with few precipitation stations which are unevenly distributed over and around the basin. Each station represents a proportion of the basin area determined by its proximity to other stations (see Figure 2.3B). Lines are drawn between adjacent stations and a grid of polygons formed from the perpendicular bisectors of these lines. The watershed forms the boundary of the peripheral ploygons. Each station precipitation value, P_n, is then weighted by its Thiessen polygon area, A_n, relative to the total basin area, A_t, in the form $P_n\ (A_n/A_t)$; the sum of these weighted values equals the spatially averaged precipitation.

The second and preferred method of spatial averaging involves the construction of an isoline map joining points of equal precipitation (isohyets - see Figure 2.3C) in the basin. The area between each pair of isohyets is measured and used to weight the average precipitation between the isohyets in the same was as is done in the Thiessen polygon method. The average precipitation between isohyets is usually taken as the arithmetic mean of their values.

The isohyetal method is more subjective and less amenable to machine processing that the Thiessen polygon method but it does allow an experienced operator to take account of non-linear variations in the pattern of precipitation caused by such things as topography and storm patterns. It is thus potentially more accurate than the Thiessen polygon method.

It is, of course, often appropriate to modify these two basic methods to accommodate particular requirements of accuracy and computational efficiency (for examples, see Cole, 1962; Clarke, 1976, Hickin, 1978).

2.3 (d) The Character of snowmelt

The second major source of precipitation, snowfall, is generally a smaller component of the basin hydrologic cycle than is rainfall. Nevertheless, it may be a very substantial part of total precipitation

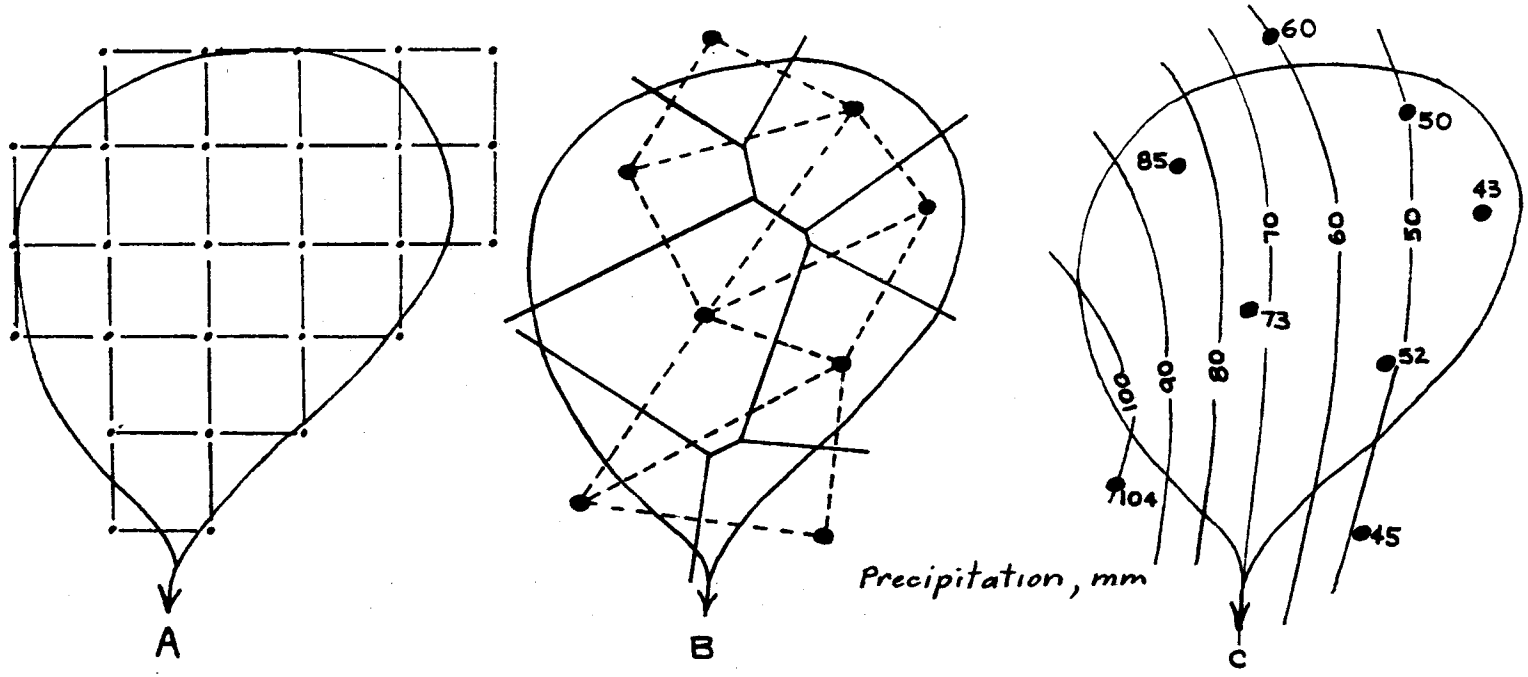


Figure 2·3 : Spatial averaging methods for determining mean basin precipitation

A : The regular grid method.

B : The Thiessen polygon method.

C : The isohyetal method.

in high latitude/altitude areas. Furthermore, in areas where snowfalls occur regularly each winter, snowmelt is usually a very important factor influencing the magnitude of the Spring flood. This influence is both in the form of direct runoff contribution and in the form of basin conditioning (wetting) for rainfall runoff. It is also the case, however, that flood peaks in snowfall areas are more commonly the direct response to rainfall. It is in this type of situation that the distinction between floods and the general seasonal rise in river levels becomes rather blurred.

The principal difference between the influence of rain and snow on runoff is that, whereas rain immediately after ground impact infiltrates the soil or forms overland flow, snow is usually stored for a varying length of time on the basin surface. Typically, snow falls and accumulates during the winter and is slowly released from storage as meltwater during the spring. The snow may completely melt during the summer or it may accumulate to form glaciers in high latitude/altitude areas. Water stored as glacial ice often will not contribute to runoff until it moves downvalley to a melting environment (see Sharp, 1960; Andrews,1974, for a discussion of glacial budgets). A large proportion of the water content of snow and ice is directly returned to the atmosphere through sublimation and evaporation, or infiltrates the soil as meltwater to become groundwater.

Snow may become an important contributor to runoff when relatively high temperatures produce a melting rate so high that soil moisture reaches and is maintained at maximum capacity well before the supply of snow is exhausted. This process is most likely to occur in the steep and often rocky (relatively impermeable) low-order drainage basins. These conditions prevailed in the extreme in British Columbia in 1948 when a long high snow-yield winter was abruptly terminated by a late warm spring, producing the disastrous Fraser River flood of that year (Hutchinson, 1950).

Any model which attempts to describe the movement of water from a snowpack to the stream channel must address two basically different problems: how is the snowbank converted to water, and how is this meltwater delivered to the stream channel? The latter problem is common to both rain and snow hydrology and we will for the moment focus on the former.

The melting of ice at a point in the snowpack is essentially a thermodynamic process, the rate and amount of melting being dependent on the net heat exchange between the snowpack and its environment. The phase change from $0^{o}C$ ice to water requires the input of some 335,000 joules of heat energy for melting each kilogram of ice. The various sources and processes influencing heat transfer to and from a snowpack are listed below:

1. absorbed shortwave radiation from the sun (R_s)
2. net longwave radiation from the earth and atmosphere (R_b)
3. condensation and vapourisation of water in the air (R_c)
4. convective transfer of heat by the wind (R_h)
5. the heat content of rain (H_r)
6. conduction of heat from the ground (H_g).

The meltwater (M) produced by the net transfer of heat (ΣH) from all sources to the snowpack is given by

$$M = \frac{R_s + R_b + R_c + R_L + H_r + H_g}{334{,}960B} = \frac{\Sigma H}{334{,}960B} \qquad (2.2)$$

where M is the water equivalent of snowmelt (mm), H is the algebraic sum of all heat contributions (joules/metre^2, J/m^2) and B is the thermal quality of the snowpack, defined as the ratio of heat required to melt unit weight of snow to that of ice at 0°C (averaging 0.95 to 0.97 for 5 to 3 per cent liquid water). The constant 334,960 is the heat input in J/m^2 required to produce 1 kg of water from ice at 0°C. Because 1 kg is the mass of 1000 cm^3 volume of water, it is also the heat required to produce 1 mm/m^2 of meltwater.

Most of the following discussion of the character of the various heat components in equation 2.2 is based on the findings of the United States Army Corps of Engineers (U.S.C.E.). Their studies and those of other agencies such as the U.S. Weather Bureau, summarised in the 1956 report "Snow Hydrology", remain the most comprehensive analysis of this component of the hydrologic cycle.

Direct radiant energy from the sun is absorbed by the snowpack and generally forms the most important contributor of heat to meltwater production in open low-slope sites. The net amount of heat absorbed by the snowpack is predictably dependent on factors such as latitude, aspect, time of the day and season, clarity of the atmosphere (degree of cloud and smog cover), type of vegetation cover, and the ability of the snow surface to reflect incoming radiation (albedo). Conventionally, the albedo is taken as 80 per cent for fresh snow and assumed to decrease exponentially to about 40 per cent for melting late season snow (U.S.C.E., 1956).

The intensity of insolation or shortwave solar radiation, expressed in units of J/m^2 per unit time, can be measured directly using a solarimeter or similar instrument or it can be estimated by a variety of methods (see U.S.C.E., 1956; Gray et al, 1970).

The melt component produced by shortwave radiation (M_{rs}) can be expressed as

$$M_{rs} = \frac{(1 - r)\, R_{si}}{334{,}960B} \qquad (2.3)$$

where r is the albedo (as a decimal fraction) and R_{si} is the effective solar radiation in J/m^2/day. Equation 2.3 is shown graphically in Figure 2.4A.

For example, if we consider a typical melting situation in which B = 0.97 and r = 0.60 for an average Canadian June when $R_{si} \doteq 2.929 \times 10^7$ J/m^2/day (cloudless sky; see Hay, 1969), equation 2.2 indicates a meltwater yield from insolation equal to 36 mm/day.

Net longwave radiation from the earth and atmosphere is also an

important factor influencing snowmelt. However, although snow can be regarded as a perfect black body, and longwave radiation emission from the snow surface thus can be estimated from Stefan's Law, the return radiation from the atmosphere for varying degrees of cloudiness and from the vegetation cover (see U.S.C.E., 1956; Gay and Knoerr, 1975) is exceedingly difficult to estimate. For this reason the U.S.C.E. Manual (1960) suggests a simple alternative based on empirical relationships involving air temperatures. Under clear skies in open conditions, the daily smowmelt from longwave radiation (M_{rl}, mm) is given by

$$M_{rl} = 0.970\ T_a - 2.34 \tag{2.4}$$

where T_a is the air temperature (^{o}C) over the snow surface at the 3-metre level. Equation 2.4 indicates that heat is actually lost from the snowpack (negative meltrates) for air temperatures less than 22°C. This condition would thus seem to be the norm under clear skies.

The corresponding expression for the meltrate under complete cloud cover is given in equation 2.5:

$$M_{rl} = 1.326\ T_c \tag{2.5}$$

where T_c is the temperature (^{o}C) of the cloud base. Under these conditions, positive meltrates will occur for all cloud base temperatures above freezing. If the temperature at the 3-metre level is substituted for the cloud base temperature, equation 2.5 also describes the meltrate for the snowpack under a forest canopy:

$$M_{rl} = 1.326\ T_a \tag{2.6}$$

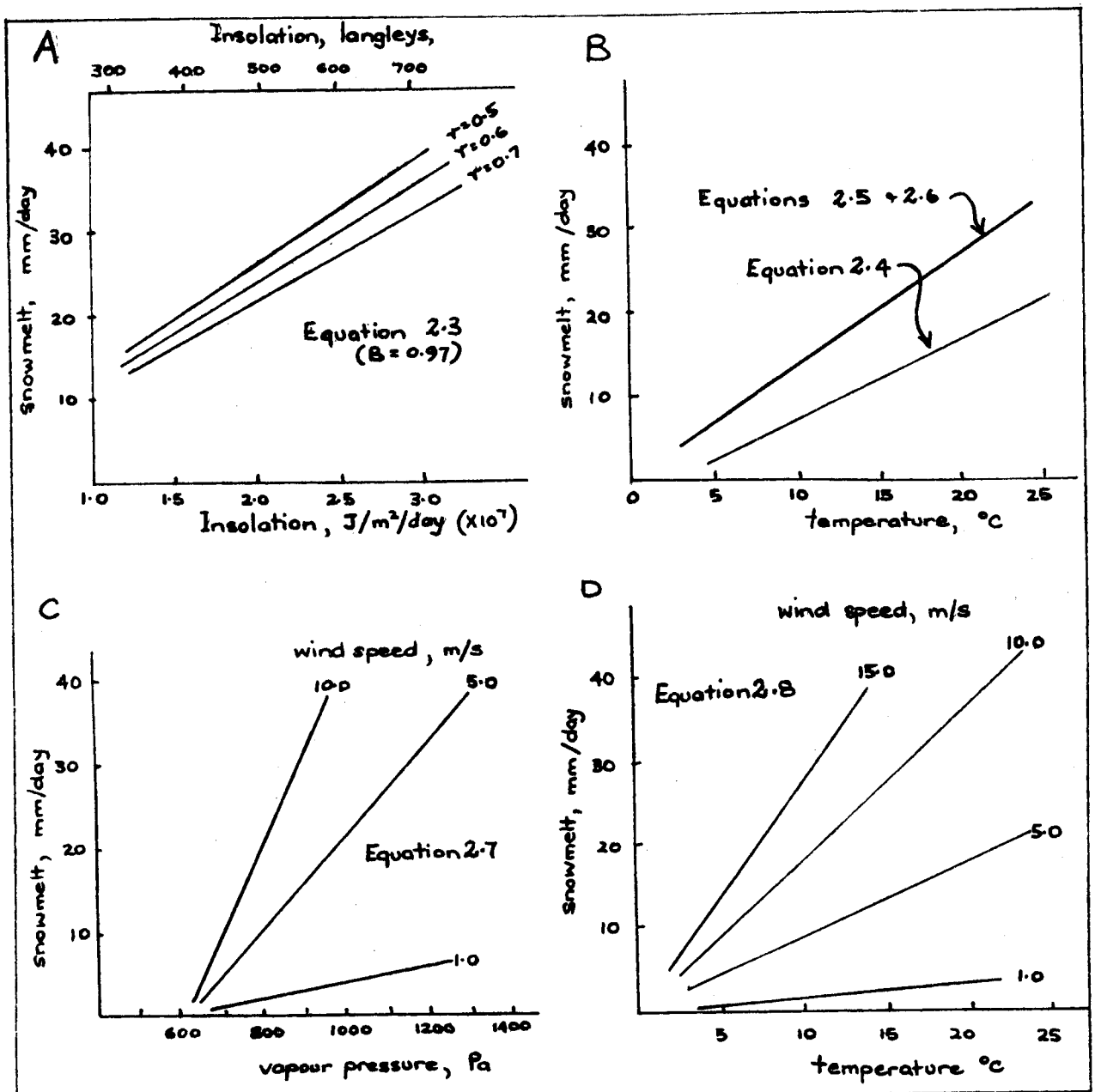
Equations 2.4, 2.5 and 2.6 are shown as graphs in Figure 2.4B.

Clearly vegetation is a very important influence on the radiation budget and on the rate of snowmelt production. A detailed analysis of radiant energy supply of the forest is provided by Gay and Knoerr (1975).

A third important source of heat for melting comes from that released by the process of condensation. When water vapour at 0°C condenses onto the snow surface as 0°C water, the phase change liberates about 2.55 x 10^{6} Joules of heat energy for every kilogram of condensed water. In other words, the latent heat from water condensing onto the snow surface will produce a meltwater mass or volume equal to 7.6 times that of the condensate

$$\left[\frac{\text{latent heat from 1 kg of condensate}}{\text{heat required to melt 1 kg ice}} = \frac{2.55 \times 10^{6}\ \text{J/kg}}{3.55 \times 10^{5}\ \text{J/kg}}\right].$$

The amount of heat absorbed in this way by the snowpack is considered to be primarily a function of the vapour-pressure gradient between the snow surface and the overlying air, and of the wind speed. Vapour pressure is the pressure that would be exerted by the water vapour in the air if all the other gasses were removed. The maximum amount of water that can be held as vapour in the atmosphere depends on temperature (see Figure 2.5) and the air is said to be saturated



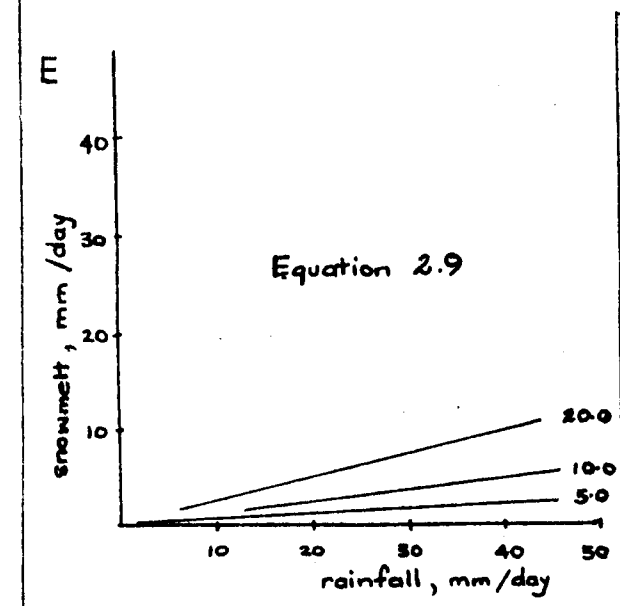


Figure 2.4: Relations among snowmelt rate and atmospheric variables

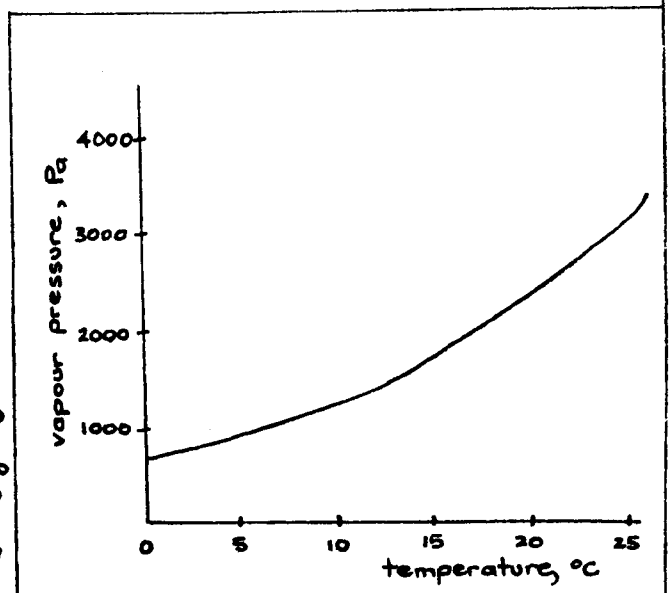


Figure 2.5: Relation between air temperature and saturation pressure

when this amount is reached. The vapour pressure at a melting snow surface is 611 Pascals (P_a). The rate of condensation is clearly dependent on wind speed because air which has lost water vapour through this process must be replaced by moister air for condensation to continue. The rate at which air moves and is replaced over the snowpack is, of course, the wind speed.

The snowmelt produced by condensation may be estimated from the experimentally derived relation (see Figure 2.4C):

$$M_e = 0.0109\ (e_a - 611)\ V_n \qquad (2.7)$$

where M_e is the snowmelt in mm/day, V_n and e_a are respectively the wind speed in m/s at 15 m above the snow surface and the vapour pressure in P_a at the 3 m level.

For example, snowmelt will be 21 mm per day in an air mass of vapour pressure 1000 P_a with a wind speed of 5.0 metres. From the structure of equation 2.7 it is clear that, in air masses with vapour pressures of less than 611 P_a, evaporation rather than condensation (and thus melt) will take place as a result of this exchange process. In such cases there is no melt produced although the volume of the snowpack will be reduced by the evaporative loss.

Snowmelt produced by convection results mainly from heat transferred from warm air advected over the snow surface. The contribution of convection melt to the total is generally small compared with other factors, although it may be important in maritime situations. It may also be important in some mountain range locations. For example, the Rocky Mountains foothills of Western Alberta are particularly subject to rapid snowmelt by heat from the warm chinook. The chinook is a wind formed when easterly flowing air descends on the leeside of the Rockies and is adiabatically warmed; January 1966 saw Pincher Creek, Alberta, experience a chinook induced temperature increase of 21°C in four minutes (Barry and Chorley, 1976)! Less extreme but significant temperature increases related to this type of activity are commonly reported from areas such as the eastern foothills of the Colorado Rockies and of the New Zealand Alps, as well as from the northern flanks of the European Alps and the mountains of central Asia.

The expression developed by the U.S.C.E. (1956) to estimate snowmelt due to heat convection from the air is one of the so-called "mass-transfer equations". It is based on the rather complex theory describing the turbulent transfer mechanism in fluids (see Tennekes and Lumley, 1972, for an introduction to turbulence theory and section 2.6 for some further discussion of the process). Fortunately, simple approximation of the complex terms derived from the theory of turbulence have been shown to be useful in practical hydrology (U.S.C.E., 1960). This simplified approach considers chiefly the temperature gradient from the snow surface to the overlying air, the wind speed, and the air density. A temperature gradient must exist if heat is to be transferred from the air to the snow. This heat transfer will reduce the temperature gradient and the air must be replaced (by wind)

for the convection to continue. The rate of turbulent transfer of energy is dependent on the eddy viscosity of the fluid which in turn is dependent on fluid density.

In the original U.S.C.E. (1956) equation the density term is represented by p/p_o, the ratio of air pressure at the snowpack to that at sea level, respectively. This ratio is rather conservative, however, and varies from unity at sea level to 0.7 at an elevation of a little more than 3000 m. For areas with moderate topographic relief we can take p/p_o as a constant of 0.8, and the convective melt is then given by

$$M_c = 0.183\ T_a\ V_n \tag{2.8}$$

where M_c is the snowmelt in mm/day and T_a and V_n are respectively the air temperature in ^oC at the 3-metre level, and wind speed in m/s at the 15-metre level (see Figure 2.4D).

Thus under typical melting conditions with mean daily air temperature of about 10^oC, and an average wind speed of 5 m/s, the snowmelt due to this factor would be about 9 mm.

When rain warmer than 0^oC falls on the snowpack surface, heat is transferred to the snow and causes melting. The amount of melting, however, is usually rather small. Although one kilogram of water liberates 4187 J of heat energy for every 1^oC fall in temperature, it requires 334,960 J or eighty times the former amount of heat to melt on kilogram of ice. Remembering that 1 kg of water has a volume of 1 mm x 1 m^2, the daily melt due to the heat of raindrops is given by

$$M_p = \frac{4187\ P_r T_w}{334\ 960} = 0.0125\ P_r\ T_w \tag{2.9}$$

where P_r and T_w are respectively the one day rainfall in mm, and the raindrop temperature (approximately equal to mean free air temperature) in ^oC (see Figure 2.4E). Thus is would take about 16 mm of one day rainfall at 5^oC to produce 1 mm of snowmelt. This may seem contrary to the common observation that the combination of rain and melting snow yields large amounts of runoff. However, the large amounts of runoff associated with these conditions is produced largely and directly by the rainfall which is unable to soak into the snowmelt-saturated snowpack and soil in the basin. Thus most of the rain forms runoff rather than entering storage.

Heat conducted to the snowpack from the ground is generally thought to be insignificant relative to the other factors which produce snowmelt. Clearly we could cite specific exceptions such as the high latitude geothermal area of Reykavik, Iceland, but it is one of just a few such cases. It is more usual that heat conducted to the snowpack from the ground averages an amount that would melt about 0.5 mm/day (U.S.C.E., 1960). Over the period of a long winter it might amount to several centimetres of melt (M_g), and could be a significant factor in ripening the snow and in adding moisture to the soil (Bruce and Clark, 1966).

The total melt, M, produced at a point location by heat from the six sources discussed above, is given by

$$M = M_{rs} + M_{rl} + M_e + M_c + M_p + M_g \tag{2.10}$$

Of course, in hydrologic problems the melt from an area, rather than that from a point source, is usually the primary concern. In the U.S.C.E. (1960) studies it was found that, given certain simplifications, the expressions for the melt components outlined above could be combined to yield basin-wide melt estimates. These estimates have been developed for two basin conditions: melt during rain periods, and melt during rain-free periods. The sets of equations describing snowmelt in these two conditions are modified by the application of empirical coefficients which reflect the influence of a variety of forest cover. Although it is beyond the scope of this background chapter to examine these various basin-melt equations in detail, it will be instructive to derive one of them to illustrate the types of assumptions on which they are based. For example, let us consider the often critical case of melt during rain in a fairly open drainage basin.

Insolation during periods of rain is commonly assumed to be constant and close to 1.675×10^6 J/m^2 for an open area; it may be smaller for a densely forested area. If the early melt-season albedo is taken as 65 per cent, then equation 2.3 yields (with B = 0.97)

$$M_{rs} = \frac{(1 - 0.65)(1.675 \times 10^6)}{(3.345 \times 10^5)(0.97)} = 1.81 \text{ mm} \tag{2.11}$$

Longwave radiation melt for rainy days is taken as that under complete cloud cover. Because cloud-base temperature is sensibly equal to that at normal instrument height, equation 2.5 can be modified to read

$$M_{rl} = 1.\ 326\ T_a \tag{2.12}$$

Equations 2.7 and 2.8 can be combined to yield an expression for condensation - convection melt, M_{ce}:

$$M_{ce} = V_w\ [0.0109\ (e_a - 611) + 0.183\ T_a] \tag{2.13}$$

The vapour pressure in equation 2.13 can be adequately represented by dewpoint temperatures. The dewpoint temperature, T_d, is the temperature to which the air must cool at constant pressure and water vapour in order to reach the point of water saturation. The relationship between saturation vapour pressure and dewpoint temperature (see Figure 2.5) is not linear but it can be assumed to be so for a small range of temperature (say 0°C to 15°C). Substitution of the vapour pressures with dewpoint temperatures in equation 2.13 (dewpoint temperature at the surface of the melting snow is 0°C) and adjustment of the coefficient to account for the change in units, yields

$$M_{ce} = V_w\ (01600\ T_d + 0.183\ T_a) \tag{2.14}$$

Condensation - convection melt during rain is assumed to occur

under saturated air, so that $T_a = T_d$. For these conditions equation 2.14 becomes

$$M_{ce} = 0.783\ T_a\ V_w \tag{2.15}$$

For basin melt estimates this equation is further modified by a basin condensation - convection coefficient, K, ranging from 1.0 for unforested plain to 0.30 for a heavily forested area

$$M_{ce} = 0.783\ K\ T_a\ V_w \tag{2.16}$$

Basin-wide snowmelt from rain is small and is expressed by equation 2.9 if it is assumed that the air is saturated and $T_w = T_a$. This amount together with a nominal 0.50 mm for ground-heat melt yields the final component in the basin melt equation:

$$M_{pg} = 0.0125\ P_r\ T_a + 0.50 \tag{2.17}$$

Grouping equations 2.11, 2.12, 2.16 and 2.17 yields the total basin snowmelt equation for open and partly forested areas (0 - 60 per cent cover):

$$M = T_a\ (0.783\ K\ V_w + 0.0125\ P_r + 1.326) + 2.31 \tag{2.18}$$

For a heavily forested area (60 - 100 per cent cover) it can be assumed that insolation is reduced and that wind effects are negligible; equation 2.18 accordingly can be modified to read

$$M = T_a\ (0.0125\ P_r + 2.057) + 0.50 \tag{2.19}$$

During rain-free periods, solar and terrestrial radiation become more important melt producing factors; also the degree of forest cover becomes very significant. Equations describing basin snowmelt during rain-free periods include empirical coefficients of proportionality to account for orientation and mean slope of the basin surface, and estimates of cloud and forest cover. If you are interested in the details of these equations, they are discussed in U.S.C.E. (1956, 1960) and outlined in most snow hydrology texts (for examples, see Garstka, 1964; Davar, 1970).

The generalised physical equations describing the snowmelt processes represent the most sophisticated approach to snowmelt-volume prediction. Often they are not as accurate for a particular basin, however, as a more direct empirical approach. This latter type of model involves correlating the observed snowmelt runoff with some temperature index or with a period length during which the temperature exceeds some specified magnitude (degree-day method). Examples of these models are described by Garstka (1964) and Davar (1970).

Quite apart from computational errors associated with some of the questionable internal assumptions of snowmelt models, the accuracy of all approaches to snowmelt prediction is limited by our inability to specify exactly the nature of the snowpack and how it will vary over

time. In this context the pattern of snowbank thinning and contraction is very important. For example there is a need for snowmelt models to reflect the fact that although snow tends to accumulate by elevation, it generally melts by aspect, south facing snowpacks retreating more rapidly than those on north facing slopes (in the northern hemisphere). Evaluations of four recently developed snow models is provided by Baker and Carder (1977). Problems of forecasting short-term runoff from snow and ice recently have been reviewed by Colbeck (1977).

2.4 Interception and surface storage

These two factors form a common pair of hydrologic controls in the sense that they are both mechanisms for storing water on or above the basin surface for some period of time.

2.4 (a) Interception storage

Rainfall is intercepted by vegetation and is redistributed as interception loss, throughflow, and stemflow. Interception loss is the water which is retained by leaves and later evaporated, throughflow is the water which drips through and from the leaves to the ground surface, and stemflow is water which trickles along twigs and branches, eventually to flow down the trunk to the ground.

When rain begins much of the water is caught in leaf and stem depressions or is simply held by surface tension forces as drops on leaf and bark surfaces and edges. Although evaporation rates under conditions of rain are very small (see Section 2.6) the wetted surface area of a single tree is considerable and that of a forest is quite sufficient to return as much as thirty per cent of the precipitation directly to the atmosphere (Kittredge, 1948). Continuation of rain at greater than the rate of evaporation (as is usually the case) results in the growth of drops stored on various parts of the vegetation until surface tension forces are overcome and the drops move downward from leaf to leaf, perhaps running together to form still larger drops. Eventually all parts of the tree will reach maximum water storage capacity (interception capacity) and further interception of rain will be offset by water moving out of the tree as throughfall and stemflow. Actually the throughflow and stemflow will always be less than the precipitation above the tree canopy by an amount equal to the interception loss by evaporation.

There are two main factors influencing the amount of interception loss from vegetation: interception capacity and meteorological conditions. Most of what we know about these factors is based on conditions in forests where measurements of the processes are more easily obtained than in herbaceous vegetation.

The interception capacity of vegetation is a function of the vegetation structure and morphology. These in turn depend on species composition, age, and density of the stands. For example, coniferous trees intercept more rainfall than deciduous trees in full leaf (see Table 2.1). Although the deciduous canopy is denser than that of the conifers, the broad leaf of the former promotes drop coalescence and flow whereas the separate needles of the decidious trees tend to retain individual raindrops and do not promote flow (Geiger, 1957). As

we would expect, unlike coniferous trees, deciduous species display a seasonal variation in interception capacity and loss in response to changes in leaf cover (see Table 2.1 and Johnson, 1942; Geiger, 1957; Law, 1957; Wisler and Brater, 1959; Penman, 1963; Zinke, 1967).

The total leaf area per unit ground area of grasses and herbs and of closed canopy forests are very similar (Lull, 1964). For this reason it has been suggested that their interception capacities and losses during the season of maximum development are also of a similar magnitude (Lull, 1964; Ward, 1967; Reynolds, 1967). However, it is clearly difficult to measure interception in this type of vegetation and the few attempts to do this have produced rather variable results (Horton, 1919; Clark, 1937; Lee, 1942; Beard, 1956; Burgy and Pomeroy, 1958; Kontorshchikov and Eremina, 1963; Lull, 1964).

The meteorological conditions influencing interception loss vary in importance with rainfall duration. For short period storms in which interception capacity is not filled, all of the precipitation may be evaporated. In such a case the interception loss is limited by water supply and will vary directly with precipitation. For long duration storms in which the interception capacity is filled by a rate of rainfall which at least equals that of the evaporative loss (the normal situation) water supply is no longer a limiting factor and interception loss is independent of the amount of rainfall being intercepted. In this case interception loss is at a maximum corresponding to the potential evaporation for the tree. Although we will consider the factors influencing evaporation at a later stage (in Section 2.6), we should for the moment note that during rain, often one of the most significant of these is wind activity. Wind can aid the evaporative process, as we saw in the discussion of snowmelt, by constantly replacing saturated air with some having a capacity to receive small amounts of water vapour. On the other hand, during a short duration storm,wind can considerably reduce the interception capacity and loss of a tree or shrub by agitating the foliage and causing premature throughfall. Thus, in general, wind will increase the total interception loss for a long duration storm and decrease it for a storm of short duration.

The total interception loss attributable to a given storm has been related to the interception capacity of the vegetation and the evaporation rate in the widely adopted equation proposed by Horton (1919):

$$I_{ri} = S_v + R_v E t_R \qquad (2.20)$$

in which I_{ri} is the total interception loss for the projected area of the canopy (in mm)

S_v is the interception capacity of the vegetation for the projected area of the canopy (in mm)

R_v is the ratio of the vegetal surface area to its projected area

E is the evaporation rate from the vegetal surface (in mm/hour)

and t_R is the duration of rainfall.

Forest Type	RAINFALL						SNOWFALL
	Gross Interception		Stemflow		Net Interception		Net Interception
	With leaves,%	Without leaves,%	With leaves,%	Without leaves,%	With leaves,%	Without leaves,%	Leaves on all species except hardwoods and aspen-birch
Northern hardwood	20	17	5	10	15	7	10
Aspen-birch	15	12	5	8	10	4	7
Spruce-fir	35	--	3	--	32	-	35
White pine	30	--	4	--	26	-	25
Hemlock	30	--	2	--	28	-	25
Red pine	32	--	3	--	29	-	30

Table 2.1: Rainfall and snowfall interception rates as a percentage of precipitation for several forest types (after Lull,1964)

A

Grain-size class	Infiltration capacity (mm/hr)
clays	0 - 4
silts	2 - 8
sands	3 -12

B

Material	Infiltration capacity (mm/hr)
clay loam	2.5 - 5.0
silt loam	7.5 -15.0
loam	12.5 -25.0
loamy sand	25.0 -50.0

Table 2.2: Variation in infiltration capacity with texture of sediments.

A: After Kirkby,1969

B: After Kohnke and Bertrand, 1959

Although equation 2.20 has been used in one form or another in many studies, its usefulness is limited because it only applies to individual rainstorms (not to longer periods such as daily or monthly averages) and to rainfalls which exceed the interception capacity. Linsley et al (1949) suggested that the equation be modified to include a term reflecting the exponential change in interception loss with precipitation (for the reasons related to limiting factors discussed above). Meriam (1960) accordingly proposed the following equation:

$$I_{ri} = S_v (1 - e^{-P/S_v}) + R_v E_{t_r} \qquad (2.21)$$

in which P is precipitation (in mm) and e is the base of the natural logarithms. As the ratio of precipitation to interception capacity increases, ln $(-P/S_v)$ approaches zero and equation 2.21 reduces to equation 2.20.

Purely empirical relationships such as some suggested by Horton (1919) may be used as an alternative to the above equations where the latter appear to be unsatisfactory. These formulae and others like them, however, include empirical constants which are based on very meagre experimental and field data. It is important to keep in mind that, given the current state of the art in this area of hydrology, it is not possible to calculate interception losses with a very high degree of accuracy.

2.4 (b) The measurement of interception loss

A far more reliable indicator of interception losses can be derived from direct measurement as a basis for solving the definitional equality:

$$I_{ri} = P - P_g - S_f \qquad (2.22)$$

where P_g is throughfall (precipitation under the canopy) and S_f is stemflow. Precipitation is measured with standard rain gauges (see Section 2.3 b) located above and below the canopy. Care must be taken, however, to ensure representative sampling by avoiding areas of unusually concentrated or deficient throughfall, the influence of local turbulence (particularly above and close to the canopy), etc. (see Geiger 1957; Law, 1957; Penman, 1963; Haupt, 1973).

It is practically impossible to measure stemflow for vegetation with numerous small stems. Streamflow can be measured in forest studies, however, by attaching and sealing collars to the circumference of the tree trunks. Stemflow enters the small opening in the top of the collar and flows through a pipe to a recording instrument such as a tipping bucket rain gauge. The collar works on the same principle as house gutters designed to collect runoff from the roof. A simple and effective method of stemflow measurement is described by Thompson (1964).

The main source of error in stemflow measurement is one of sampling. For example, Law (1957) found variation of 100 per cent between stemflow for trees in the same forested plot; there is clearly a need to obtain measurements from a number of trees (see also Kimmens, 1974).

Fortunately, for the sake of accurately solving equation 2.22, there is some evidence that stemflow is usually only a small proportion of the precipitation. Law's 1957 study suggested that the proportion is no more than 7 per cent and previous work by Geiger (1957) and a subsequent study by Brookes and Turner (1964) similarly estimated relative stemflow to be small at less than 5 per cent of the precipitation.

Studies by Rowe (1941), Kittredge (1948) and Lull (1964), however, indicate that the amount of stemflow largely depends on the roughness of the bark, with values ranging from 1 to 15 per cent and from 2 to 3 per cent of precipitation for respectively smooth and rough bark.

The relative stemflows measured by Freise (1936) in the rainforest of Brazil are considerably higher than the values cited above. His study indicates that almost half the amount of precipitation initially flows down the tree trunks and that stemflow actually reaching the ground is about 28 per cent of the precipitation. The difference between the two values results from equal amounts of abstraction by evaporation from, and absorption by, the bark.

The results of Freise (1936) appear to be anomalous but can hardly be rejected for this reason. Once again, as is so often the case in the relatively youthful earth sciences, there is a need for further research in order to reconcile the disparate results of too few studies.

The preceding discussion has been concerned with the interception of rainfall. We might expect that because snow is delivered to and stored in the vegetation canopy rather differently than is rain, differences will also occur in the amounts of interception. It has been argued (Geiger, 1957) that snow reaches the forest floor more easily than rain because the combination of its own weight and agitation by the wind causes snow to be quite unstable in the canopy. Also, conditions favouring snow accumulation in the canopy are those which promote very low evaporation rates. In any case, melting is far more likely than the more energy consuming evaporative process (see Section 2.3 d) in many areas.

Evidence concerning the interception of snow is meagre and plagued with discrepancies (see Jeffrey, 1970). For example, work by West and Knoerr (1959) in the Sierra Nevada suggests low snow interception losses (8 per cent of the winter fall), while later work by Miller (1962) in the same general area suggested a much greater loss (equivalent to 18 mm of daily snow evaporation (see also Miller, 1977). Other investigations have indicated that amounts of interception of snow and rain are sensibly equivalent (Johnson 1942; Rowe and Hendrix, 1951; Lull, 1964 - see Table 2.1). Lull (1964) also notes the expected contrast between the relatively high snow interception by coniferous trees and the relatively low amounts of snow intercepted by the leafless deciduous trees (see Table 2.1).

The inconsistencies in these data on snow interception undoubtedly result in large part from the difficulty of obtaining representative measurements. Snow interception must be measured in the very locations where drifting and variability of snow accumulation are most pronounced (Ward, 1967; Haupt, 1973).

2.4 (c) Surface storage

Closely allied to interception storage is surface storage. Any water which is stored on the surface of the ground for any length of time constitutes surface storage. Although it may be in the form of liquid water or snow, the latter has received considerable attention in Section 2.3 d and we will here confine the discussion to water.

Many authors also draw the distinction between short-term storage (detention) and long-term storage. Detention refers to the temporary storage of water on the ground surface wherever the intensity of rainfall exceeds the ability of the ground to absorb the water. In other words, it is the storage effect of overland flow. Detention storage is temporary in the sense that once rainfall intensity declines to less than the rate of infiltration into the soil, this storage component quickly disappears. Because it only exists during part of each storm event, evaporative loss from detention storage is usually very small. We shall see at a later stage (Section 2.7 a) that overland flow rarely occurs in many drainage basins; in these cases detention storage does not constitute a component of the hydrologic cycle.

Long-term storage (a few days or longer) usually represents a far larger abstraction of water from the basin bydrologic cycle than does detention. Most surface storage takes the form of standing bodies of water in depressions in the ground surface. It may occur in depressions ranging in size from a small shallow pool to a vast lake system such as those that exist in the Canadian prairies (see Stitchling and Blackwell, 1957). Before water can flow over the surface of a drainage basin the surface storage must be filled. In prolonged wet conditions the many depressions are filled to overflowing and are therefore interconnected so that water flows through the system of ponds to the main stream. In dry periods, however, only the larger depressions will contain water and general runoff from rain is delayed until the depressions are once again filled with water.

Water is lost from surface storage by infiltration and by evapotranspiration, processes which are the subjects of the next two sections.

2.5 Infiltration, soil moisture, and groundwater

As we noted earlier, infiltration is the term used to describe the downward movement of water from the surface of the ground into the soil below. This water eventually occupies one or more of the moisture zones shown in Figure 2.6A (see Ward, 1967). At the surface is the soil zone which first receives water from precipitation. From here water either returns to the atmosphere by evapotranspiration or it percolates downward towards the water table. After the water moves through this intermediate zone of downward percolation, it encounters the capillary fringe immediately above the surface (water table) of the saturated zone. In the capillary fringe, water is held above the water table against the force of gravity by surface tension effects (capillarity; see Kirkham, 1964). In well drained sites the water table and the capillary fringe are usually well below the ground surface. In poorly drained areas, such as a floodplain, however, the capillary fringe may reach to the ground surface completely displacing the intermediate zone. The location of the soil moisture zones are usually in a continual state of flux as they adjust to each new

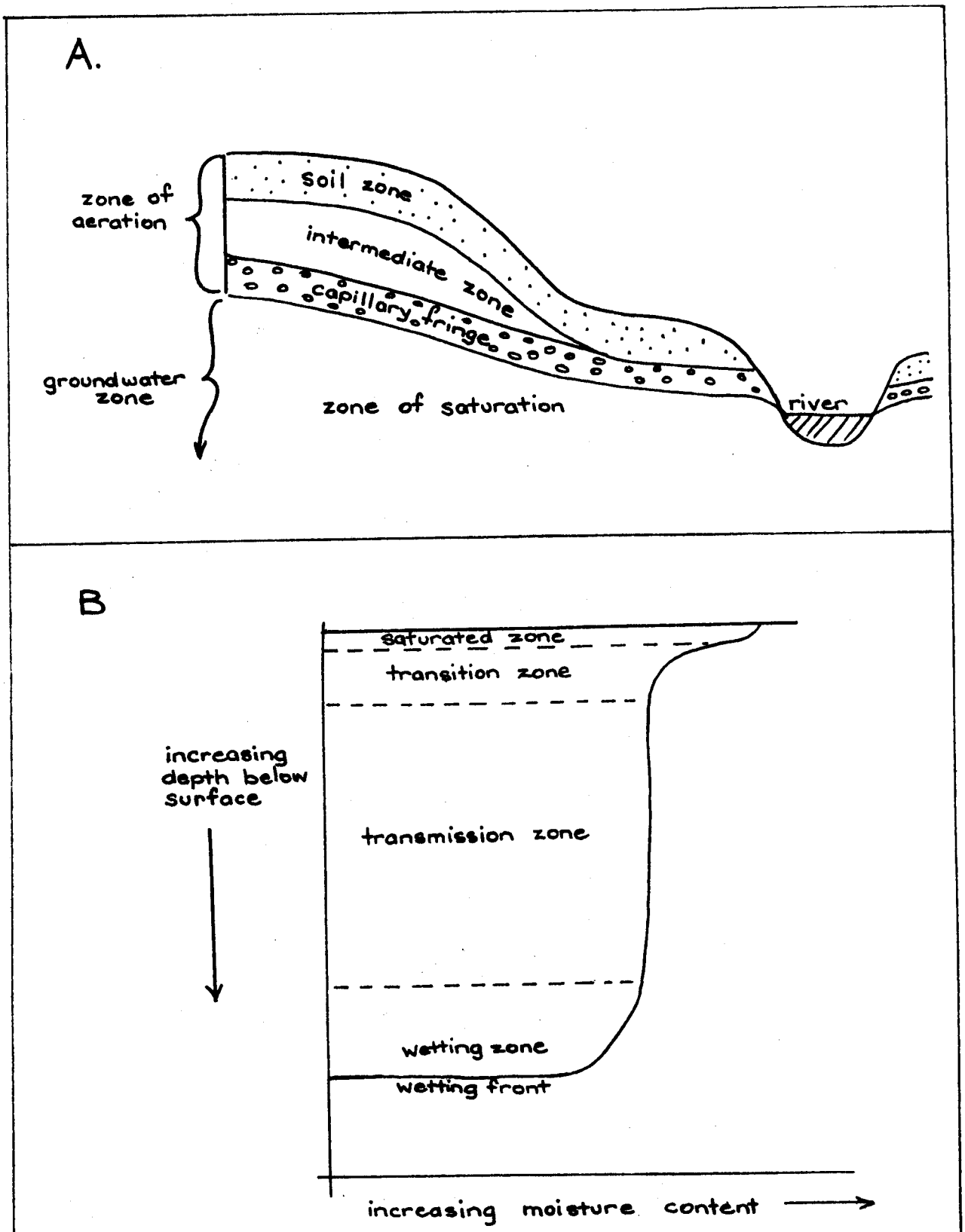


Figure 2.6: Soil moisture zones
A: General soil moisture zones
B: Moisture zones during infiltration (after Bodman and Colman, 1943)

rainfall and infiltration event. It is common practise to reserve the term soil moisture for the water stored in the unsaturated layers above the water table (the zone of aeration) and the term groundwater for that below the water table.

During infiltration water moves into the soil and intermediate zones as a wave of moisture. Early experimental studies (Bodeman and Coleman, 1943; Coleman and Bodeman, 1944) demonstrated the existence of several moisture zones associated with this wave (see Figure 2.6B). During infiltration the first centimetre or so of soil quickly comes to near saturation level. Immediately beneath this saturated soil is a strong negative moisture gradient across a few centimetres of soil depth, below which there is a transmission zone in which moisture levels change little with depth. The lower limit of the transmission zone is the wetting front - the front of the moisture wave. Provided infiltration occurs at a high enough rate or is sufficiently prolonged, the wetting front will eventually reach the water table and contribute moisture to the groundwater reserves.

The rate of infiltration, when less than the infiltration capacity, is largely dependent on rainfall intensity at the ground surface; as rainfall intensity increases so does the infiltration rate. This simple direct relationship is somewhat complicated by the fact that an increase in rainfall intensity is associated with an increase in raindrop size (Law and Parson, 1943) which can increase the surface soil compaction and thus reduce the infiltration capacity of the soil.

Rainfall can also indirectly affect the infiltration capacity of the soil by increasing soil moisture content, influencing the activity of earthworms and burrowing animals, and by altering the structure of the soil.

In general terms the infiltration capacity, f_c, is given by the Philips' (1957) infiltration equation:

$$f_c = A + B/\sqrt{t} \qquad (2.23)$$

in which A is the transmission constant of the soil, B is the diffusion constant of the soil and t is the elapsed time from the beginning of rainfall. The transmission term represents unimpeded flow through a continuous network of soil pore spaces and the diffusion term represents flow in small discrete steps from one pore to another. The latter type of random movement is essentially a response to the differential forces of capillarity that exist between wet and dry soils (see (Ward 1967 for a detailed discussion). The capillarity force gradient and the related potential for diffusion flow are greatest at the wetting front. Because infiltration occurs from above and the dry soil occurs at depth, the net movement of this diffusion flow is downward. It also follows that, in general, diffusion flow declines as soil moisture content increases and does not occur at all in saturated soil.

The infiltrating water thus has two components: a transmission component which is constant and represents a steady flow through the soil, and a diffusion component which is an initially rapid and then increasingly slow displacement of air from the soil pores. In actual soils these two types of water transfer occur together in all pores. A typical graph of infiltration capacity versus time is shown in Figure 2.7. The rapid initial decline in infiltration rate represents adjustments of the diffusion term to the increasing soil moisture conditions over time, and the infiltration asymptote represents the constant transmission term of equation 2.23.

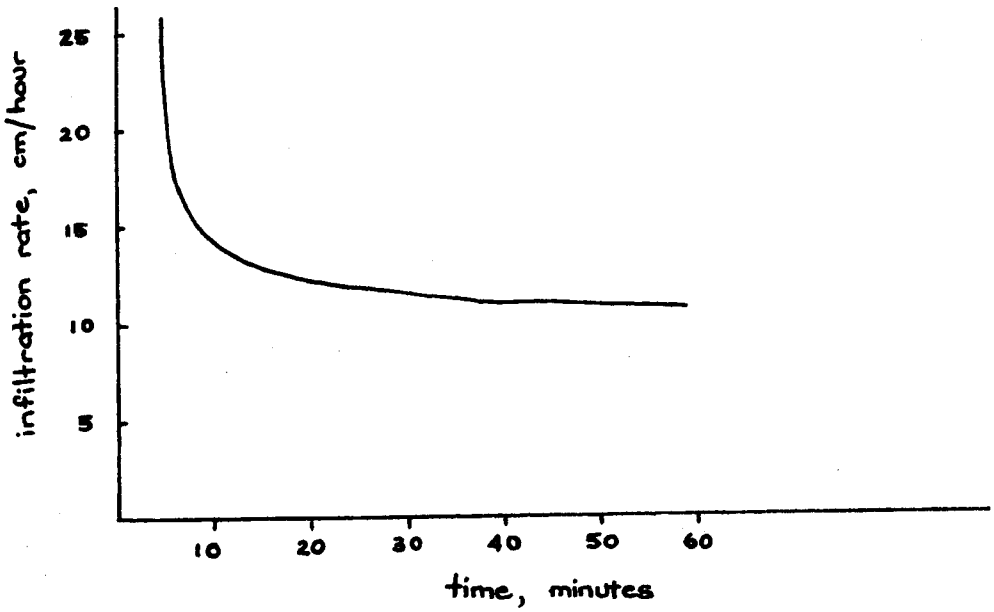
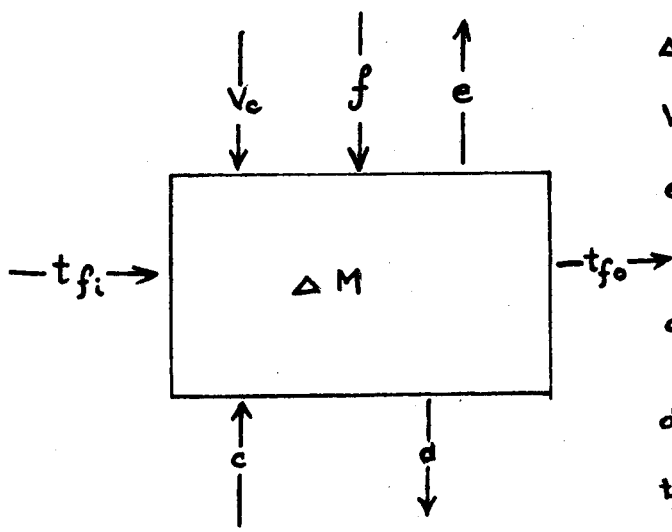


Figure 2·7: Infiltration capacity in relation to elapsed time After Spence (1972).



ΔM = rate of change in soil moisture content

V_c = rate of moisture addition from condensation

e = rate of evapotranspiration

c = rate of moisture addition by capillary rise from zone of saturation

d = rate of moisture loss to zone of saturation by percolation

t_f = the net rate of change in throughflow, where t_{fi} is the rate of in flow, and t_{fo} is the rate of outflow.

Figure 2·8: The components of the water balance in soils.

The rate of change in soil moisture is determined by the behaviour of the components of the soil water-balance equation (modified from Philip, 1964):

$$\Delta M = f + c - d - E + V_c + t_f \qquad (2.24)$$

the terms of which are defined in Figure 2.8. The condensation term is probably negligible in most cases (see Penman, 1940; Garr, 1953; Marshall, 1959). Equation 2.24 clearly describes a feedback system in which changes in soil moisture and infiltration rate are dependent on each other.

The magnitude of both the transmission and diffusion components in equation 2.23 is dependent on the type of material into which the water is infiltrating. The transmission constant tends to be greatest for material in which pores are large and interconnected; diffusion flow is most efficient through material with small interconnected pores. The net effects on infiltration rates of different types of material are shown in Table 2.2 and 2.3A. It is generally the case that infiltration capacities increase as grain size of the material increases. It is important to remember, however, that soil is a naturally layered substance usually with organic-rich light textured material overlaying leached medium textured soil which in turn is underlain by heavier textured soil at depth. Some soils may contain almost impermeable clay pans, others have caliche layers, still others contain a sort of plumbing system created by the rotting of plant roots and the movement of burrowing animals. Infiltration rates will accordingly be highly variable, even within the one soil type.

In many cases, however, it is not the soil but the soil surface which is the factor limiting infiltration capacity. For example, where vegetation cover is sparse, the impact of raindrops on the soil surface can cause appreciable soil compaction which will markedly reduce infiltration rates. Compaction of this type appears to be more effective in clay-rich rather than sandy soils (Wisler and Brater, 1959). A similar but more localised effect is caused by the hooves of grazing animals and the wheels of vehicles (Doreen and Henderson, 1953).

Another surface process which can reduce infiltration capacities in the inwashing of fine material into the surface pore spaces. The soil surface is essentially clogged by fine material, particularly after a long period of dry weather (Lowdermilk, 1930; Penman, 1963; Morin and Benyamini, 1977). On the other hand, dessication of clay-rich soils may promote the development of cracks in the soil which can markedly increase infiltration rates (Penman, 1963; Ward, 1967).

The presence of ice in or on the soil is also a factor which will obviously influence infiltration capacity (see Post and Dreibelbis, 1942; Willis et al, 1961; Larkin, 1962; Kuznik and Bezmenov, 1963; Gillies, 1968). In areas of permafrost the ice simply behaves as bedrock and infiltration rates are often negligible (see Table 2.3A); elsewhere ice lenses can locally reduce infiltration rates.

These types of surface influences are seen in the extreme in urban areas where roads, sidewalks, parking lots, buildings, etc., seal the soil surface and promote negligible infiltration and extremely high runoff during storms. The hydrologic cycle in these circumstances is so modified that its study has become a new sub-discipline known as urban hydrology (see Jens and McPherson, 1964).

In non-urban areas, most of the surface processes influencing infiltration are also modified, if vegetation is present. For example, raindrop impact is unimportant in such cases, and forest litter filters out fine material that otherwise might clog soil pores, (see Penman, 1963), and sun-crack development is also inhibited. Permafrost is particularly sensitive to the insulating effects of vegetation, (see Nanson and Beach, 1977). The general effect of vegetation cover is to increase the infiltration capacity of the soil (see Table 2.3). Although the degree of the effect varies among plant species, most evidence suggests that, within a specie, the infiltration capacity will increase as the density of the vegetation increases (for example, see Smith and Leopold, 1942; Woodward, 1943).

2.5a: The measurement of infiltration capacity

Infiltration capacity is measured by two basic types of instrument: flooding type infiltrometers and sprinkling infiltrometers. Flooding infiltrometers vary in specific design and size but most are either tubes or rings which are let into the soil surface and filled with water from a graduated burette at a rate (infiltration capacity) necessary to maintain some constant head (see Figure 2.9A). Tubes measure the infiltration rate through a column of soil while rings, because of their shallow depth, also reflect lateral movement of water into the soil.

The sprinkling infiltrometer is designed to sample a larger area than the flooding type, and to provide a closer analogue to actual rainfall. Basically it consists of a water-sprinkled plot of known size in which infiltration rate is taken as the difference between rate of water application and rate of runoff (see Figure 2.9B). The plots range in size from a few square centimetres in the case of small portable instruments to fixed installations with sprinkled areas of the order of $100m^2$.

Both types of instrument measure only relative infiltration rates (Musgrove and Holtan, 1964) and the sprinkling infiltrometers appear to be more representative of large areas of soil and yield results which are closer to actual absolute values than those indicated by the flooding type (Gregory and Walling, 1973).

Detailed descriptions of infiltrometers of various types are given by Sharp and Holtan (1940), Barnes and Costell (1957), Hermsmeier et al (1963), McQueen (1963), Musgrove and Holtan (1964), Swanson et al (1965), Selby (1970), and Gregory and Walling (1973).

2.5b: The measurement of soil moisture content

Techniques for the measurement of the amount of water held in

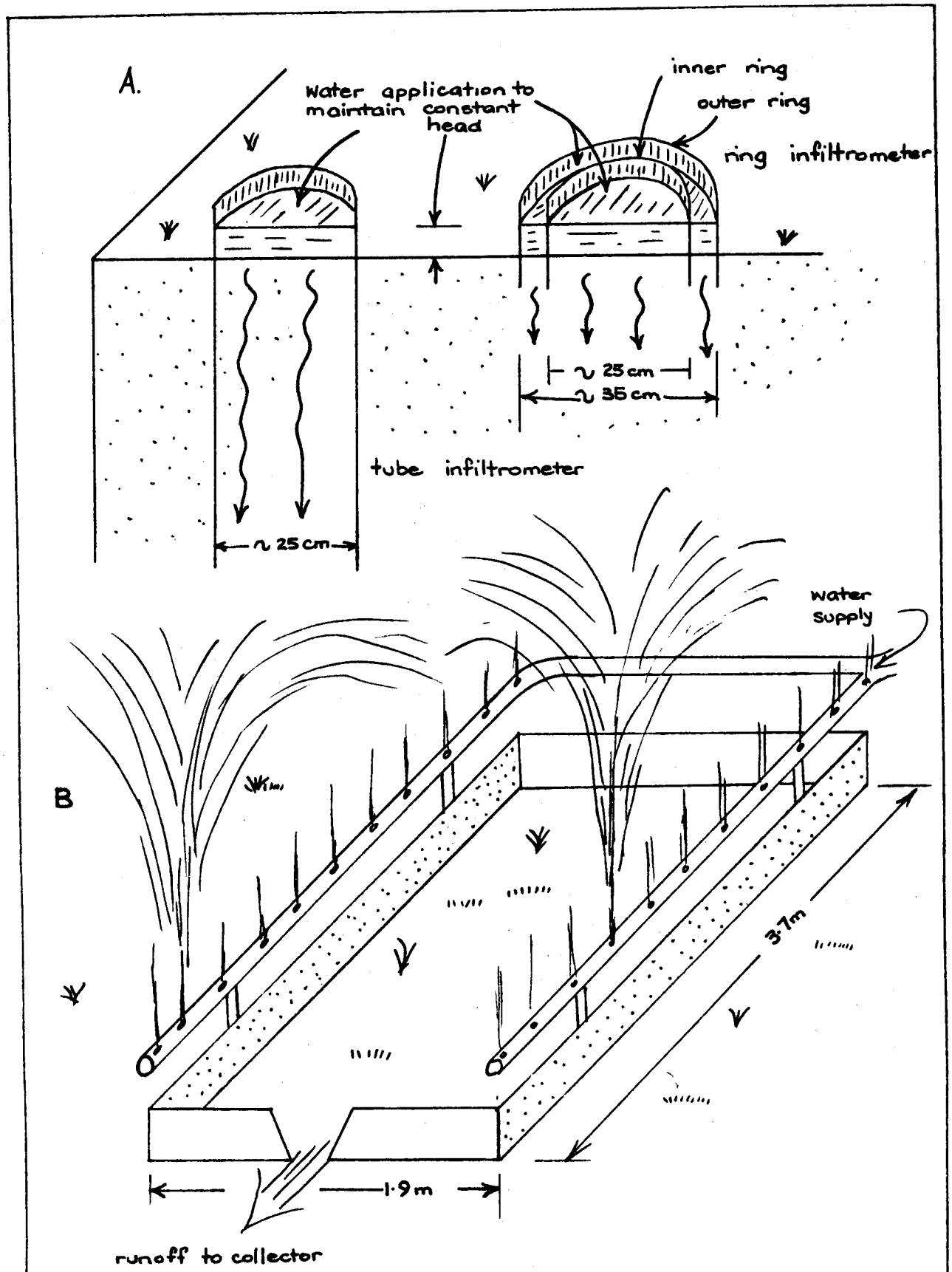


Figure 2·9: Instruments for measuring infiltration rates
A: tube and ring infiltrometers
B: a fixed sprinkling infiltrometer.

A

Soil profile characteristics	GROUND COVER CONDITIONS					
	Bare soil	Row crops	Poor pastures	Small grains	Good pastures	Forest
I	8	13	15	18	25	76
II	3	5	8	10	13	15
III	1	2	3	4	5	6
IV	0.5	0.5	0.5	0.5	0.5	0.5

Category I : Coarse and medium textured soils over sand and gravel (glacial outwash), coarse open till, or coarse alluvial deposits
II : Medium textured soils over medium textured tills
III : Medium and fine textured soils over fine textured clay-rich till
IV : Shallow soil (less than 3/5m.) over bedrock.

B

Ground Cover	Infiltration capacity, mm/hr
Old permanent pasture	61
Moderately grazed permanent pasture	20
Heavily grazed permanent pasture	15
Strip cropped or mixed cover	11
Weeds and grain	10
Clean tilled	9
Crusted bare ground	8

Table 2.3 A: Infiltration capacities (mm/hr) for a variety of soil and vegetation types (after Ayers, 1959)

B: Infiltration capacities for a variety of ground cover types on similar soils (after Holtan and Kirkpatrick, 1950)

a sample of soil are reviewed in most standard texts in related subjects (for examples, see Baver, 1956; Kirkham, 1964; Holmes et al, 1967; King, 1967; Gregory and Walling, 1973).

The gravimetric method, the most common and only direct way to measure soil moisture content (see Reynolds, 1970), involves weighing the soil sample before and after oven drying at 105°C. The weight difference is that of the water and the moisture content is usually expressed as a percentage weight of that for the even-dried soil. Although this method is direct and is used to calibrate other instruments, it is a time consuming analysis which must include precautions against errors resulting from soil moisture changes in the sample as it is moved from the field to the laboratory. Furthermore, each sample provides only a point measurement in time and space and the technique is thus unsuited to continuous monitoring of spatially integrated soil moisture changes over time.

Several methods have been developed to overcome these limitations by providing a continuous soil moisture record of an indirect nature in relatively undisturbed soils. The first of these, the electrical resistance method, is based on the principle that the electrical conductivity of the soil varies with soil moisture content. A pair of electrodes, imbedded in some porous material capable of rapidly coming to moisture equilibrium with the soil, when buried, provides a continuous record of electrical resistance, and by calibration, of soil moisture content.

Another indirect method of soil moisture determination is the neutron scattering technique (for example, see Stone et al, 1955; Van Bavel, 1965; Bell and McCulloch, 1966). Neutrons emitted from a radioactive source lowered into the soil are slowed as they collide with various elements including the hydrogen in soil water. The count of these slow neutrons can be calibrated to provide a continuous and spatially integrated (de Vries & Kring, 1961) record of soil moisture content. Although most instruments of this type are relatively expensive, their use may be justified in some cases by the appreciable labour saving of about 85 per cent over other methods (see Stone et al, 1961; Cohen & Tadmore, 1966).

Finally, soil moisture content can be measured by calibration with capillary pressure (also termed soil tension and soil suction) measured by a tensiometer. In its simplest form a tensiometer consists of a porous cup connected by a water column to a manometer or vacuum gauge. As soil moisture increases water flows into the cup and is drawn back out as soil moisture decreases. Changes in the "suction" of the soil, and thus in the soil moisture content, are recorded by the manometer. Simple tensiometers function best in moist conditions and more sophisticated instruments must be used in dry soils (see Peck and Rabbidge, 1966) although the basic principle remains the same.

2.5c: The measurement of some groundwater properties

The most common method of measuring the height of the water table, particularly in small drainage basins, is the direct observation of the water surface in bore holes and wells. The position

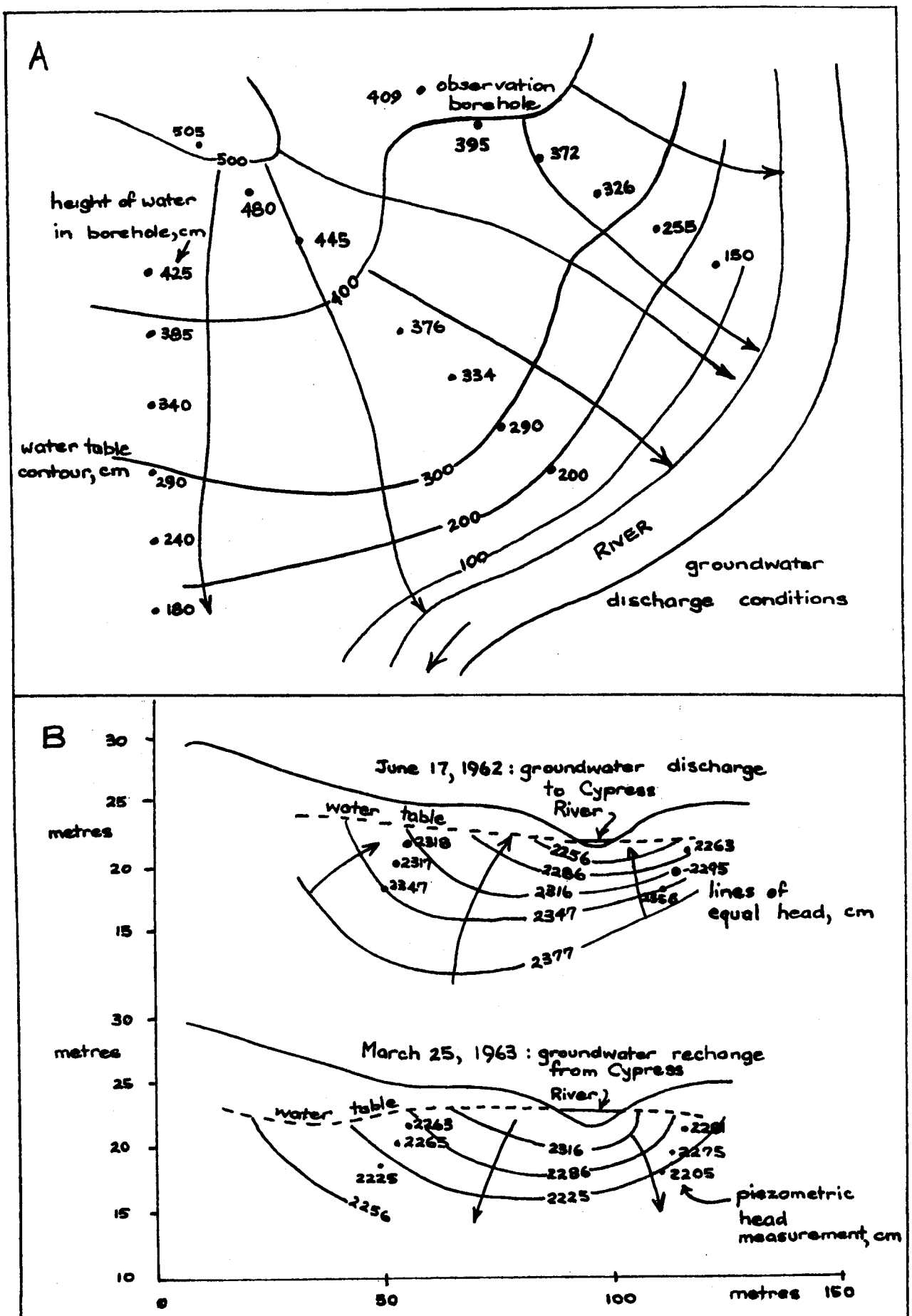


Figure 2.10: Groundwater flows in response to
A: a sloping groundwater surface, and
B: the piezometric gradient beneath a river (after Meyboom et al 1966)

of the water surface is simply measured manually with a graduated tape or continuously monitored by a float recorder such as that used to record river stage (see Section 2.7c).

Movement of water within the groundwater body is a three-dimensional phenomenon induced by pressure gradients existing there. In the water table zone the horizontal movement of water is indicated by the normals or perpendiculars through the contour pattern describing the groundwater surface (see Figure 2.10A). The vertical component of groundwater movement can similarly be depicted by normals through isolines of equal pressure (equipotentials) in a vertical section through the groundwater reservoir. Figure 2.10B shows two such examples illustrating that, during high flow, rivers tend to supply the groundwater reservoir with water from the channel, and that during low flows in the channel the groundwater flow reverses to nourish the river.

Point pressures within the groundwater body are measured by a piezometer, a small diameter tube open only at the point of measurement. Water in the piezometer tube thus rises, not to the level of the water table, but to a height (pressure head) consistent with local pressure at depth. These types of data may be used in a variety of flow equations describing the physics of flow through porous media to estimate the groundwater discharge per unit cross sectional area; one of these, the D'Arcy equation, will be considered in section 2.7(b).

Groundwater movement may also be assessed more directly by tracing the flow route of dyes and radioactive solutions from one borehole to another in the groundwater zone. Studies using tracer techniques are described by Halevy et al (1967) and Drew and Smith (1969).

Most aspects of groundwater hydrology receive a thorough review in the recent text by Bouwer (1978)

2.6: Evapotranspiration

Evapotranspiration is the collective term describing evaporation of water from wet surfaces and the process by which vegetation absorbs water through its root system and either uses it directly to build plant tissue or passes it back to the atmosphere principally by evaporation from leaf surfaces (transpiration).

The physical controls on evaporation were recognized by Dalton in 1802 and since then many theoretical and practical investigations conducted by scientists from a dozen disciplines have created a very substantial literature on the subject (for examples, see King, 1961; Sellers, 1965; Thornthwaite and Hare, 1965 for detailed accounts of the theory of evapotranspiration, and Veihmeyer, 1964, Ward, 1967; Barry, 1969 and Gray et al, 1970, for more general reviews).

We will not explore this literature in detail but it will be useful for us to examine some of the basic concepts involved, and to consider some representative research findings.

2.6 (a) The nature of evaporation from free water and soil surfaces

We have already encountered some of the concepts relevant to a discussion of evaporation in the analysis of the snowmelt process (section 2.3 (d)). The net transfer of water molecules into the air only occurs if a vapour pressure gradient exists between the evaporating surface and the air. As we noted earlier, it is strictly the partial pressure of the water vapour in the atmosphere that is of concern here. Evaporation involves a phase change and some source of energy must be available to provide the necessary latent heat of vapourisation. At the molecule scale this means increasing the kinetic energy of the water molecules to the point where some are sufficiently agitated to break the covalent bonding between the hydrogen and oxygen atoms, thus allowing the molecules to leave the water surface as water vapour. When a water molecule leaves the water surface the total kinetic energy of the water body deceases and we say that it has cooled. Of course, the process of phase change across the water/air interface is one of exchange; only if more molecules leave the water surface than join it from the air, does positive evaporation actually occur. When the numbers of molecules escaping from the water surface equal those falling back to the fluid, an equilibrium is reached between the pressure exerted by the escaping molecules and the pressure of the surrounding air; this equilibrium condition is known as saturation.

The source of energy to bring about these physical changes may be solar radiation, sensible heat from the atmosphere, ground, and vegetation, and from the evaporating water body itself. Generally, solar radiation is the principal direct source of energy for evaporation.

The nature of the dependence of evaporation rate (E_v) on the vapour pressures of the water body (e_w) and of the air (e_a) is often formally stated as Dalton's law, the general form of which is

$$E_v = C\,(e_w - e_a) \qquad (2.25)$$

where C is a coefficient of proportionality reflecting the influence on evaporation rates of factors other than vapour pressures. The principle factors determining the magnitudes of the vapour pressures (and thus of evaporation) are temperature, wind activity, atmospheric pressure, water quality, and the shape and size of the water body.

The dependence of vapour pressure on temperature is illustrated in Figure 2.5. From equation 2.25 it is clear that equal increases in the temperature of both air and water may not increase the rate of evaporation because it is proportional to the difference in vapour pressures. If sensible heat of the water body is providing the latent heat of evaporation then evaporation rate will decline to zero as the temperature of the water cools to that of the air.

Wind activity, as we noted in section 2.3 (d), is a very important factor influencing evaporation rates. Advection of air above the water body directly changes the vapour pressure by replacing moist air and it may also supply additional heat energy for evaporation. Of course, the converse of these possibilities may also be true. Many of the evaporation equations based on Dalton's law

include a term to account for the significant influence of wind on vapour pressures (see Table 2.4).

The effect of atmospheric pressure on evaporation rates is not well understood. In theory a decrease in barometric pressure (i.e., a reduction in the number of air molecules per unit volume of atmosphere) should allow an increase in the rate at which water molecules escape from the free water surface. Such changes in barometric pressure are associated with moving air masses and changes in the altitude of observation. In practice, however, barometric pressure changes are also accompanied by other meteorological changes (such as in temperature and vapour pressure) and the influence of pressure change alone is exceedingly difficult to isolate (for example, see Peck and Pfankuch, 1963).

The rate of evaporation from water surfaces exposed to the same atmospheric conditions may also vary in response to the quality of the water. For example, the vapour pressure of water declines as salinity increases. Early studies showed that evaporation from sea water with an average salinity of about 3.5 per cent, is some 2 to 3 per cent less than that from fresh water (Lee, 1927; Rohwer, 1933; Adams, 1934).

The final important control on vapour pressures, the size and shape of the water body, is not an independent factor but exerts its influence through temperature and wind effects. The seasonal temperature regime of a small shallow lake will normally approximate the seasonal air temperature regime so that water temperature will be at a maximum in the latter half of the summer and at a minimum in the latter half of the winter. In a large water body, however, the water temperature regime is dampened with respect to the air temperature regime because of the relatively large amounts of energy required to change the mean water temperature. During the summer the incoming radiation slowly heats the water to a considerable depth but its temperature remains below that of the overlying air and little energy is available for evaporation. During the winter when air temperatures decline below that of the water, heat stored in the water during the summer supplies the energy required to maintain relatively high evaporation rates (see Figure 2.11). An example of this heat storage effect was provided by Morton (1967) who calculated that mean annual evaporation from Lake Ontario is 813mm whereas that from the larger and deeper Lake Superior is only 546mm.

The shape of a water body, in conjunction with wind direction, can also influence evaporation rates. If the wind direction is along the axis of an elongated lake, for example, the air may be saturated before it moves over the water surface, thus reducing evaporation rates towards the leeward shore. The same wind blowing across the narrow dimension of the lake, however, would sustain high evaporation rates (oasis effect) over the whole lake surface (see Figure 2.12).

Water in soils occurs in films on grain surfaces and fills the interstices between them. It should not be surprising, therefore, that evaporation from soils is governed by the same meteorological factors as those considered above in relation to a free water surface.

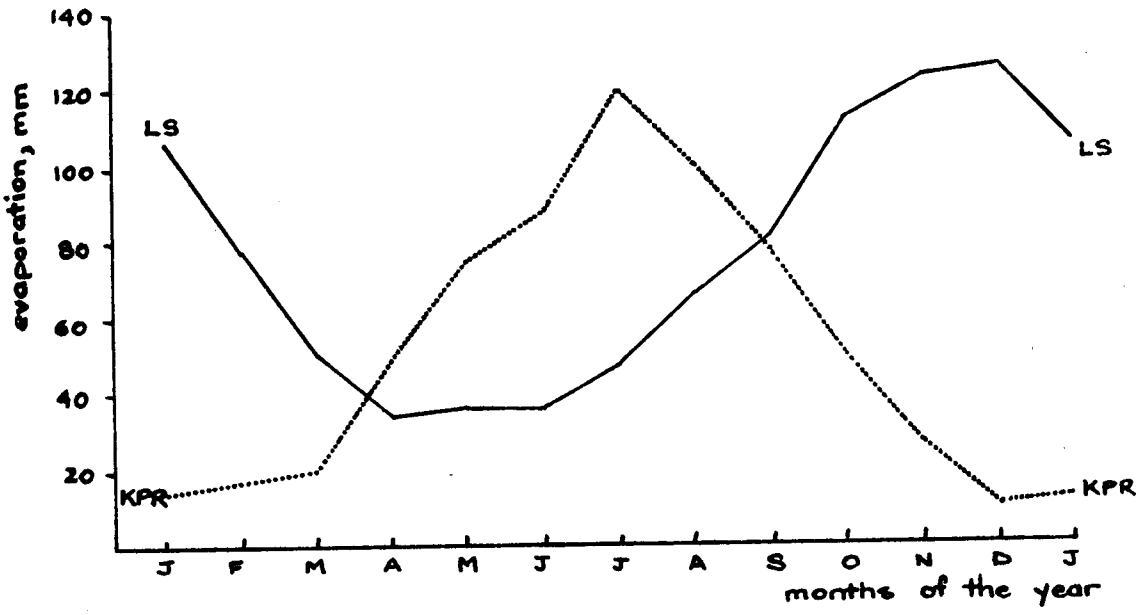


Figure 2·11: The seasonal march of evaporation from Kempton Park reservoir (KPR) near London (maximum depth, about 10m) and that from Lake Superior (LS: maximum depth about 400m). Adapted from an original diagram by Ward (1967).

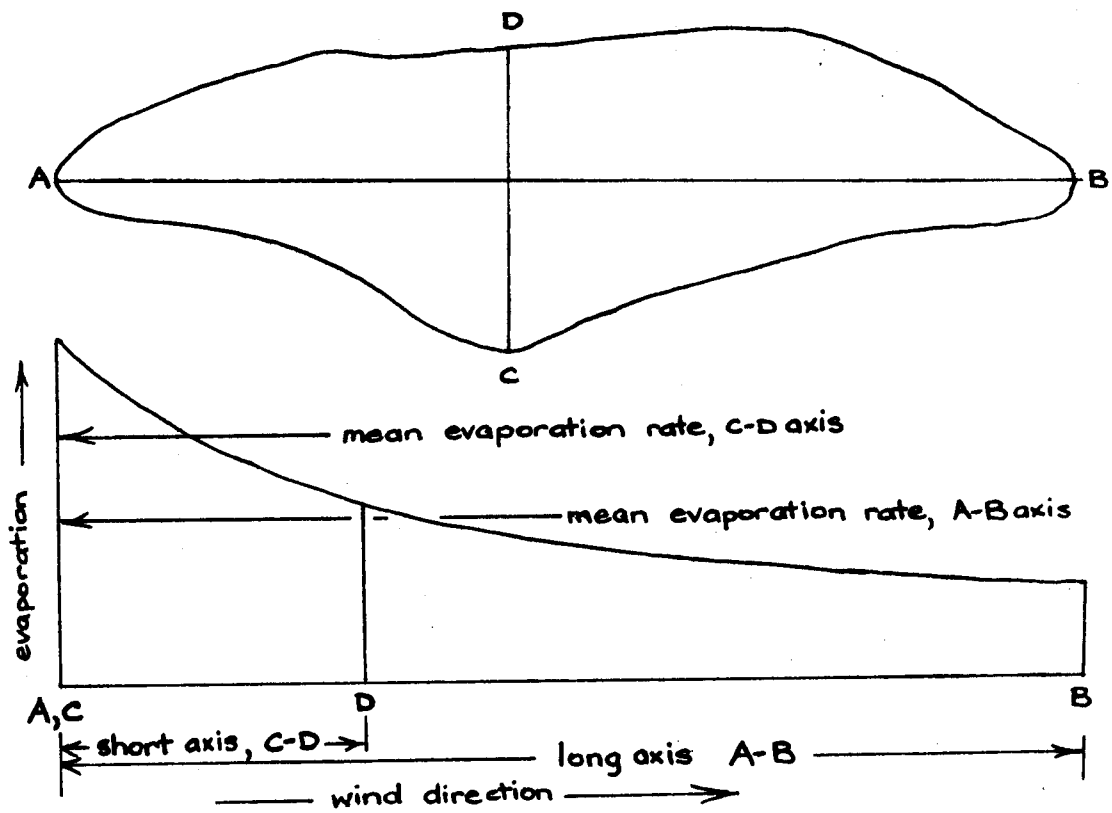


Figure 2·12: The influence of lake shape on the mean rate of evaporation

Source	Equation	
Dalton (1802)	$E_v = C(e_w - e_a)$	(2.25)
Fitzgerald (1886)	$E_v = \psi(e_w - e_a)$; $\psi = 0.4 + 0.199V_n$	(2.26)
Meyer (1915)	$E_v = C(e_w - e_a)\psi$; $\psi = 1 + 0.1V_n$	(2.27)
Horton (1917)	$E_v = 0.4(\psi e_w - e_a)$; $\psi = 2 - e^{-0.2V_n}$ For large areas, $E_v = E_v \times (1-P) + P\left[\frac{\psi-1}{\psi-h}\right]$	(2.28)
Rohwer (1931)	$E_v = 0.771(1.465-0.0186B)\psi(e_w-e_a)$ $\psi = 0.44 + 0.118\,V_n$	(2.29)
Lake Hefner (1954)	$E_v = 0.00177V_n(e_w - e_a)$	(2.30)
Lake Mead (1958)	$E_v = 0.001813V_n(e_w-e_a)t\left[1-0.03(T_a-T_w)\right]$	(2.31)

B = mean barometric pressure in inches of mercury at 32°F.
C = coefficient of proportionality; C = 15 for small shallow water, and C = 11 for large deep water in equation (2.27).
e = base of natural logarithms.
e_a = vapour pressure in air based on monthly mean air temperature and relative humidity at nearby stations for small bodies of shallow water, or based on information about 30 feet above the water surface for large bodies of deep water, in equation (2.27); or mean vapour pressure ofsaturated air at dew-point temperature, in mb, for equations (2.30) and (2.31), or in inches of mercury in other equations.
e_w = maximum vapour pressure (inches of mercury) corresponding to monthly mean air temperature at nearby stations for small bodies of shallow water, or corresponding to water temperature instead of air temperature for large bodies of deep water, in equation (2.27); or mean vapour pressure at the water surface in mb in equations (2.30) and (2.31), or in inches of mercury in other equations.
E_v = rate of evaporation in inches per 30-day month in equation (2.27), and in inches per t days in equation (2.31), or in inches per 24 hours in other equations.
h = relative humidity. P = fraction of time when wind is turbulent.
t = number of days in evaporation period. T_a = average air temperature, °C + 1.9°C
T_w = average water-surface temperature, °C. ψ = wind factor
V_n = monthly mean wind velocity, in miles/hr at 30 feet above the ground in equation (2.27), or mean wind velocity near the surface of the ground or water, in knots, in equation (2.31), oi in miles/hr in other equations.

Table 2.4: Some evaporation equations based on Dalton's Law (From Veihmeyer, 1964)

There is, however, a major difference between evaporation rates from free water-surfaces and soils: water supply. In the case of evaporation from a free water-surface, water is so plentiful that it never becomes a limiting factor. Evaporation rates from soils, however, are often less than the corresponding rates from free water surfaces because there is simply not enough water in the soil to allow the meteorological potential for evaporation to be realized. For this reason, evaporation from a non-limiting water supply such as a lake surface, is often termed the potential evaporation and soils usually have an evaporation opportunity which is less than 100 per cent (potential).

The most important factor directly affecting the evaporation opportunity of a soil is its moisture content. Early in this century the nature of the relationship between evaporation rates and soil moisture was well known (for example, see Whitney and Cameron, 1904); evaporation rate from a moist soil decreases rapidly at first as surface moisture content declines and rates thereafter decline slowly to zero (see Figure 2.13). The low rates of evaporation that occur following the initial drying of the first few centimetres of surface soil are basically limited by the rate of capillary rise of soil moisture from deeper layers within the soil. Keen (1927) was the first to note, however, that capillary lift contributes little to total evaporation losses after the water table falls to about 1 metre below the soil surface (35 cm in coarse sand and 85 cm in heavy loam).

These soil moisture/evaporation relationships will be modified by the presence of vegetation. Vegetation cover absorbs and radiates back to the atmosphere much of the incoming solar energy that otherwise would go to heat the soil. It also reduces air movement close to the ground and thus promotes high vapour pressures there. The generally reduced evaporation rates from soils under vegetative cover have been reported by many investigators (for examples, see Hursh, 1948; Rowe, 1955; Olivier, 1961). The insulating effects of forest-floor litter seems to be particularly effective in reducing evaporation from the soil (Baver, 1956).

Snow, for the purposes of our discussion, can be regarded as a special type of soil. Evaporation from snow, discussed in section 2.3 (d), is usually very small compared with that from a free water surface and from other types of soil. The amount of snow evaporation (sublimation) is small because of the typically low air temperatures close to the surface of the snowpack and because of the large energy expenditure associated with the phase change from solid to gas.

2.6 (b) The measurement of evaporation

Estimates of evaporation from open water-surfaces are usually obtained using evaporation pans. These are small water-filled tanks set on or into the ground, or fixed to rafts floating on the water surface. Evaporation is measured as the amount of water that must be added to the pan to maintain a constant depth while allowing for any input from precipitation. Evaporation pans come in a variety of sizes and specific designs but the most commonly used type is the 'Class A' pan of the United States Weather Bureau (see Figure 2.14A).

Because of the heat storage and oasis effects described in section 2.5A, pans overestimate evaporation from larger water bodies and measurements must be adjusted by applying a pan coefficient. Experience in the United States indicates that these coefficients range from 0.6 to 0.8 (see Gray et al, 1970). Pan measurements are best suited to estimating seasonal or annual evaporation from lakes and are much less reliable for monthly and shorter-period estimates (see McKay and Stichling, 1961).

Pan measurements provide no indication of evaporation from soil surfaces and measurements for this purpose must be obtained using lysimeters. A lysimeter is a vessel or container placed below the ground surface to intercept and collect water moving downward through the soil (Lull, 1964). They range in complexity from simple percolation gauges (Lapworth et al, 1948) to large sophisticated weighing devices (King et al, 1956; Pelton, 1961). Lysimeters have also been adapted to remotely monitor snowpack properties (see Thompson, De Vries and Amorocho, 1975).

The percolation gauge consists of a soil block enclosed by a metal container (Figure 2.14B). Evaporation from the soil surface is taken as the difference between precipitation measured by an on-site raingauge, and the amount of percolating water collected from the base of the metal enclosure. The instrument is best suited to long-term measurements because changes in storage within the soil block may render short-term measurements inaccurate. Some account of storage changes may be made by repeated measurements of soil moisture.

Weighing lysimeters are similar to percolation gauges but are designed to accurately monitor changes in soil moisture storage. This is accomplished by an installation capable of measuring the weight of the whole soil block over time. Perhaps the most famous of these is the installation at Coshocton in Ohio in which the lysimeters can weigh a 59,000 kg soil mass to within $\pm$ 2.3 kg (Harold and Dreibelbis, 1951). The mass of the soil in this case is determined hydraulically by a manometer recording water displacement. A similar installation at Hancock, Wisconsin, is shown in Figure 2.14C.

Other instruments have been devised to directly measure evaporation rates but they do not provide measurements of the same reliability as those obtained from evaporation pans and lysimeters. The most commonly used of these, the atmometer, is described in some detail by Veihmeyer (1964) and Shannon (1966).

2.6 (c) Transpiration: its character and measurement

Transpiration is the process by which the root systems of plants extract water from the soil and convey it through the xylem cells of the wood to the leaves and thence to the atmosphere. In a living plant there is a continuous column of water which extends from the root in the soil through the stem to the walls of the mesophyll cells forming the leaves. Covering the entire surface of a leaf is a single cell layer or epidermis which consists of many pores or stomata. It is through the stomata that almost all plant water

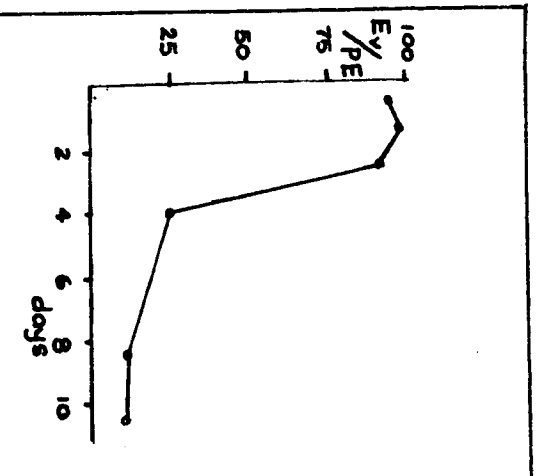
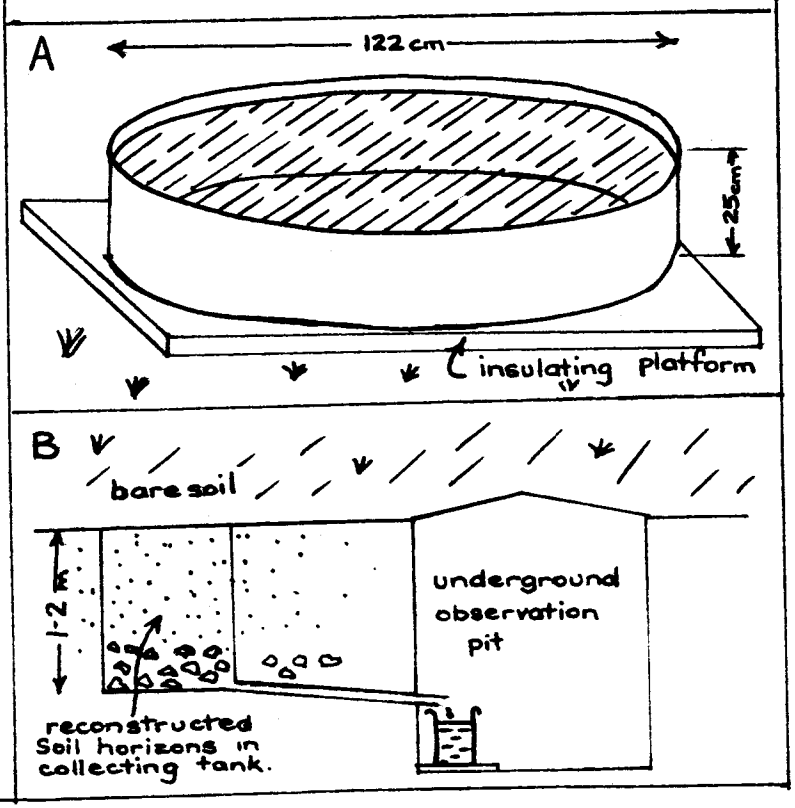


Figure 2.13: Evaporation/potential evaporation from initially saturated bare soil near Alice Springs, Australia, (data from Slatyer, 1961).



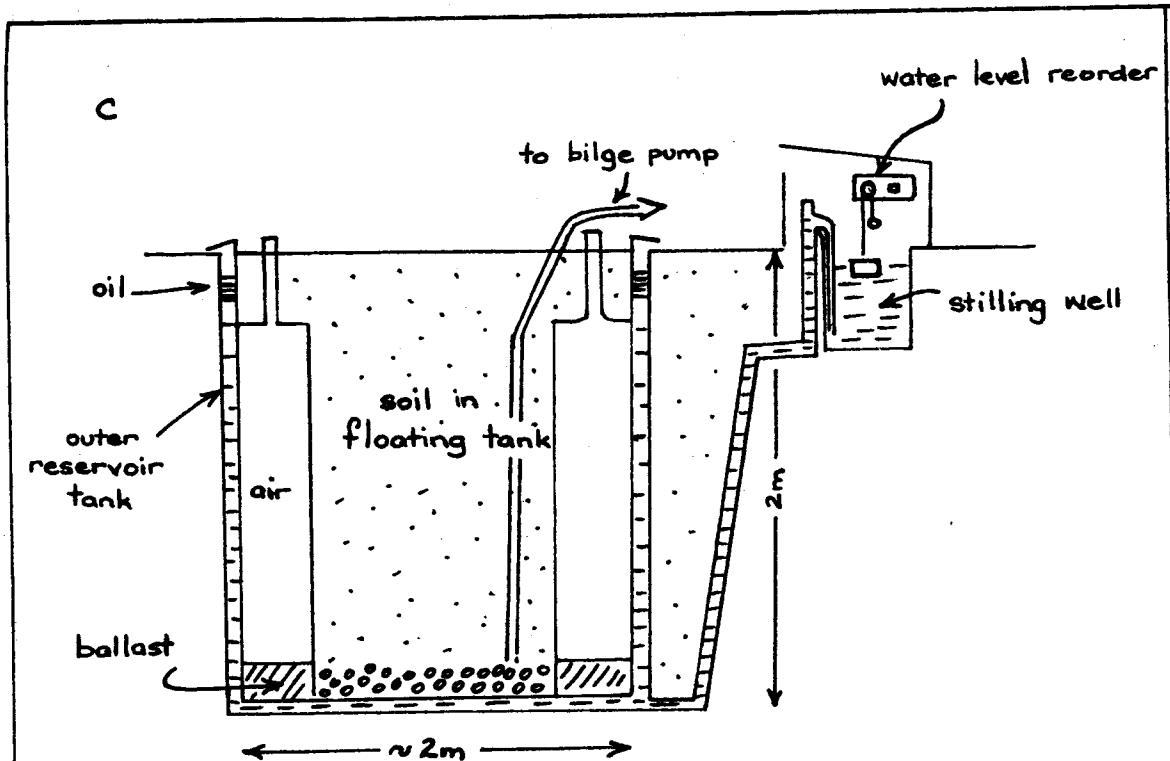


Figure 2·14: Installations for measuring evaporation and transpiration of water.

A: U.S. Weather Bureau Class A evaporation pan.
B: A percolation gauge
C: A lysimeter; changes in water mass in the floating tank is recorded hydraulically by displacement.

finally escapes to the atmosphere as water vapour. The stomata respond to light by opening during the day and closing at night; they are also responsive to capillary tension within the plant and will tend to close during times of moisture stress. Most transpiration of water (about 95 per cent) occurs during daylight hours (de Vries and van Duin, 1953, Monteith, 1956; Tanner, 1957). The transfer of water through the stomata to the atmosphere is the initiating process causing movement of water from the soil to and through the plant. It is a passive process as far as the plant is concerned but it serves the necessary function of nutrient and water supply, and it contributes in a small way to cooling leaf surfaces. In all but areas completely devoid of vegetation (a rare occurrence, even in so-called deserts), transpiration will be the more or less dominant component of the total water loss from the land surface (Ward, 1967).

Provided that leaf stomata are at least half open, and that available moisture exceeds the plant's water requirements, the meteorological factors affecting transpiration rates are the same as those governing evaporation. This is because movement of water through the stomata is an evaporative process and takes place when vapour pressure in the air is less than that in the spaces beneath the leaf stomata. We must remember, however, that the meteorological factors governing evaporation rates from vegetation relate to leaves and not to water and soil surfaces.

In addition to the meteorological controls on the rate of transpiration, there also may be some plant controls. It is perhaps surprising to some that there should be any doubt about this point. Research has led most investigators to conclude, however, that provided the water supply rate exceeds the limiting, transpiration rate set by meteorological conditions, the transpiration rate is largely independent of the type of vegetation (for example, see Thornthwaite, 1944; Russell, 1950; Hagan and Peterson, 1953; Thornthwaite and Hare, 1955; Green, 1956; Penman et al, 1956; Tanner, 1957).

The generally accepted view is that transpiration is basically a leakage process which occurs during the intake of carbon dioxide through the stomata (see Penman, 1963). Because regulation of the transpiration rate is physiologically unnecessary, no plant mechanism exists for that purpose. Thus, given adequate water supply, the rate of transpiration depends almost entirely on the rate at which water in the leaves can be converted to water vapour. This rate, in turn, depends only on meteorological controls.

We therefore might expect that, the greater the amount of solar radiation absorbed by the plant, the larger will be the transpiration rate. That is, in a large tree the total leaf area is greater than it is in a small plant and we might reasonably assume that a 50-metre tree will transpire more than a young 5-metre tree. It would seem, however, that this assumed difference only is true up to a point; increased shading and decreased ventilation is associated with most tree growth and these factors suppress transpiration. The shading of lower leaves by the outer canopy leaves can actually reduce the absorbed radiation of a large tree to less than that

absorbed by a younger one (Ward, 1967).

Because transpiration takes place principally from the leaves most directly in receipt of solar radiation, it has been argued that any area of vegetation which presents a continuous surface to the sun will transpire the same amount of water regardless of the type of vegetation (pasture, arable crops, forest, etc.) provided three main conditions are satisfied (Ward, 1967):

1. That the stomata opening regime is sensibly the same from one plant species to another.

2. That they absorb the same amount of solar radiation. If all species are presenting a flat leaf surface, the principal control on absorbed radiation will be the leaf albedo. Monteith (1959) found the albedo of most agricultural crops to be a constant 25 per cent. Although this figure may be as low as 15 per cent for coniferous forests (Penman, 1963), in general all green vegetation will absorb about the same proportion of shortwave radiation according to the traditional view.

3. That a sufficient supply of water is available at all times. If this were not so, deep rooted plants would absorb and transpire more soil moisture than shallow rooted plants (see Croft, 1950).

Botanical research in recent years, however, has focussed on inter-species differences in transpiration and has led many investigators to question the validity of the above generalisation. For example, Shepherd (1972) has reopened the debate on independent stomatal control of transpiration by some species. Brady et al (1974) have emphasised the importance of plant growth on transpiration rates; Szeicz et al (1969) have stressed the difference in the aerodynamic properties of tree canopies among species. Several forest microclimate models include structural parameters that vary from one species of tree to another (for example, see Waggoner and Reifsnyder, 1968; Waggoner et al, 1969). Miller (1977), drawing together these and other similar types of studies (principally the works of Sibbons, 1962; Konstantinov, 1968; Dilley and Shepherd, 1972; Rauner, 1972; McNaughton and Black, 1973), concluded that the concept of constant transpiration rates from all types of continuous vegetation cover is outmoded and should be replaced by a more complex model based on species variability.

We must remember, however, that progress in the natural sciences is often characterised by periods in which opposing viewpoints alternately dominate the research effort. It is through this type of research debate that we eventually gain a more balanced view of nature. But the process is usually a lengthy one and it would seem to be far from complete in this case. Perhaps this more balanced perspective may recognise that any generalisation about vegetation and transpiration rates will not be independent of scale. In large drainage basins measuring thousands of square kilometres and including a range of vegetation types, generalisations are more likely to apply than they are in a first order basin of just a few hectares in area.

Although it is known that plants have difficulty in extracting soil moisture when the supply is limited, the precise nature of the relationship between soil moisture content and transpiration rate is also a controversial issue yet to be resolved. The two basic views of the process are depicted in Figure 2.15. Veihmeyer (1927) and Veihmeyer and Hendrickson (1927, 1955) argued from experiments on potted prune trees that actual evapotranspiration (E_{vt}) occurs at the same rate as when moisture supply in the soil is unlimited (potential evapotranspiration, PE) over the range of soil moisture from field capacity to wilting point. Field capacity is the quantity of water normally held in the soil pores against the force of gravity. Once wilting point of the plant is reached E_{vt}/PE declines rapidly to zero. This view is also supported by Halkias et al (1955) and Gardner and Ehlig (1963), and in slightly modified form by Penman (1949) and Pearl et al (1954).

The opposing view is that evapotranspiration decreases as soil moisture declines. For example, Thornthwaite (1948, 1954) and Mather (1963) maintain that the decrease in E_{vt}/PE as soil moisture declines below field capacity, is a logarithmic function of soil suction (see Figure 2.15). A general decrease in E_{vt}/PE with decreasing soil moisture is also supported by Kramer (1952), Lassen et al (1952), Makkink and van Heemst (1956) and Visser (1963, 1964).

In time it will likely be shown that these two opposing views are probably end members of a continuous range of possibilities conditioned by soil type and climatic conditions. For example, field capacity ranges from 25 mm in shallow sandy soil to 550 mm in deep clay loams (Barry, 1969). Chang (1965) believes that Veihmeyer and Hendrickson's results (see Figure 2.15) apply to a heavy soil with vegetation cover in humid cloudy conditions whereas rapidly declining E_{vt}/PE might be expected in vegetated sandy soils in arid conditions (also see supporting work by Holmes, 1961).

It will be evident from much of the preceding discussion that transpiration, as a single hydrologic component, is very difficult to measure. Most estimates reported in the literature have been based on measurements of the rather artificial transpiration from phytometers. These are large vessels containing vegetation in a soil which has the surface sealed to prevent evaporative loss. The transpiration rate is measured as the water depth equivalent of the phytometer mass-change per unit time (Veihmeyer, 1964). Transpiration rates could similarly be determined by planting a weighing lysimeter with vegetation and sealing the surrounding soil surface. Such direct measurements of transpiration that are available (Meinzer and Stearns, 1929; Veihmeyer, 1927, 1938; Molcanov, 1955; Smirnov and Odinovka, 1954) indicate that transpiration rates are usually at least two to three times evaporation rates from the surrounding soil. It should be noted, however, that measurements of transpiration relative to evaporation rates, are highly variable, ranging over almost two orders of magnitude !

Fortunately, the fluvial geomorphologist is generally far more concerned with total evaporation from all types of surfaces (evapotranspiration) than he or she is with the contribution of particular components such as transpiration. Potential evapotranspiration is

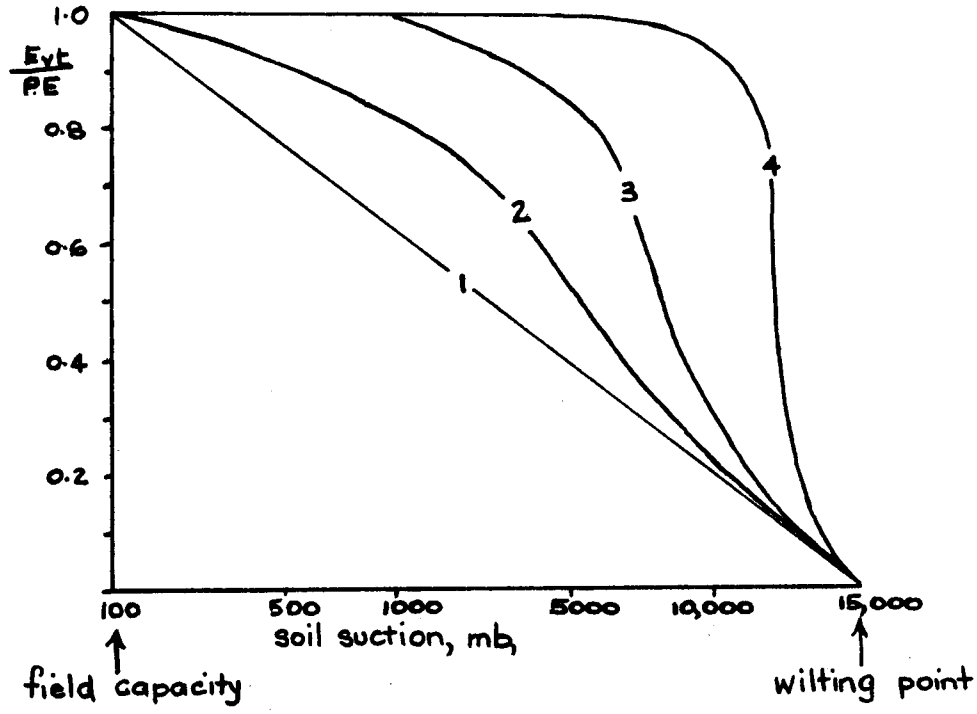


Figure 2·15: Relative evapotranspiration related to soil moisture according to
(1) Thornthwaite and Mather (1954),
(2)+(3) Schematic curves for vegetated sandy soil under high evaporation stress and vegetated clay-loam under low evaporation stress, respectively (after Barry, 1969), and
(4) Veihmeyer and Hendrickson (1927).

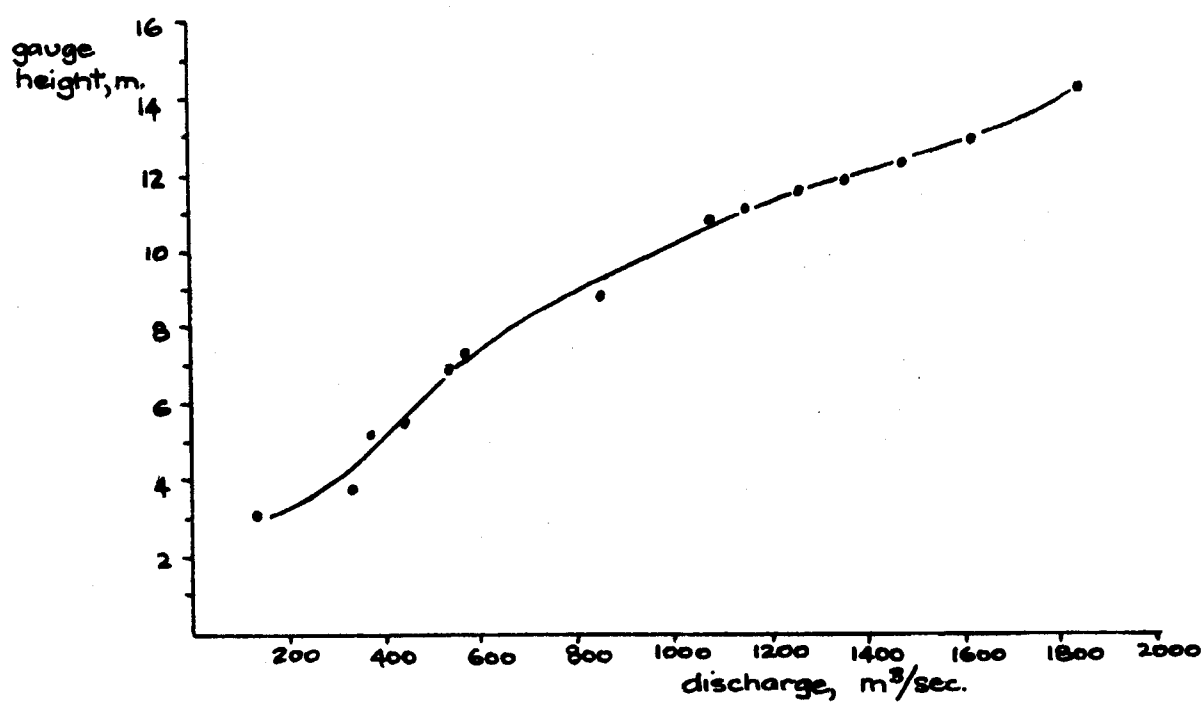


Figure 2·16: The discharge rating curve for the Colo River at Upper Colo, NSW, Australia.

usually measured as some function of loss from evaporation pans (for example, see Hargreaves, 1966), and actual evapotranspiration can be measured using lysimeters in which soil and vegetation conditions are matched as closely as possible with those in the surrounding area.

Less direct methods of estimating evapotranspiration include the monitoring of water tables and the analysis of the basin water balance. The first method involves computations of daily evapotranspiration on the basis of diurnal fluctuation in the water table observed in bore holes. Early successes were claimed for this method (Blaney et al, 1933; White, 1932; Gatewood et al, 1950) but it has seen little recent use. The second method involves the solution of equation 2.1 by assuming that the storage components remain constant (net groundwater storage is commonly measured from observation-well data). Evapotranspiration is essentially taken as the difference between mean precipitation input and basin outflow. In the short-term this method is generally unsatisfactory because of difficulties in determining the flow quantities to the necessary level of accuracy. It does appear, however, to provide reasonable estimates of long-term evapotranspiration (see Sanderson, et al, 1967) and is apparently reliable in the short-term if it is applied to unusually watertight basins (Edwards and Rodda, 1970).

2.6 (d) The estimation of evapotranspiration

In most basin hydrologic studies evapotranspiration is not measured but rather is estimated using relationships among environmental parameters established elsewhere. Almost two centuries of research since the formulation of Dalton's Law have yielded a bewildering array of equations designed to predict evaporative loss from natural surfaces. They range from simple expressions of the Dalton type (see Table 2.4) to complex simulation models based on large numbers of variables. In spite of the time and effort spent on formulating and modifying these predictive tools, however, an accurate and general equation is yet to be developed. This fact should come as no surprise because, as we have seen, the controls on evapotranspiration are exceedingly complicated and may always defy complete understanding. As it stands, almost all of the equations predict potential rather than actual evapotranspiration, thus avoiding the difficult problem of accounting for soil moisture and plant controls. Ward (1971) has reviewed many of the relevant formulae and the discussion here will simply focus on several of the more popular ones (see Table 2.5) in order to illustrate the general approaches to estimating evapotranspiration. These selected equations, like those in Table 2.4 are shown in their original or commonly used forms, thus they are not expressed in units of the S.I. system of measurement.

Equations for predicting evapotranspiration rates can be divided into two basic types: those that reflect essentially physical models and those that are essentially empirical or statistical statements. Of course, no equation is purely an expression of a physical model (no empirical component) nor is any one

a purely statistical statement; each includes elements of the other. There are in turn two types of approach to estimating evapotranspiration through physical models: the mass transfer approach, and the energy budget method.

The mass transfer approach is essentially an attempt to evaluate the constant in equation 2.25. It is based on the fact that the wind-velocity profile above the ground reflects in part the resistance offered to air movement by surface roughness and by turbulence near the ground. Turbulence in any fluid can be envisaged as an upward movement of discrete eddies generated at the boundary and carrying with them their properties including water vapour content. A turbulent eddy leaves the boundary with an initial quantity of kinetic energy which decays until it is exhausted and the eddy ceases to exist. This rate of decay, or rate of eddy diffusion as it is termed, determines the rate at which water vapour can be transferred from an evaporating surface to the atmosphere. It can be related to the wind velocity gradient above the ground (velocity varies with the logarithm of height in the turbulent layer of air) through the so-called von Karman constant (see Part III for a more detailed discussion of these fluid mechanics principles).

Two of the most widely used equations based on these mass transfer ideas are those proposed by Thornthwaite and Holzman (1939) and Sverdrup (1946), common forms of which are shown in Table 2.5.

The usefulness of equations (2.32), (2.33), and others like them (see Table 2.4), is limited for two reasons. First, they are based on a very simple model of turbulence which often may not describe the poorly understood actual eddy diffusion process. For example, the von Karman "constant", usually assigned a value of 0.4, may vary considerably from this value. From the structure of equations (2.32) and (2.33) we can see that errors in predicted evapotranspiration are directly proportional to the square of deviations from $k = 0.4$. At present there are no practical means of independently estimating the magnitude of k.

The second limitation is that evapotranspiration predicted by these equations is extremely sensitive to changes in the velocity and vapour pressure gradients. Measurements of these factors need to be very accurate if the results are to be meaningful. The ideal measurement facility includes instruments which will provide continuous and simultaneous readings of the relevant atmospheric properties.

It follows that mass transfer equations of this type are not in general use, but they are used to provide independent estimates of evapotranspiration for research purposes. They have most commonly been used to provide estimates of evaporation from lake surfaces (for example, see Phillips, 1978).

The second type of approach to estimating evaptranspiration through physical models, the energy budget method, earlier in this chapter saw application to the prediction of snowmelt yields

(see equation 2.2). Conceptually, it is far simpler than the previous approach although it does present some difficult problems of measurement. It follows from the energy conservation principle that the net total of shortwave and longwave radiation received at an evaporating surface (R_n) is available for three processes: the transfer of sensible heat (R_n) and of latent heat of evaporation (LE_v) to the atmosphere, and of sensible heat into the ground (H_g), or

$$R_n = R_h + LE_v + H_g \qquad (2.37)$$

Or solving for E_v (= PE),

$$E_v = \frac{R_n - R_h - H_g}{L} \qquad (2.38)$$

Equation (2.38) assumes that minor energy transfers such as those related to chemical and biological processes are negligible in this context. Although R_n can be easily measured with a net radiometer and H_g can be calculated from the soil temperature profile, it is not possible to directly determine R_h. To evaluate this component, use is made of the ratio of energy utilised by evaporative processes to that energy conducted to and from the evaporating surface by the air as sensible heat. In 1926 Bowen related this ratio to easily measurable quantities as follows:

$$B = \frac{R_h}{LE_v} = C \left[\frac{T_o - T_a}{e_o - e_a}\right] \frac{P}{1000} \qquad (2.39)$$

in which B = Bowen's ratio

T_o, T_a = respective temperatures of the evaporating surface and of the air (°C)

e_o = saturation vapour pressure (mb) corresponding to temperature, T_o.

e_a = vapour pressure of the air (mb)

& P = atmospheric pressure (mb)

& C = a constant with an average value of 0.61 under normal atmospheric conditions.

The assumption of equation (2.39), that the vertical transfer of heat and water vapour by turbulence occurs with equal efficiency (equal eddy diffusivities), has been shown by Dyer (1967) to be valid.

Combining equations (2.38) and (2.39) yields the energy budget equation (2.34) in Table 2.5. Of course, the energy budget method can also be used to determine evaporation rates from water bodies. Indeed, it was in this context that the idea was first applied by Schmidt in 1915. In the case of lake studies care must be taken to account for heat advection by inflow and outflow of water and the method should not be applied where net heat transfer to the lake bed is unknown (probably negligible in the case of deep lakes). A study by McKay (1962) & McKay and Stichling (1961) of the Weyburn Reservoir in Southern Saskatchewan provides a useful comparison of a variety of lake evaporation estimates (also see Bruce and Rodgers, 1962, for a summary of studies on the Great Lakes system). A recent example of the energy budget approach to evapotranspiration from a forest environment is provided by Stewart (1977).

Some of the measurement difficulties noted in the preceding

Source	Equation	
Thornthwaite & Holzman (1939)	$PE = \dfrac{62.3\rho k^2(V_{n8} - V_{n2})(e_2 - e_8)}{(\ln 800/Z_o)^2}$	(2.32)
Sverdrup (1946)	$PE = \dfrac{62.3\rho k^2 V_{n8}(e_o - e_8)}{(\ln 800/200)^2}$	(2.33)
Energy budget (see text)	$PE = \dfrac{R_n - H_g}{L(1 + B)}$ where $B = \dfrac{\Delta}{\gamma}$	(2.34)
Penman (1948)	$PE = (B\frac{R_n}{L} + E_a)/(B + 1)$	(2.35)
Thornthwaite (1948)	$PE = 16b\left[\dfrac{10T_m}{I}\right]^a$	(2.36)

PE = potential evapotranspiration in mm/sec (2.32;2.33); mm/unit time (2.34); mm/day (2.35); and mm/month (2.36)
= density of the air (gm/cc)
k = von Karman constant (0.4)
V_{n2}, V_{n8} = respective wind speeds at 2 and 8 metres above the evaporating surface (ev.s.) (cm/sec)
e_o, e_2, e_8 = respective vapour pressures at the surface, and at 2 and 8 metres above the ev.s.(mb).
P = atmospheric pressure (mb)
R_n = net radiation at the ev.s. (J/m^2 or more usually, $cal./cm^2$)
H_g = sensible heat transfer to the ev.s. (J/m^2 or more usually, $cal./cm^2$)
L = latent heat of evaporation (about 2.47×10^6 J/m^3, or more usually, 590 $cal./cm^2$)
B = Bowen's ratio in which γ = 0.27 (mm mercury/^oF), the psychometric constant; and $\Delta \doteq (de_s/dT)$, the change of saturation pressure with mean air temperature (mm of mercury/^oF)
E_a = the mass transfer term in equations 2.41 and 2.42
b = daylength correction factor
T_m = mean monthly temperature (^oC)
I = Thornthwaite's heat index (see equation 2.46)
a = a cubic function of I (see equation 2.45)

Table 2.5: Some evapotranspiration formulae

discussion can be overcome by combining the two physical models of evapotranspiration. By far the most widely adopted of these combination methods was developed by Penman in 1948 (subsequently presented with minor modifications in 1952, 1954, 1956, and 1963). He expressed potential evapotranspiration as a function of net radiation (R_n) and a mass transfer term (E_a) combining saturation deficit and wind speed:

$$R_n = 0.75 \quad R_s - R_b \tag{2.40}$$

where R_s is shortwave radiation at the evaporating surface and R_b is the net longwave radiation; the constant is an absorption coefficient corresponding, in this case, to a short grass albedo of 25 per cent;

$$E_a = f(V_n)\ (e_s - e) \tag{2.41}$$

in which $f\ (V_n) = 0.35\ (1+0.01\ V_n)$ for short grass (2.42)

and V_n = wind speed at 2m above the ground (miles/day)

e_s = saturation vapour pressure (mm of mercury) at mean air temperature

and e = actual vapour pressure at mean air temperature and humidity

Penman assumed that the net heat flux to the soil is negligible and that net radiation is simply divided between heating the air and providing latent heat for evaporation. Combining equations 2.40, 2.41, and 2.42 yields the Penman equation (for short grass cover shown in Table 2.5.

Penman also suggested two empirical equations for estimating the radiation terms in equation 2.40 (expressed in evaporative units of mm/day):

$$R_s/L = (1 - r)R_a(0.18 + 0.55n/N) \tag{2.43}$$

and $$R_b/L = \sigma T_a^4(0.56-0.09\sqrt{e})(0.10+0.90n/N) \tag{2.44}$$

in which r = albedo (as a fraction); set at 0.25 in equation 2.40

R_a = the theoretical radiation intensity at the ground surface in the absence of an atmosphere (expressed in evaporation units).

n/N = the ratio of actual to possible hours of bright sunshine.

and e = actual vapour pressure at mean air temperature and humidity.

The term σT_a^4 is the theoretical longwave radiation leaving the area in the absence of an atmosphere, in which σ is Stefan's constant (to account for atmospheric absorption of radiation) and T_a is mean air temperature.

The difference in the estimates yielded by equations 2.43 and 2.44 is the net radiation expressed in evaporative units (R_n/L in equation 2.35). Where it is possible, however, it is clearly more desirable to use net radiometer measurements in equation 2.35.

Nevertheless, direct measurements are often unavailable and it becomes necessary to use these indirect estimates. On the other hand, solving Penman's net radiation equations requires that we know the station latitude, time of year, duration of bright sunshine, mean air temperature, mean vapour pressure, and mean daily wind speed ! It is obvious that it will often not be possible to apply the Penman method because the required data are not available.

The accuracy of equation 2.35 appears to vary directly with the length of the period for which evapotranspiration is being estimated (Ward, 1967). Milthorpe (1960) suggests that, unless the radiation terms are measured directly, equation 2.35 probably yields results which are only meaningful for periods longer than one week. Penman (1963) acknowledged that, because of the statistical nature of the relationship between the relative duration of bright sunshine (n/N) and the total amount of incoming radiation, his formula should not be used to determine short-term evapotranspiration. On the other hand, if direct net radiation measurements are used in equation 2.35, then even short-term estimates of PE should be tolerably accurate (Ward, 1967).

Finally, we should remember that, although the framework of the Penman equations is theoretically elegant, the accuracy of the model will be limited by the weakest of the several included empirical relationships.

The second basic type of equation used to predict PE, the essentially empirical or statistical statements, is designed to utilise data readily available from almost all meteorological stations. For this reason, the most popular of these equations, those proposed by Thornthwaite (1944, 1948, 1953, 1954) constitute the most widely used of any method to predict potential evapotranspiration rates.

Thornthwaite related the consumptive use of irrigation water to air temperature in the western United States, allowing for the influence of daylength variation. The basic formula, equation 2.36 in Table 2.5, expresses evapotranspiration as a simple power function of mean monthly temperature. The constant b is the correction factor to account for daylength variation between months and the exponent a is evaluated in terms of an annual heat index, I, as

$$a = (67.5 \times 10^{-8})I^3 - (77.1 \times 10^{-6})I^2 + 0.0179I + 0.492 \quad (2.45)$$

in which

$$I = \sum_{m=1}^{12} (T_m/5)^{1.51} \quad (2.46)$$

If each 30-day month has 12 hours of sunshine, equation 2.36 reduces to

$$PE = 16.2(10T_m/I) \quad (2.47)$$

From our examination of the evapotranspiration process in section 2.6 (c) it will be obvious that the Thornthwaite approach is a gross simplification which should not be expected to provide in all cases accurate estimates of short-term PE. Its primary purpose is to provide a first estimate based on readily available data; the many criticisms of the method (reviewed by Ward, 1967; 1971) have often been unfair in the sense that they ignore this fact. It is, however, a reflection on the state of the art in this field that the Thornthwaite method generally provides estimates of PE which are as meaningful as those given by many other more complex approaches.

2.7: Runoff and the flood hydrograph

Runoff is the residual water volume after precipitation has been discounted by all abstractions, of which evapotranspiration is usually the most important component. It is the volume of water which is essentially derived from groundwater supply, throughflow and surface (overland) flow, and which leaves a drainage basin as streamflow. It is that component of the hydrologic cycle which largely determines the character of the drainage net and the size of rivers comprising it. But more than that, the pattern of runoff as it appears in the time-distribution of streamflow, represents a synthesis of the complete basin hydrologic cycle; it contains information about the character of precipitation, evapotranspiration, interception, infiltration, surface and subsurface flows, and the basin properties on which each depends. Most of the remaining chapter will be concerned with ways of describing and analyzing the streamflow record in order to decipher some of the information it carries.

2.7 (a) The flood hydrograph and the measurement of discharge

Basin runoff from a storm event can be measured as the total volume of streamflow generated at the mouth of the basin by the storm. The most common method of determining the storm contribution to streamflow is known as hydrograph analysis.

A graph showing the height of the water surface (stage), discharge, velocity, or any other property of flowing river water with respect to time, is in the strictest sense, a hydrograph. When the graph describes such changes during a single flood event it is usually termed a flood hydrograph. Conventionally the term hydrograph, however, normally implies a discharge hydrograph and other types are specified by the appropriate preface; for example, stage hydrograph.

Stage records usually take the form of a direct plot on chart paper by an automatic water-level recorder. Recorders of this type are fixed in a stilling well near the river and record the river height through mechanical linkage to a float or as an electrical signal from a pressure transducer fixed on the bed. Stage hydrographs may also be constructed, of course, by simply noting, at appropriate time intervals, the water level with respect to a staff gauge fixed to the bank of the river channel. This method, once a standard procedure, has the obvious disadvantage that an operator

must frequently be on hand to observe and record the water levels.

The discharge hydrograph is derived from the stage hydrograph through a rating curve relating water-surface elevation to discharge. Figure 2.16 shows such a rating curve for the Colo River near Sydney, Australia. At some discharge gauging stations, such as that on the sandbed channel of the Colo River, rating curves must be adjusted from time to time because of shifting control (channel scour and fill alter the stage/discharge relationship). Gauging stations are usually located so that the problem of shifting control is minimised; ideal locations are in rock-cut sections of channel. At other gauging stations where shifting control in the natural channel cannot be avoided, control structures such as weirs are commonly installed to ensure a stable rating curve.

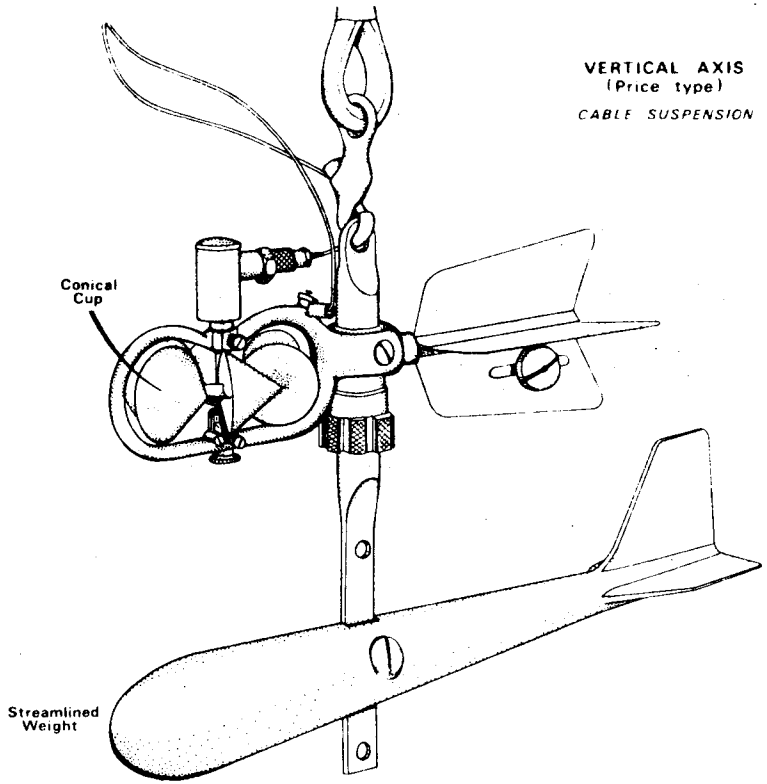
Discharge is defined as the volume of water passing a given point on the river per unit time. It is measured as the product of the channel cross-sectional area and the average flow velocity through the section (see Part III): it has the dimensions of metres^3/second (m^3/s). Flow velocity is usually measured with a current meter consisting of a propellor or cup wheel which rotates in the flow (see Figure 2.17A). The speed of rotation, proportional to the flow rate, is recorded by the operator and converted to a velocity through an appropriate rating curve.

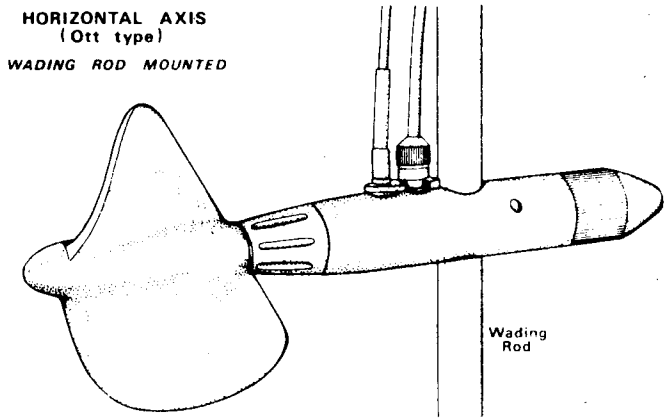
The conventional procedure for discharge measurement is to determine, at a number of verticals of known spacing, the flow depth and mean velocity (see Figure 2.17B). The verticals are usually equally spaced but may be varied to adequately represent flow depth and velocity variations in the cross section. Generally the spacing between adjacent verticals should not exceed 5 per cent of the channel width and the discharge between them should not exceed 10 per cent of the total discharge (World Meteorological Organisation, 1965); these specifications are often relaxed for very uniform channels.

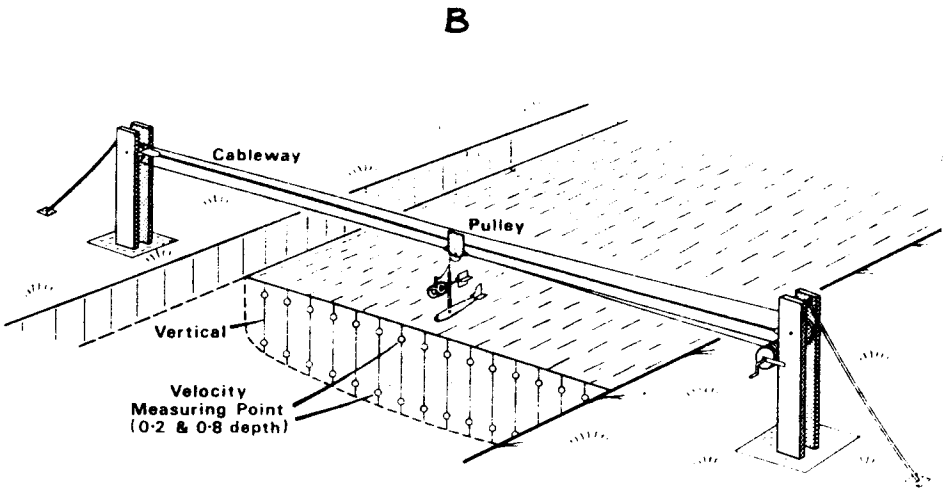
Mean velocity in a vertical is normally taken as the average of point velocities at 0.2 and 0.8 of the vertical flow depth (see Part III) and total discharge is usually computed as the sum of segment discharges, as defined in Figure 2.17B.

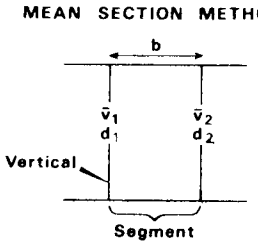
The control structures, and measuring and recording instruments commonly used at gauging stations are described in detail by Boyer (1964) and a more recent review is provided by Gregory and Walling (1973). Discharge measurement using tracer techniques, alternative methods to that described above, rely on the facts that rates of tracer travel and dilution in flowing water are dependent on flow velocity. These techniques are outlined by Gregory and Walling (1973) and are described in detail by Church and Kellerhals (1970) and Church (1975). The reader interested in pursuing these topics further will find a useful bibliography on discharge measurement techniques in Pickett et al (1977).

A typical hydrograph produced by a storm event is a single-peaked skew distribution curve (see Figure 2.18A). Before the storm event, water in the river is supplied from groundwater in the basin. This water, entering the channel as groundwater seepage

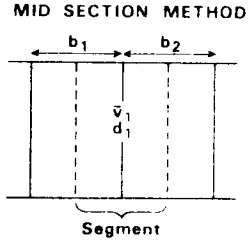








Segment Discharge = $\left(\frac{\bar{v}_1+\bar{v}_2}{2}\right)\left(\frac{d_1+d_2}{2}\right)b$



Segment Discharge = $d_1 v_1 \left(\frac{b_1+b_2}{2}\right)$

Figure 2·17 : Stream gauging instruments and installation
A : Two common types of current meter.
B : A stream gauging facility on a stream too deep to use wading rods.
(from Gregory and Walling, 1973)

through the banks and bed of the river, may sustain a very uniform base flow for long periods between rainstorms. When a rainstorm does occur, the flood hydrograph response is rarely immediate. Although there may be a slight increase in stage as a result of water falling directly into the river near the gauging station, most of the storm water takes some interval of time to find its way into the river channel. For this reason the maximum flood discharge (the peak or crest) lags behind the storm mass centre by a time interval termed the basin lag.

The storm runoff from a short-duration intense rainstorm produces a relatively rapid rise in river discharge up to the peak flow. This segment of the flood hydrograph, termed the rising limb, rises abruptly and steeply above the base flow to the flood peak. The peak flow usually marks the transmission of runoff produced during the brief maximum intensity of storm rainfall. After the peak flow, discharge and stage fall rather gradually as less intense runoff from the weakening phase of the storm continues to decline and as groundwater supply once again becomes the major source of water for the river. This segment of the hydrograph is known as the recession curve or limb; the lower part of this segment is the groundwater recession or depletion curve. It describes the slow decline in the rate of groundwater supply to the river after the storm and typically takes the form:

$$Qt = Q_o K_r^{\,t} \qquad (2.48)$$

where Q_t is the discharge at any time t after Q_o, and K_r is a recession constant. That is, the rate of release of water from storage in the ground declines exponentially with time after a rainfall event. Because of this property, the groundwater recession curve will plot as a straight line on semi-logarithmic paper (discharge on the logarithmic scale) and provides the basis of several essentially arbitrary methods of separating the surface flow, subsurface or through flow, and groundwater components of the flood hydrograph; one such method is illustrated in Figure 2.18C (see Chow, 1964 for further discussion of hydrograph partitioning).

For our purposes, the simple division of the hydrograph into direct runoff and base flow is adequate. Again, the division is subjective and usually involves drawing a straight line from the point of rise to some arbitrary point on the lower portion of the recession limb of the hydrograph. Chow (1964) advises that "this arbitrary point may be so chosen that the base-flow separation line should not be too long and, on the other hand, the base flow should not rise too high"; Figure 2.18B illustrates the application of these guidelines.

The total runoff from a given storm event is the area under the related hydrograph and above the base-flow line, and has the dimensions of m^3. Runoff is also commonly expressed as runoff volume per unit drainage area. This latter measure is the depth of water (also called the effective rainfall) which must uniformly cover the whole drainage area in order to yield a given volume of runoff. It is always less than the average depth of water received as precipitation because, as we have seen, much of the precipitation is abstracted to components of the hydrologic cycle other than runoff. The world's rivers carry an annual discharge re-

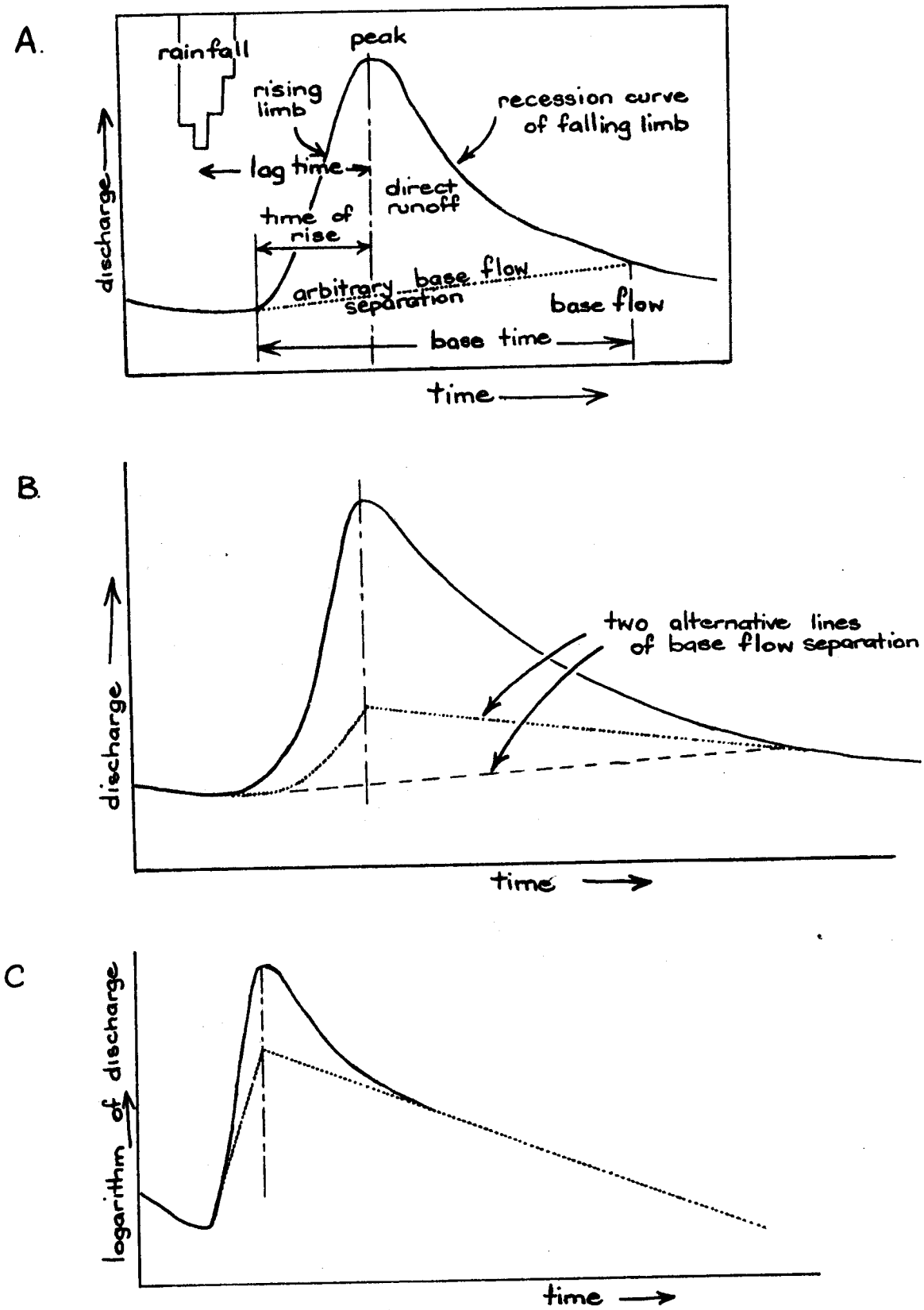


Figure 2.18: Flood hydrograph characteristics
A: Definition of hydrograph terms
B: Two methods of base flow separation
C: Base flow separation using semi-logarithmic transformation of the hydrograph.

presenting about one-third of the annual precipitation. For individual rivers the runoff/precipitation ratio may vary from almost zero to unity depending on the climatic, geologic and geomorphic characteristics of the basin.

2.7 (b) Two views of the runoff process

We have thus far considered the flood hydrograph response to a storm event without reference to the specific mechanism governing direct runoff. The mechanism, however, is far from being self-evident. There are in fact two basically different types of basin runoff models that we should briefly consider: the Horton overland flow model, and the throughflow model.

The overland flow model (Horton, 1945) is the classical theory of basin runoff in the sense that it has enjoyed longstanding and almost universal acceptance as a conceptual model. It is the concept on which are based many other hydrologic models such as unit hydrograph theory (Section 2.7c) and some flood routing models (Section 2.10). Basically the Horton model divides streamflow into two sources: overland flow and groundwater flow (see Figure 2.19A). Rain falling on the basin surface infiltrates into the soil and gradually percolates to the groundwater table, below which groundwater is stored as an underground reservoir supplying water as base flow to rivers. Should the rainfall intensity exceed the capacity of the soil to absorb the falling rain (the infiltration capacity) the water stays on the ground surface, at first to be held in hollows and irregularities as surface detention, and eventually to form overland flow.

In recent years, however, considerable attention has been given to measurement and analysis of rainfall intensities and infiltration rates in a variety of conditions (see Yarnell, 1935; Hershfield, 1961; Musgrave and Holtan, 1964; Whipkey, 1965) and it has become obvious that, for some types of drainage basins, Horton overland flow rarely occurs. Infiltration rates for many natural surfaces, with some exceptions such as bedrock, unvegetated and frozen ground, are considerably greater than all but the most exceptional rainfall intensities. This realisation has given credence to an alternative runoff model based on throughflow responses to rainfall (see Kirkby and Chorley, 1967; Kirkby, 1969). In terms of specifying the relationship between rainfall intensity and hydrograph response, the throughflow model is little different from the Horton model. The mechanism giving rise to the rainfall-runoff response, however, is entirely different (see Figure 2.19B). Basically the throughflow model states that, when rain falls onto a basin surface, all of it infiltrates into the soil. Some of the water moves downward (if the soil reaches capacity) to recharge the groundwater reservoir, but most of it flows down the hillside within the soil layers as throughflow and eventually contributes to streamflow. The concentration of throughflow near the basin surface occurs there because permeability is greatest in the light textured organic A_0 and eluviated A_1 soil horizons and very much less in the underlying clay-rich B horizon.

The throughflow discharge can be determined from a basic principle of the physics of saturated flow through porous media, known as Darcy's law:

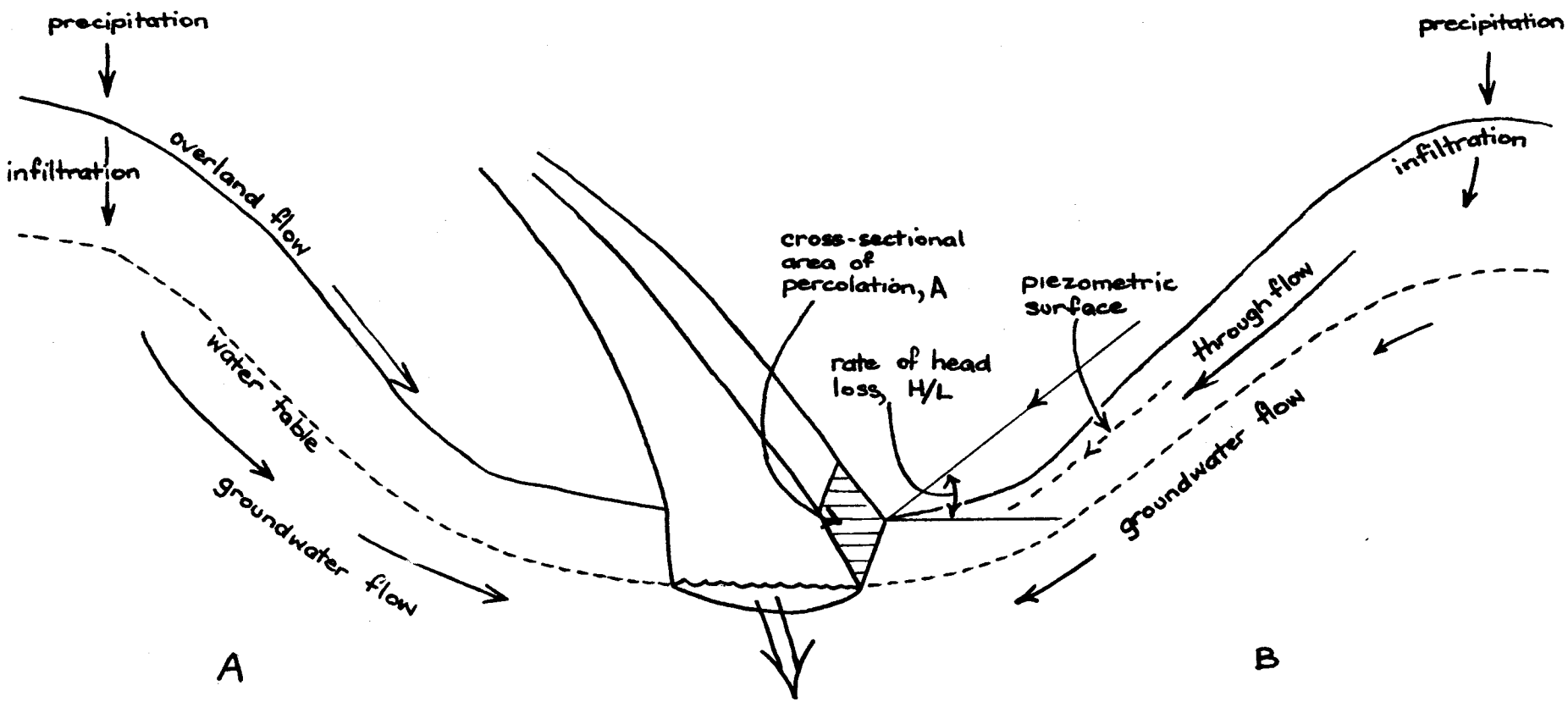


Figure 2.19 : Two views of the runoff process.
A : The Horton overland flow model.
B : The throughflow model with definition of the Darcy law parameters.

$$Q = K(H/L)A \qquad (2.49)$$

in which Q = the volume of water discharged per unit time (m^3/day)
K = the coefficient of permeability or hydraulic conductivity (m/day)
H/L = the rate of piezometric head loss (see section 2.5)
and A = the cross-sectional area of percolation (m^2)

The application of these parameters to a section of drainage basin is shown in Figure 2.19B.

The velocity of throughflow (v) towards the river can be approximated by

$$V = Q/AP_o \qquad (2.50)$$

where P_o is the porosity of the percolated material. Equation 2.50 will only yield the exact throughflow velocity if the porosity, the proportion of pore space to total volume of the percolated material, is a relative measure of interconnected pore space. Thus, even given accurate measurement of A, Q, and P_o, velocities of throughflow predicted by equation 2.50 will usually be low.

The coefficient of permeability, expressed as m^3/m^2/day, (i.e., m/day) range from 10^{-6} and 10^{-4} m/day for clay to 10^2 - 10^6 m/day for gravel (Gregory and Walling, 1973); it is determined experimentally by solving for K in equation 2.49.

Darcy's law has also been found to apply to unsaturated porous media (Childs and Collis George, 1948, 1950; Richards, 1931) if the head-loss term includes the combined gradient of suction and gravitational forces in the soil (K in this case is termed the capillary conductivity).

As we noted in section 2.5 (c), throughflow velocities can also be directly measured by timing the movement of tracers through the soil. This method is time consuming, involving the recording of dye movement downslope from one bore hole to another, but it probably provides the most accurate estimates of throughflow velocities.

There are several important distinctions between the throughflow and Horton models that we should recognise. First, studies of soil-water movement have shown that rates of throughflow are perhaps a thousand times smaller than those for overland flow (Kirkby, 1969). That is, most Horton overland flow will contribute directly to the flood hydrograph but most throughflow will be unable to reach a channel until long after the rainfall has stopped and the flood peak has passed.

Second, an implication of the first distinction is that, whereas in overland flow it is the rain water of the current storm that forms the flood water during that storm, in throughflow much of the flood water flowing into the channels during the current storm fell in previous storms. In other words, the water infiltrating into the basin surface must displace all of the downslope soil water before it can gain access to the channel. Stated another way, storm

water enters the channel in the same order as it infiltrated the soil of the basin.

The third and related distinction between the two runoff models is that, unlike the Horton model, storm runoff in the throughflow model involves only a small proportion of the total basin area. Only water from a relatively small contributing area near the channels is able to reach them in time to form part of the flood hydrograph (see Hewlett and Hibbert, 1967).

Thus, in terms of the Horton overland flow model, the direct runoff component of the flood hydrograph constitutes surface flow, and in terms of the throughflow model it constitutes subsurface flow. We should recognise, of course, that most real-world drainage basins do not conform exactly to either the Horton or the throughflow models, but instead are some combination of both.

2.7 (c) Unit hydrograph theory

Before we examine some specific environmental influences on the form of the flood hydrograph, it may be instructive to briefly consider the concept of the unit hydrograph. Because hydrograph form varies with the magnitude and time-distribution of the related rainfall event, Sherman (1932) suggested the unit hydrograph as a means of standardising these controls in order to reveal the characteristic hydrograph of a basin. The unit hydrograph of a basin is defined as a hydrograph of direct runoff resulting from one centimetre of effective rainfall distributed uniformly in time and space over the basin. This definition, together with the following assumptions, constitute the unit hydrograph theory:

(a) The runoff-producing rain (effective rainfall) over the drainage basin is uniform in time and space.

(b) The base or time duration of the hydrograph of direct runoff from an effective rainfall of unit duration is constant, regardless of the rainfall intensity.

(c) The ordinate values of all direct-runoff hydrographs of a common base time are proportional to the volume of direct runoff in each case.

These assumptions are rarely satisfied in nature and the theory should only be applied to those drainage basins where hydrologic conditions do not badly violate them. The theory clearly does not apply to any basin in which runoff is supplied by snow and ice meltwater. Ideally the basin should not be so large that spatial uniformity of rainfall is unlikely, and the storm event should be single peaked and relatively short in order to approximate temporal uniformity of rainfall. When hydrologic data are carefully selected so that they meet the above assumptions closely, unit hydrograph predictions have been found to be acceptable for many engineering applications

Actually, Sherman (1932) originally intended the term unit to specify the period of time of effective rainfall but the term

has more recently been applied to the rainfall magnitude rather than to duration. I will use the term unit hydrograph in this latter sense, although it has been suggested (Walling, 1971) that hydrographs standardised on unit depth of runoff may be better termed unit-response hydrographs.

The procedure for deriving the unit hydrograph is illustrated in Figure 2.20A; the observed hydrograph is for the Beatton River in Northeastern British Columbia during August, 1962. You will notice that, although this hydrograph generally resembles the ideal form in Figure 2.18A, it clearly is less regular. These minor irregularities in hydrograph form are typical and reflect the unique combination of climate, topography, geology, drainage-network properties, etc., which characterise every individual drainage basin. It is only when the irregularities or deviations begin to dominate the normal form that we need to abandon an ideal model.

The first step in deriving the unit hydrograph is to isolate the direct runoff by subjectively partitioning the base flow in the manner previously described. The direct runoff hydrograph is then constructed by subtracting the base flow from the observed flow for each ordinate in the range of the base time. The product of the mean direct runoff (231m /s) and the direct-runoff duration (12 days or 1.0368×10^6 secs.) yields the total volume of direct runoff ($2.395008 \times 10^8 m^3$). Dividing this volume by the drainage area to the Beatton River gauging station (1.6058×10^{10} m^2) yields an effective rainfall of 1.49 cm. The final step in the analysis involves the adjustment of the ordinates of the direct-runoff hydrograph to yield an effective unit rainfall of 1 cm. Clearly, dividing each ordinate value by 1.49 achieves this end and results in the unit hydrograph shown in Figure 2.20A. The effective rainfall duration in this case is approximately 12 hours and the unit hydrograph is more properly termed the 12 hour unit hydrograph. In other words, it is the hydrograph which would result from 1 cm of effective rainfall evenly distributed over a 12 hour period (i.e., at a rate of 0.083 cm/hour).

To assess the influence of extended effective rainfall duration on the form of the hydrograph, ideally we should examine longer-record data. It is often the case, however, that such data are not available and we must resort to generating synthetic data. For example, it is possible, by simple integral summation of unit hydrographs, to derive hydrographs produced by varying rainfall durations. Consider the Beatton River hydrograph shown in Figure 2.20A. Although it was produced by an effective rainfall duration of approximately 12 hours, we can estimate the form of the hydrograph for the same rainfall intensity but twice the effective rainfall duration by offsetting the unit hydrograph a further 12 hours and adding the ordinates to the 24-hour hydrograph and repeating the process for another additional 12-hour offset. This procedure, first suggested by Morgan and Hullinghors (1939), is sometimes called s-hydrograph analysis because the envelope curve defined by repeated offsets assumes a distorted s-shape as it becomes asymptotic to a discharge corresponding to the rate of effective rainfall (see Figure 2.20B). In other words, it is the hydrograph resulting from continuous effective rainfall at some specified rate.

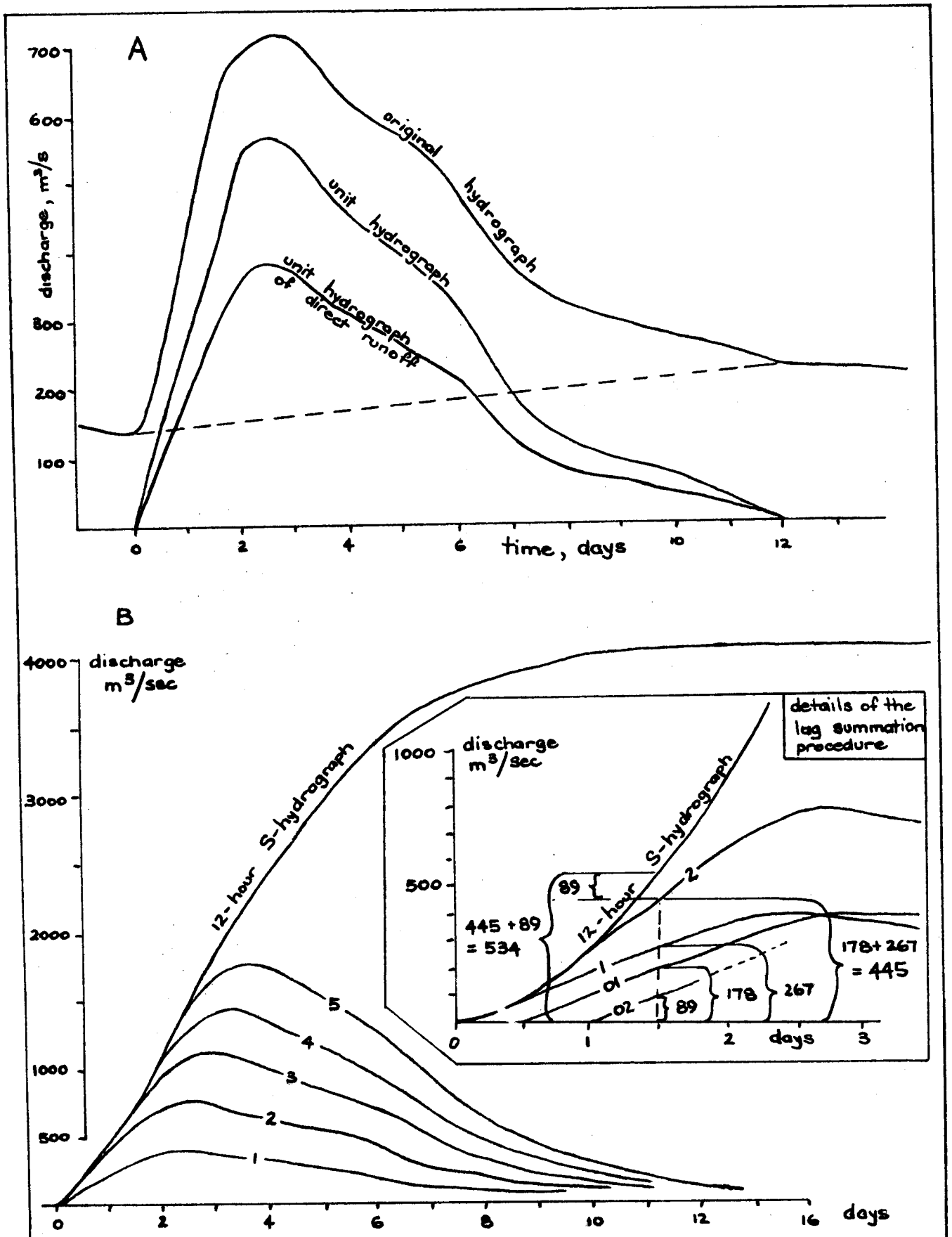


Figure 2.20 : Unit hydrograph derivation
A: Unit hydrographs derived for the Beatton River, British Columbia
B: Unit hydrograph derivation for simple multiples of the original 12-hour duration; the envelope curve is the S-hydrograph.

Just as the s-hydrograph can be composed by the integration of a number of unit hydrographs it can also be readily differentiated or decomposed into its component hydrographs of a specified duration by the graphical technique shown in Figure 2.21A. Although it has been useful to visualise the s-hydrograph as the product of this type of integral summation, it is more easily constructed, however, as the cumulative discharge graph of the corresponding unit hydrograph. Offsetting this s-hydrograph by some duration t_o and subtracting the offset ordinate from the original ordinate at a given time t, yields the t ordinate for the corresponding hydrograph. The process is repeated for the given offset until sufficient data points are available to define the hydrograph in full.

Because hydrographs produced in this way are of longer duration than the original hydrograph, they represent more than the equivalent of 1 cm of effective rainfall and therefore are not unit hydrographs. The hydrograph ordinates must be divided by the number of offsets corresponding to the specified duration. For example, a 48-hour duration hydrograph will have ordinates four times greater than those of the 12-hour unit hydrograph from which it was derived; thus, each ordinate of the former curve must be divided by four to yield a comparable unit hydrograph.

Several hydrographs for the Beatton River have been derived in this manner and are shown in Figure 2.21B. They clearly illustrate the application of unit hydrograph theory to river flow-response problems. For example, the five derived unit hydrographs describe how a given effective rainfall ranging from 12 to 240 hours duration influences the magnitudes of the time of rise and of the flood peak.

Throughout the foregoing discussion of unit hydrograph theory I have presented the basic arguments in graphical terms in the hope that it might aid the understanding of the principles involved. The derivation of unit hydrographs using numerical methods, however, is much more direct and less likely to include errors than the corresponding graphical procedures. Several hydrograph problems are solved using numerical methods in the illustrative examples at the end of this chapter.

For the reader interested in examining this topic further you will find detailed discussions of unit hydrograph theory (and related topics such as the conceptual models of the instantaneous unit hydrograph) in most standard hydrologic works (for example, see Linsley Kohler and Paulhus, 1949, 1958; Chow, 1964).

We will find a need to return more than once to this useful concept of the unit hydrograph during the following discussion of the physical mechanisms which determine the shape of the flood hydrograph.

2.8: Factors controlling flood hydrograph characteristics

Most of the important controls on the form of the flood hydrograph can be deduced from a careful consideration of the basin hydrologic cycle. Clearly, any factors which influence the input of precipitation, the loss or storage of this water, or its transfer from

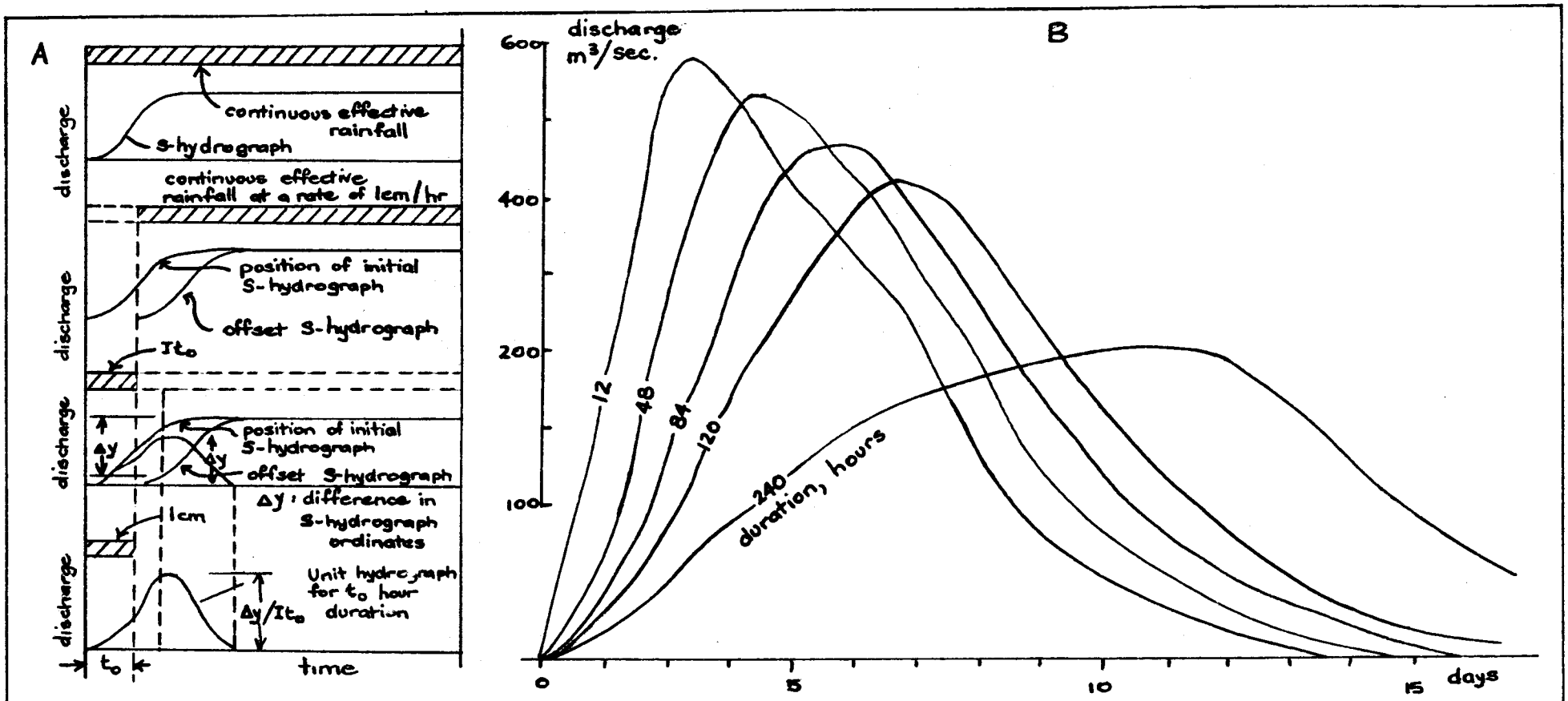


Figure 2:21
A : Derivation of the unit hydrograph from the S-hydrograph (After Chow, 1964)
B: The Beatton River hydrograph (see Figure 3·20A) for several flow durations.

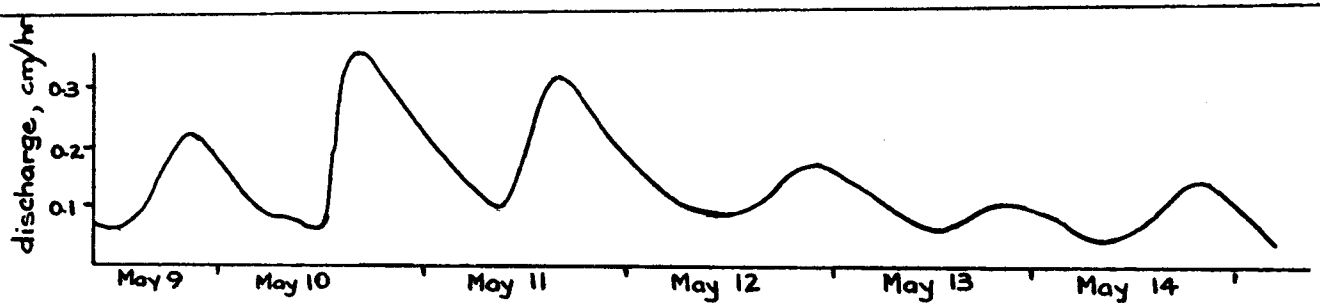


Figure 3·22: Diurnal fluctuation in snowmelt runoff from a subarctic snowpack (after Dunne et al, 1976)

hillslope to and through the stream channel, will have an influence on the size and shape of the hydrograph. Accordingly we can develop a list of controlling factors such as that shown in Table 2.6.

It is not my intention to examine every possible effect on the flood hydrograph of variations in all of the potential influences discussed throughout this chapter ! Instead, I will confine the discussion to major hydrologic responses to the three general categories of controls listed in Table 2.6.

2.8 (a) Input controls

The nature of precipitation is, of course, a fundamental control on the form of the flood hydrograph. If the precipitation is in the form of snow it usually must be regarded as an abstraction from the basin hydrologic cycle until the Spring thaw. Then, depending on the quantity of snow and the meltrate, the stored water may simply contribute to a general rise in base flow or it may produce a snowmelt flood-peak similar to that produced by a rainstorm. Although the distinction between base flow and direct runoff is not particularly meaningful in the snowmelt situation, the general hydrograph response is the same as it is to a rainfall event. That is, in this context, rainfall and snowmelt intensities, durations, and spatial distributions, can be considered interchangeable and most of the remaining discussion will be in terms of rainfall events. In general we can expect that, other things remaining constant, the greater the snowmelt rate, the greater will be the magnitude of base flow in the channel. Furthermore, if snowmelt contributes to direct runoff, the flood hydrograph peak will increase as the snowmelt rate increases. A short burst of snowmelt runoff will produce a sharply peaked hydrograph and sustained snowmelt will produce a broad flood peak. Snowmelt flood hydrographs which peak and recede over many days, typically display a "sawtooth" discharge trace produced by the diurnal fluctuation in temperature and meltrate (see Figure 2.22).

Equation 2.1 would also lead us to expect, and it is commonly observed, that the total volume of storm runoff (the area under the flood hydrograph above the base flow) usually increases with increasing precipitation input. This general rainfall-runoff correlation forms the basis of several runoff models (for examples, see Young, 1948; Langbein et al, 1949; Kohler and Linsley, 1951; Butler, 1957, Sutcliff and Rangeley, 1960).

A direct plot of runoff versus rainfall for individual storms, however, does not usually produce a high degree of correlation (Chow, 1964) partly because it does not distinguish between two very important aspects of rainfall: intensity and duration. A storm rainfall of 20 mm might be produced by rain falling at a rate (intensity) of 20mm/hour for one hour duration or it could be produced, for example, by five hours of rainfall at one fifth of this intensity. Each equal-volume rainfall event will produce flood hydrographs which are very different to each other.

Horton (1933) envisaged the flow types of hydrograph response shown in Figure 2.23 as direct results of variations in rainfall intensity. His Type 0 hydrograph represents no hydrograph response

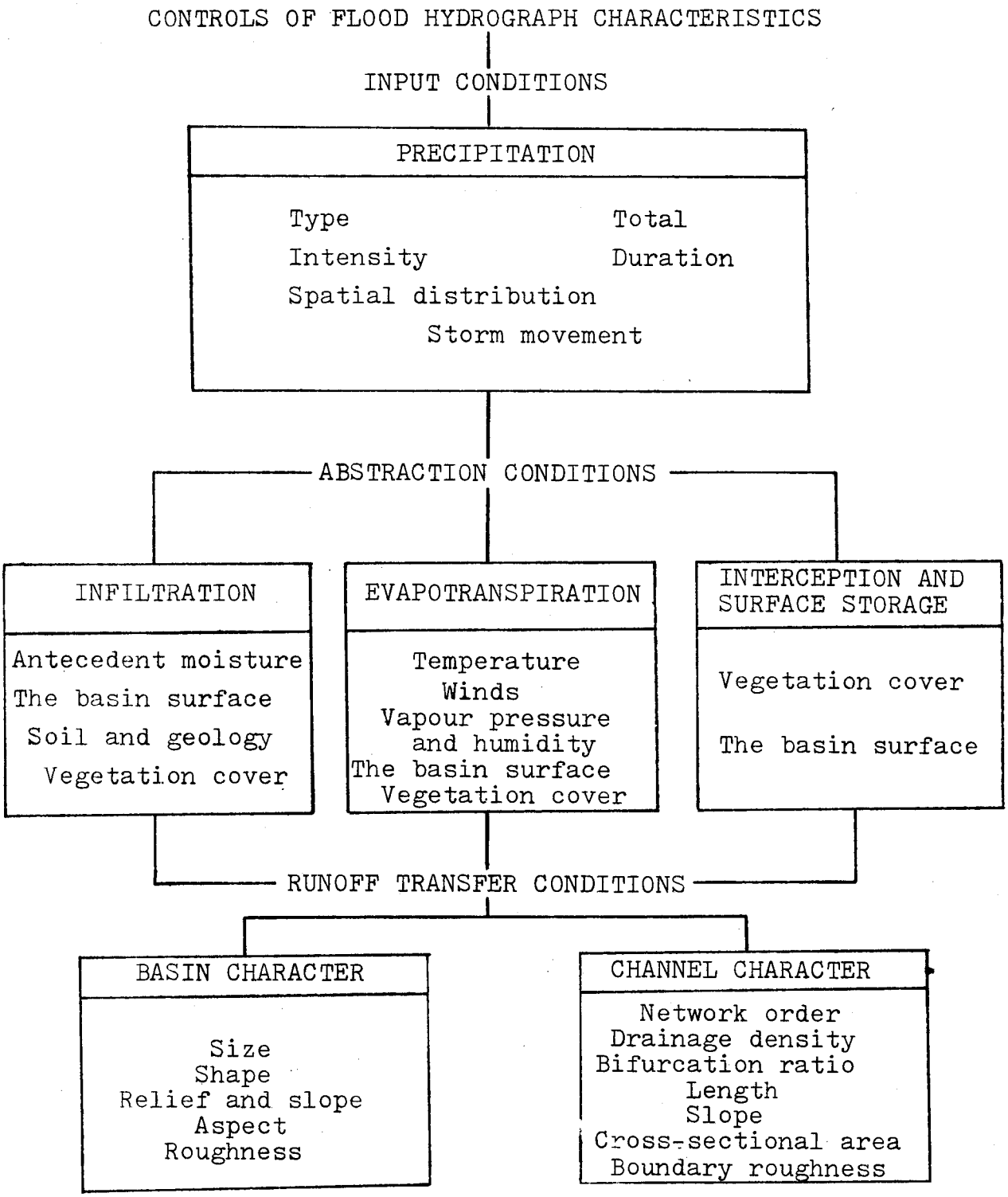


Table 2.6: Controls of the flood hydrograph characteristics

in spite of rainfall. It occurs when rainfall intensity is so low that not only does it fail to exceed the infiltration capacity of the soil but it is also unable to replace the initial deficit in soil moisture. Thus there is no transfer of water from the soil to groundwater and consequently no increase in groundwater runoff. The resulting hydrograph corresponds to the normal dry weather recession curve. In strict terms there must actually be a slight increase in streamflow because some of the rain will fall directly in the stream channels. These Type 0 conditions are of some significance because repeated events could bring the soil moisture to the maximum amount that can be held in the soil against the downward force of gravity (field capacity), and further rain of the same low intensity will contribute to the groundwater reservoir.

In Type 1 conditions (Figure 2.23B) the rainfall intensity is greater than that for Type 2 but it remains less than the infiltration capacity of the soil and so surface runoff is produced. The rainfall intensity is high enough, however, to replace the initial soil moisture deficit and add water to groundwater supplies. The result is that the recession curve is shifted upwards in response to the increase in groundwater discharge (base flow), after which the normal recession pattern is re-established. Of course, it is also possible that the rate of groundwater supply from rain is less than the rate of groundwater depletion in which case there would be no hydrograph response.

In Type 2 conditions (Figure 2.23C) the rainfall intensity is greater than the infiltration capacity and the resulting surface runoff increases streamflow. The infiltration which does occur, however, is not sufficient to replace the initial soil moisture deficit and the runoff peak is simply superimposed on the continuous recession curve.

The Type 3 conditions (Figure 2.23D) involve a combination of high rainfall intensities and high infiltration rates so that a runoff peak is produced and groundwater reserves also are increased. This is clearly a case of a runoff peak being added to the hydrograph of Type 1 conditions.

We should recognize, however, that these two alternative models represent ideal cases. Most natural basins conform neither to the Hortonian nor the throughflow models, but are instead some combination of both.

Unit hydrograph theory appears to qualitatively describe the influence of rainfall intensity on hydrograph form reasonably well. Figure 2.20A reflects an assumption of the theory that rainfalls of the same duration but differing intensities, will yield hydrographs of the same base time but of varying flood-peak magnitude. If other factors remain constant, it is clear that the magnitude of the flood peak will increase as rainfall intensity increases.

This relationship forms the basis of the so-called "rational" formulae for predicting peak discharges in rivers (see Bruce and Clark, 1966).

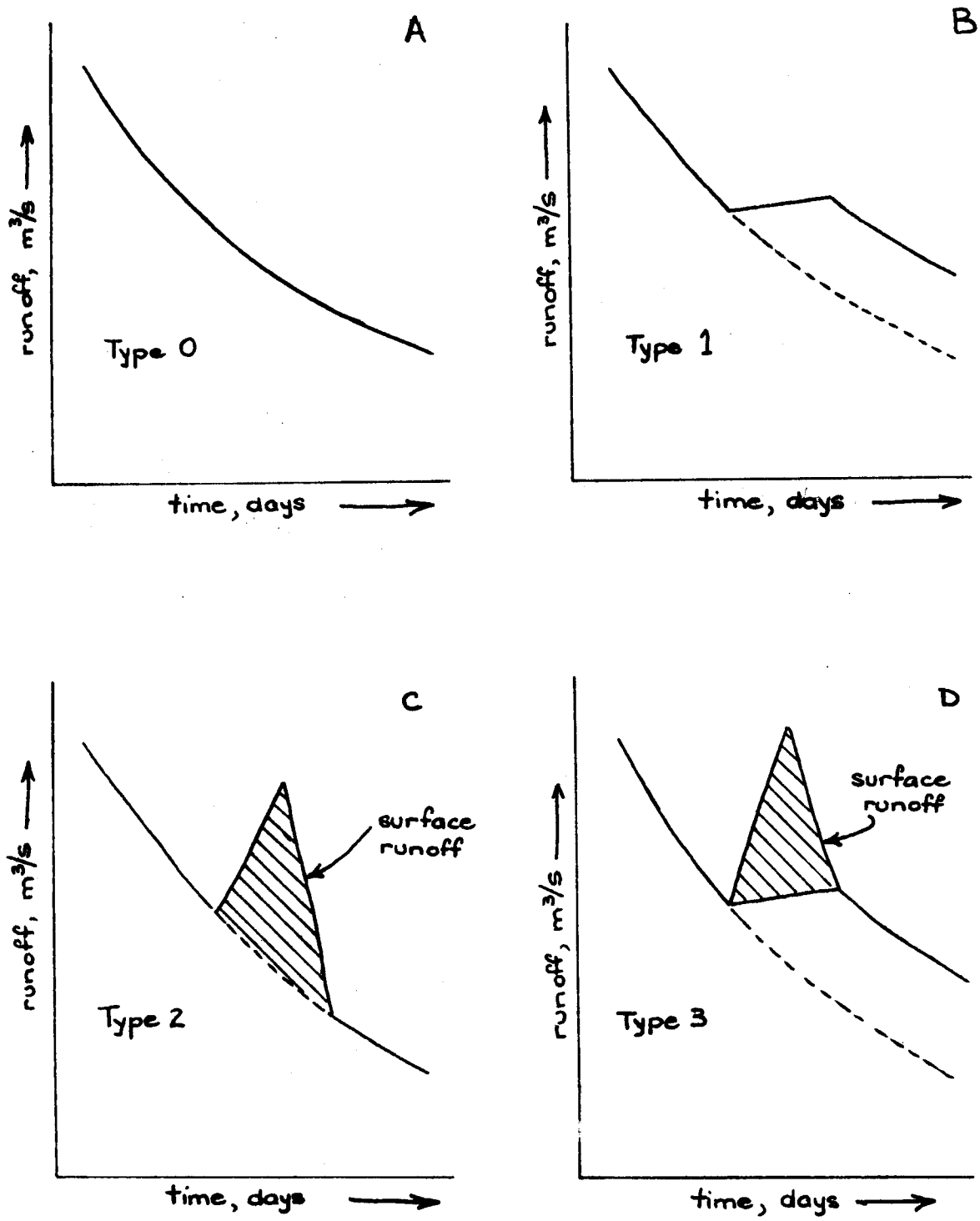


Figure 2·23: Hydrograph response to variations in rainfall intensity and infiltration rate (based on Horton, 1933).

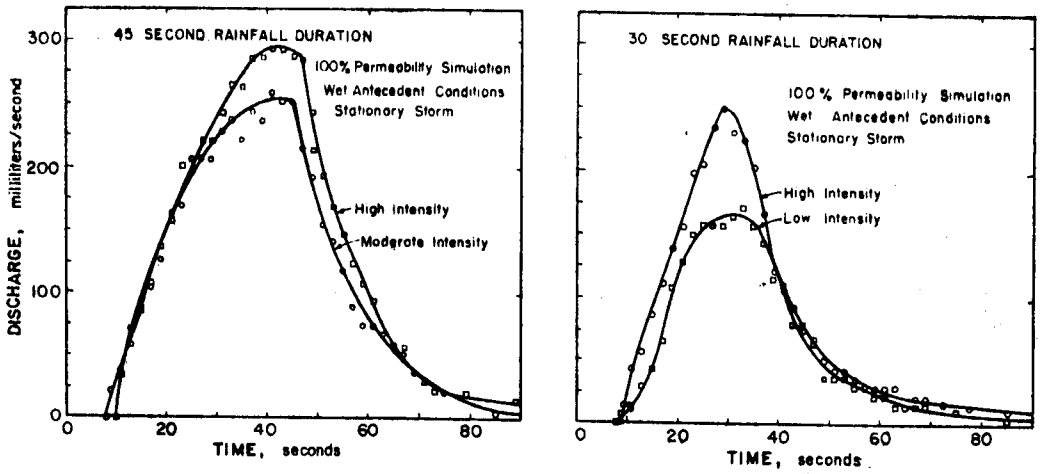
The correlation between rainfall intensity and the magnitude of the flood peak has been demonstrated in an interesting experimental study conducted by Roberts and Klingeman (1970). They documented hydrograph response to a variety of simulated rainfall and basin properties in a laboratory model of a catchment. Figure 2.24A shows the runoff hydrograph from the moist but permeable surface of the model basin under conditions of uniformly distributed rainfall of varying intensity. An increase of about 20 per cent in the intensity of rainfall for two 45-second-duration rainfalls produced a 20 per cent increase in the magnitude of the flood peak. Similarly, an increase of about 40 per cent in rainfall intensity for two 30 second simulated storms produced a 40 per cent increase in the magnitude of the flood peak. Figure 2.24A also displays an offset of the rising limbs during the 30-second rainfall; at the low intensity of rainfall a relatively long period is needed to saturate the basin surface because some volume of rainfall must first contribute to surface abstractions such as interception, infiltration and surface storage. The time required is dependent on the intensity of rainfall.

Roberts and Klingeman (1970) also examined the influence of rainfall duration on the shape of the flood hydrograph. Their results (see Figure 2.24B) indicate that the simple integral summation assumption of unit hydrograph theory (see Figure 2.20B) appears to be valid in this case. As rainfall duration of storms of the same rainfall intensity increases, the flood response is described by a series of nested hydrographs on a common rising limb. Each displays increasing flood peak magnitudes and basin lags until equilibrium discharge is established for the given rainfall intensity. We should not expect the flood hydrograph for a natural basin to respond in quite the same systematic way as that for the experimental basin, because in the latter case, antecedent soil moisture conditions and the areal distribution of rainfall were not variables and they are almost always so in natural conditions. Nevertheless, we can say with some confidence that, if other factors remain constant, the quality of changes in the flood hydrograph responses to increasing rainfall duration, are well represented by the unit hydrographs of Figure 2.20B.

But of course rainfall intensity and duration are not the only aspects of precipitation which can influence the shape of the flood hydrograph. Another important factor is the spatial distribution of precipitation within the watershed. For example, a storm centred over the gauging station would produce a flood hydrograph with a relatively small basin lag and a large flood peak. The same storm centred over the headwater areas of the basin would yield a flood hydrograph with a greater basin lag and a smaller flood peak. In the former case the precipitation has only a short distance to travel from the point of impact to the gauging station and the hydrograph response is abrupt and pronounced; in the latter case the more subdued hydrograph response is the result of runoff abstraction during the relatively long journey to the gauging station.

These ideas of travel time and flow abstraction from stationary storms can easily be extended to the case of a moving storm. Roberts and Klingeman (1970) simulated downbasin and upbasin storm movement in their laboratory experiments and produced the flood hydrographs shown in Figure 2.25. In the case of upbasin movement of the storm

A



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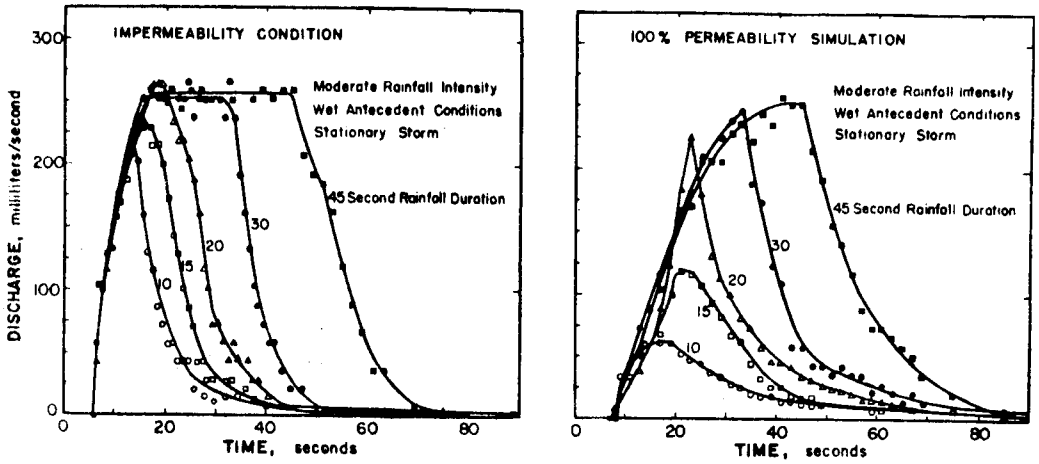


Figure 2.24: The relation of simulated rainfall characteristics ties to hydrograph response

A: The effect of rainfall intensity on the runoff hydrograph

B: The effect of rainfall duration on the runoff hydrograph.

(After Roberts and Klingeman, 1970)

the hydrograph time of rise was very short because of the initial proximity of the storm to the gauging station. The flood peak was relatively low and prolonged, however, because of the increasing distance (and travel time) between the storm centre and the gauging station. In the case of the downbasin storm movement, on the other hand, the time of rise is greater because of the relatively long travel time at the start of the storm. Furthermore, the flood peak is considerably greater because the stream flow from upbasin precipitation moves with the storm centre to a concurrent arrival at the basin mouth. Because of the time-concentrated delivery of runoff to the gauging site in this latter case, both the rising limb and the recession curve are steeper than those in the case of upbasin storm movement.

Again, we should recognise that storm character and movement over natural basins are likely to be much more complex than in this simple simulation model. Nevertheless, the general principles illustrated by these experiments should apply in all cases.

2.8 (b) Abstraction conditions

The next major group of controls on flood hydrograph characteristics listed in Table 2.6, the abstraction conditions, include infiltration, evapotranspiration, interception and surface storage.

High infiltration rates, promoted by low soil-moisture status, low-angle and rough basin surfaces (surface flow is retarded allowing more opportunity for infiltration), permeable basin materials, and dense vegetation cover, in the Horton runoff model gives rise to hydrographs with relatively subdued flood peaks with a large basin lag-time and a long time to rise. Similarly, the same storm event associated with low infiltration rates promoted by the converse conditions, give rise to rapid hydrograph response and a large flood peak. The response of flood hydrographs in the throughflow model is more complex. If infiltration rates at the surface are high and there exists at a shallow depth an impermeable soil horizon, then throughflow velocity will be relatively high. We also know from equation 2.49 that the throughflow velocity increases as the hydraulic gradient (approximately equal to hillslope angle) increases. Thus we can expect rapidly responding and peaked hydrographs in basins with steep slopes. On the other hand, if no permeable layer exists at depth, the rate of throughflow, other things being constant, will increase as the soil moisture content of the soil increases up to saturation level.

On balance most natural drainage basins tend to behave rather like the Horton model with respect to the infiltration rate/hydrograph-response relationship, particularly if the soil moisture content is high. These conditions especially apply in basins in which snowmelt keeps much of the surface materials close to saturation, and in all basins immediately following heavy or prolonged rain.

The influence of permeability and antecedent moisture conditions is graphically illustrated by the experimental study by Roberts and Klingeman (1970). Figure 2.26A shows the increases in hydrograph lag-time and in the time of rise accompanying increases in simulated basin-permeability. Although all hydrographs reached peak equilibrium for this 45-second rainfall, one of shorter duration would also have produced a decline in the flood leak as permeability increased, (see Figure 2.24B).

The experimental hydrograph for wet and dry antecedent conditions, is shown in Figure 2.26B. In dry conditions the hydrograph is displaced to the right by the time it takes for the "soil" moisture to reach capacity in some part of the basin. In natural conditions a typical storm would not produce sufficient rain to reach equilibrium discharge and the flood peak in the dry antecedent conditions would not only have a greater lag-time but it would also be smaller.

The antecedent moisture condition is regarded by many investigators to be sufficiently important to warrant its inclusion as one of the few variables in many simple empirical hydrograph-response models (for example, see Butler, 1957; Glasspoole, 1960; Gregory and Walling, 1973).

Because infiltration rates are related either directly to basin geology or indirectly through the surficial materials that develop as weathering products, some authors have reported runoff/precipitation ratios for characteristic pedalogic and geologic settings. For example, Serra (1954) reported this ratio to be respectively 0.81, 0.63, and 0.17 for basalt, granite, and moraine basins in southern France (also see Stafford & Troxell, 1944; McDonald & Langbein, 1948; Lvovitch, 1957; Ayers, 1966; Wright, 1970).

The principal direct influence of evapotranspiration on the flood hydrograph is to reduce the total volume of storm runoff relative to precipitation (runoff yield).

Interception and surface storage similarly exert an influence on the flood hydrograph by limiting the volume of precipitation available for runoff. But it is a type of theshold control which operates only up to the point when the various types of storage reach capacity. Thereafter, interception and surface storage exert a negligible influence on the flood hydrograph properties. The effect of these storage elements is to shift the hydrograph to the right in the manner depicted in Figure 2.26B. Because natural basin surfaces display a normally heterogeneous spatial distribution of surface storage and interception capacity, during a storm some parts of the basin will contribute runoff before others. The net effect is for the hydrograph lag-time and time of rise to increase as surface storage and interception capacity increases. Increases in storage capacity, with other things constant and precipitation volume less than the total capacity, will tend also to depress the flood peak. Basins with large surface storage capacity, such as those formed in the knob and kettle and hummocky moraine country of southern Ontario and the Great Lakes States, may require very substantial precipitation before any direct runoff is produced.

As an agent of hydrograph control, probably the single most important physical abstraction element, is the vegetation cover. Temporal changes in vegetation cover are commonly produced by natural disturbances such as insect defoliation, lightening fires, and wildlife grazing pressures, and catastrophic changes certainly can be wrought by human activity. Because of forest industry practices in several countries there has been ample opportunity to observe the net effects on hydrograph response to forest changes ranging from selective cutting of large single-species trees to clear cutting of the complete vegetative cover. Indeed, the literature on the hydrologic responses to

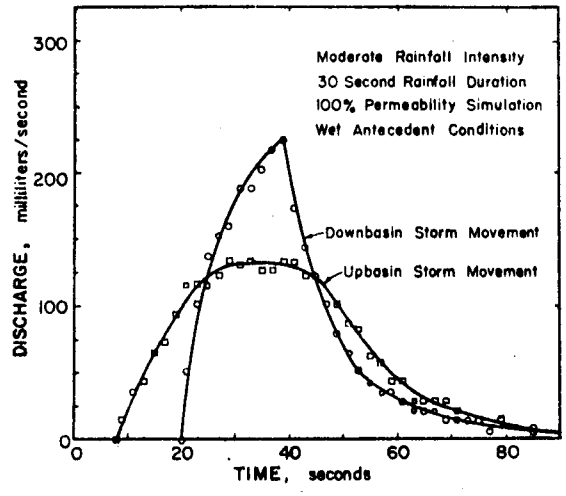


Figure 2·25: The effect of simulated storm movement on the runoff hydrograph for a laboratory basin. (from Roberts and Klingeman, 1970)

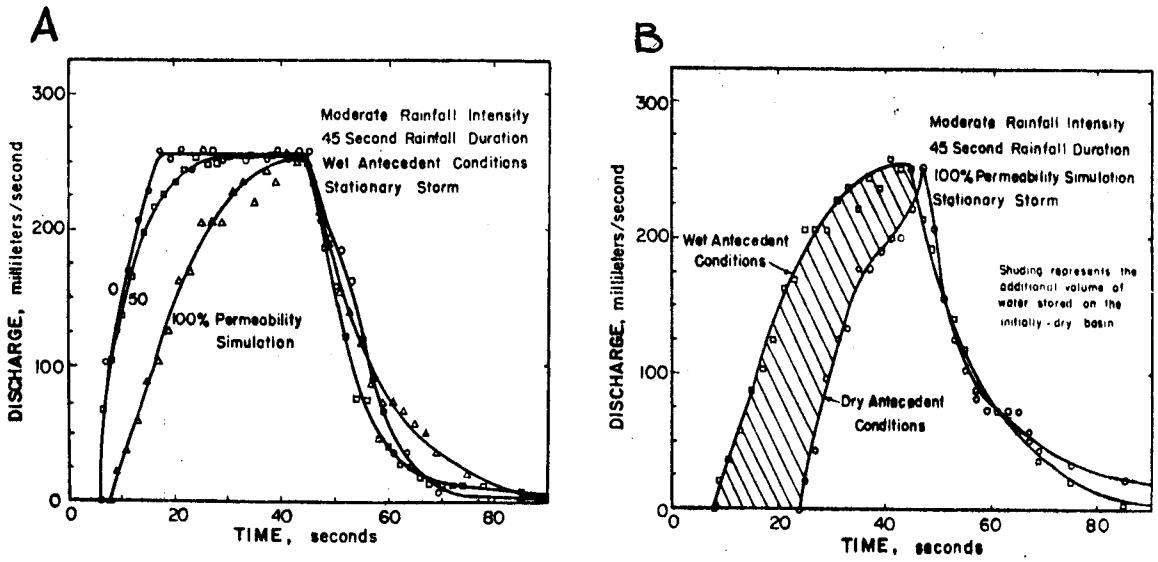


Figure 2·26: The effect of basin surface properties on the runoff hydrograph for a laboratory model.

A : The relation between permeability and the hydrograph shape

B: The relation between antecedent moisture conditions and the hydrograph shape.

(From Roberts and Klingeman, 1970)

the harvesting of forests is quite voluminous; much of it is reviewed by Penman (1963), Hibbert (1967), and Jeffrey (1970).

In general these studies indicate that, as the proportion of vegetation cover in the basin is reduced by logging, there is an increase in runoff yield. For example, Goodell (1958, 1967) reported that when an experimental basin in pine forest was cut over in a pattern of alternate clear-cut strips, the annual runoff yield was immediately increased by 31 per cent. Complete cutting, but not removal, of vegetation in experimental basins in Coweeta, North Carolina, was reported by Lieberman and Hoover (1951) to have resulted in a 100 per cent increase in the median basin runoff yield. Most of this increased yield, however, was in the form of higher base flows resulting from the reduction in interception and evapotranspiration; infiltration rates were likely unaffected because the cutover material remained in place as ground protection. Figure 2.27A shows the marked increase in groundwater levels that were observed after clearcutting of the beech forest in a Danish experimental basin. The low summer and early fall water levels before clearcutting reflects the soil-water loss by evapotranspiration. After clearcutting this water is no longer withdrawn and groundwater levels were much less variable (Holstenet-Jorgensen, 1967; see also Thomas and Benson, 1970).

Provided that forest regeneration is permitted, the hydrologic changes following logging operations are only temporary. For example, in the experiments at Fraser, Colorado (Goddell, 1958, 1967), although the initial increase in runoff yield was 31 per cent during the first two years after cutting, a further 5 years of forest regeneration reduced the runoff yield to 22 per cent above the pre-cut conditions. Similar patterns of recovery have been described by many other authors (see in Sopper and Lull, 1967).

Fire is the most common natural cause of widespread damage to vegetation. A typical runoff response to a complete and intense forest fire is a marked increase in the magnitude of the peak flood, a decrease in the lag-time and time of rise of the hydrograph, and steepening of the recession curve (for example, see Brown, 1972). In cases of fired vegetation, the effects of reduced infiltration rates are often intensified by the increased rates of overland flow possible in the absence of the otherwise obstructing grass and herb layer.

In cases where the basin vegetation is not completely burned, the hydrograph response is dependent on the location and relative area of the burn. Figure 2.27B shows the unusual hydrologic response to a local burn close to the gauging station. The first and smaller flood peak consists of runoff derived from the burned area, and the main flood peak is formed by runoff from the rest of the basin.

Vegetation regeneration after a forest fire produces the same recovery characteristics in the flood hydrograph that are observed after logging. For example, in the Brown (1972) study of the Yarrongobilly River in eastern Australia, the fire related secondary peak in the flood hydrograph (see Figure 2.27B) had disappeared after five years of vegetation recovery.

The general influence of modern logging practices and forest fires on the character of the flood hydrograph is summarised in Figure 2.27c.

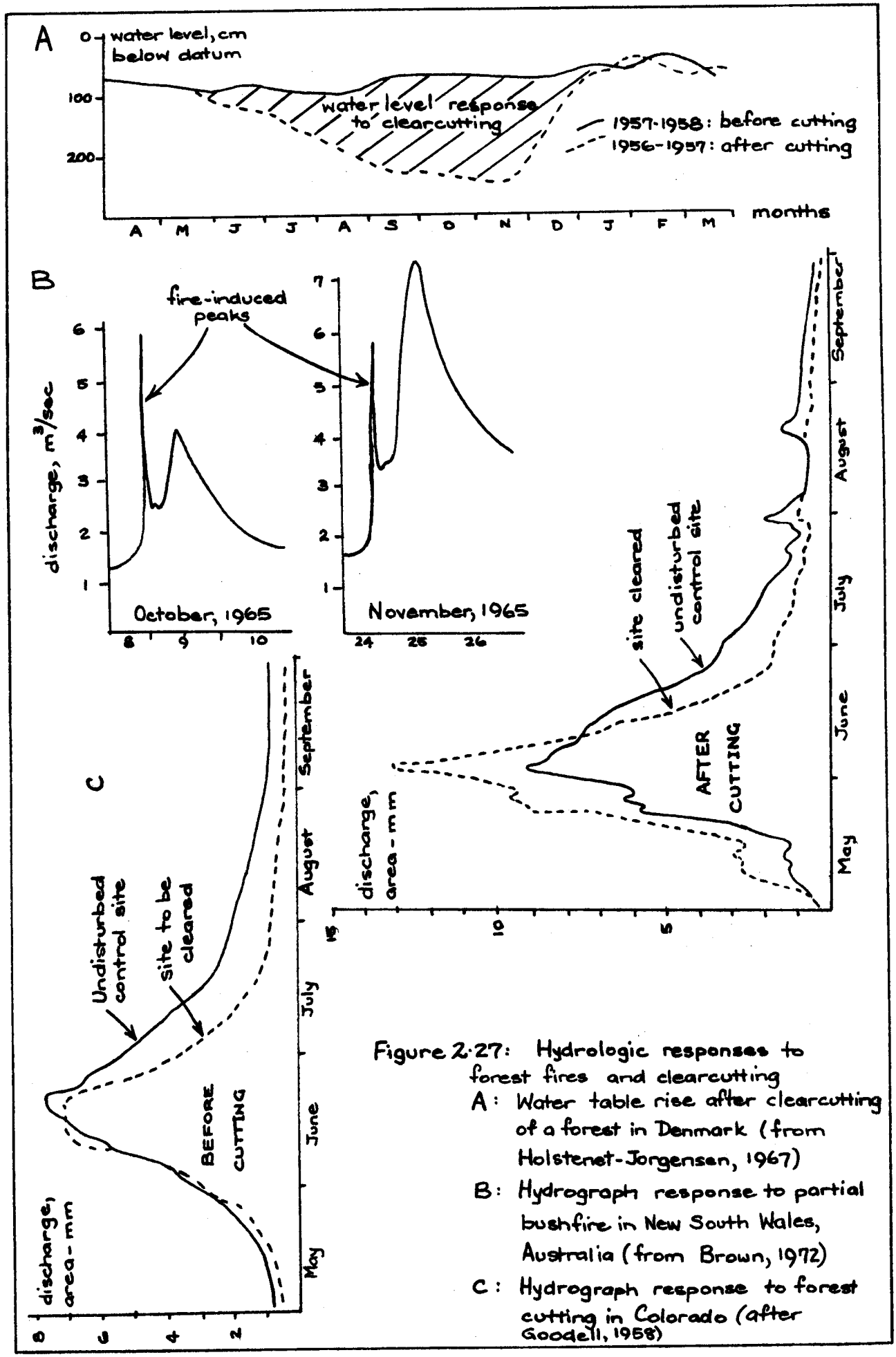


Figure 2·27: Hydrologic responses to forest fires and clearcutting

A: Water table rise after clearcutting of a forest in Denmark (from Holstenet-Jorgensen, 1967)

B: Hydrograph response to partial bushfire in New South Wales, Australia (from Brown, 1972)

C: Hydrograph response to forest cutting in Colorado (after Goodell, 1958)

2.8 (c) Runoff transfer conditions

The final group of flood hydrograph controls listed in Table 2.6, the runoff transfer conditions, includes all aspects of basin and channel geometry which influence the volume and delivery rate of water reaching the gauging station.

The total volume of water constituting the basin input from a storm is the simple product of the average precipitation and the basin area. If other factors remain constant, it is obvious that the flood peak will increase as the size of the basin increases. Figure 2.28A shows typical graphs relating mean annual flood (Q) to drainage area (A_d) of the form

$$Q = a' A_d^{b} \qquad (2.51)$$

in which b is commonly between 0.5 and 1.0 and a' depends largely on regional topographic and precipitation characteristics. Equation (2.51) forms the basis of the many so-called rational flood formulae (see Gray and Wigham, 1970) three examples of which are shown as equations (2.52) to (2.54) in Table 2.7. We might expect that Q should be a linear function of A_d but equation (2.51) is simply a statistical statement and of course other factors are not constant in such an empirical relationship. The exponent b, at less than unity, implies that the flood discharge per unit drainage area declines as the size of the basin increases ($dQ/dA_d = ba'A_d^{b-1}$ where $b-1 \doteq -0.50$); see Figure 2.28B. This decline is partly because small basins are usually steeper and often have less well vegetated slopes and a smaller depth of moisture-storing soil than larger basins.

Because drainage area is geometrically related to most linear aspects of the basin planform (see Part I), flood discharge peaks also tend to increase as river length increases if other factors remain constant (see Morisawa, 1967). Although river length (or basin length) is rarely preferred over basin area as an independent variable in equations to predict flood peaks, it is commonly used as a predictor of flood hydrograph time of rise and basin lag-time. Clearly, as the size of the basin increases, so does the average distance (and travel time) from the watershed to the basin mouth. These relationships are reflected in many empirical formulae such as those expressed by equations (2.55) and (2.56) in Table 2.7 (also see Taylor & Schwartz, 1952). Furthermore, because the amplititude/wavelength of a flood wave tends to decline as the wave travels downstream, the time base of the flood hydrograph also tends to increase as channel length increases (see equation 2.57 in Table 2.7).

The relationship between flood hydrograph form and drainage basin shape is shown schemmatically in Figure 2.29A. De Wiest (1965) suggests that the hydrograph form will directly reflect the relative basin area within given travel times from the gauging station. Although this correspondence may apply to a completely controlled environment, there is little evidence to indicate the extent to which the normal range of basin shape will influence the form of the hydrograph. On the one hand, basin shape has found application in some hydrograph studies (the inverse of Horton's form factor - see Part I) conducted by the U.S. Army Corps of Engineers (U.S.C.E.,1949-1954). On the other hand, a subsequent investigation of 25 basins in the Appalachian Plateau by Morisawa (1958) yielded rather inconclusive results about the relation of basin shape to flood response. On the basis of Morisawa's results,

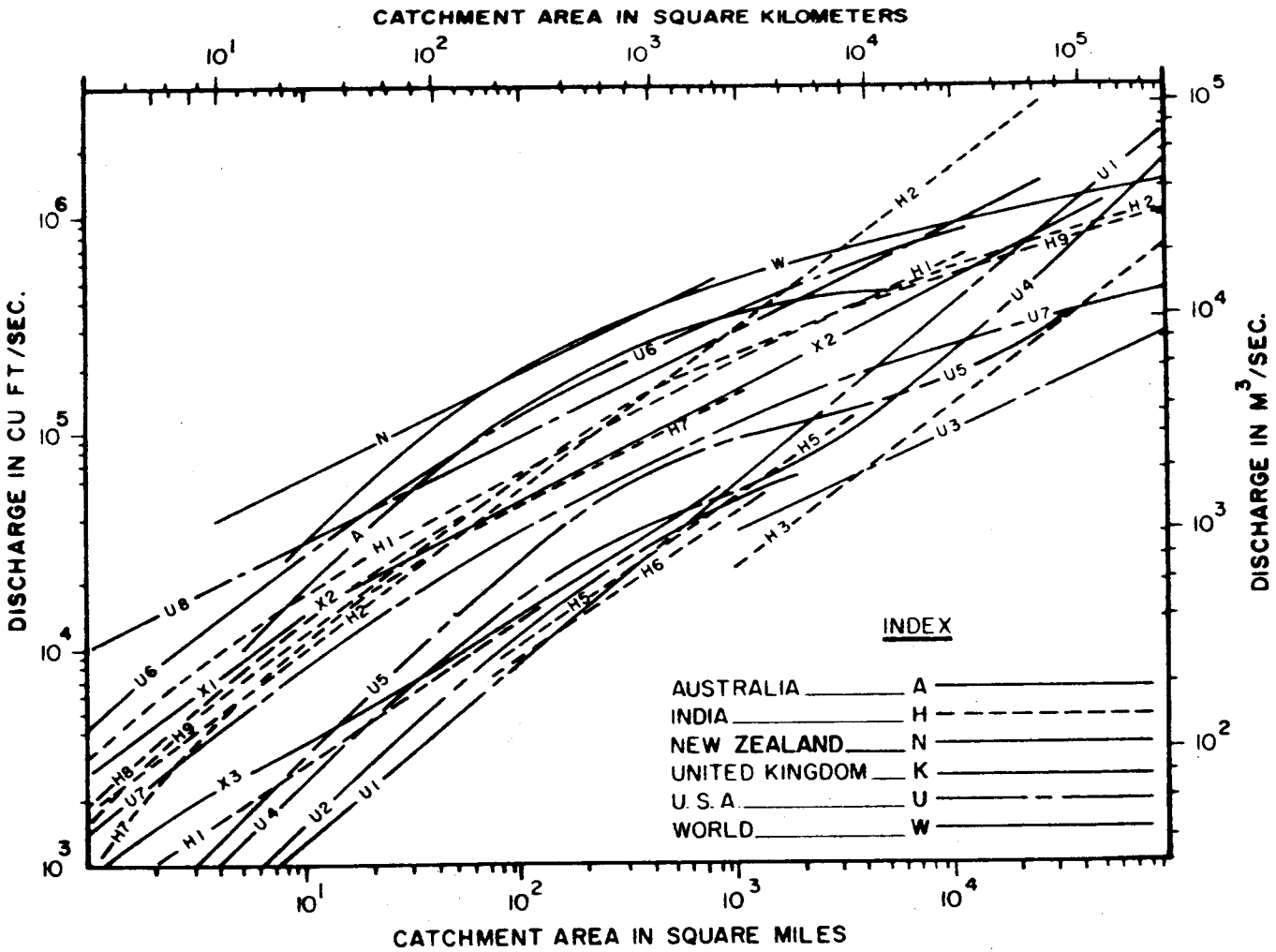


Figure 2·28A: Empirical flood formulae in common use in various countries (from Gray and Wigham, 1970)

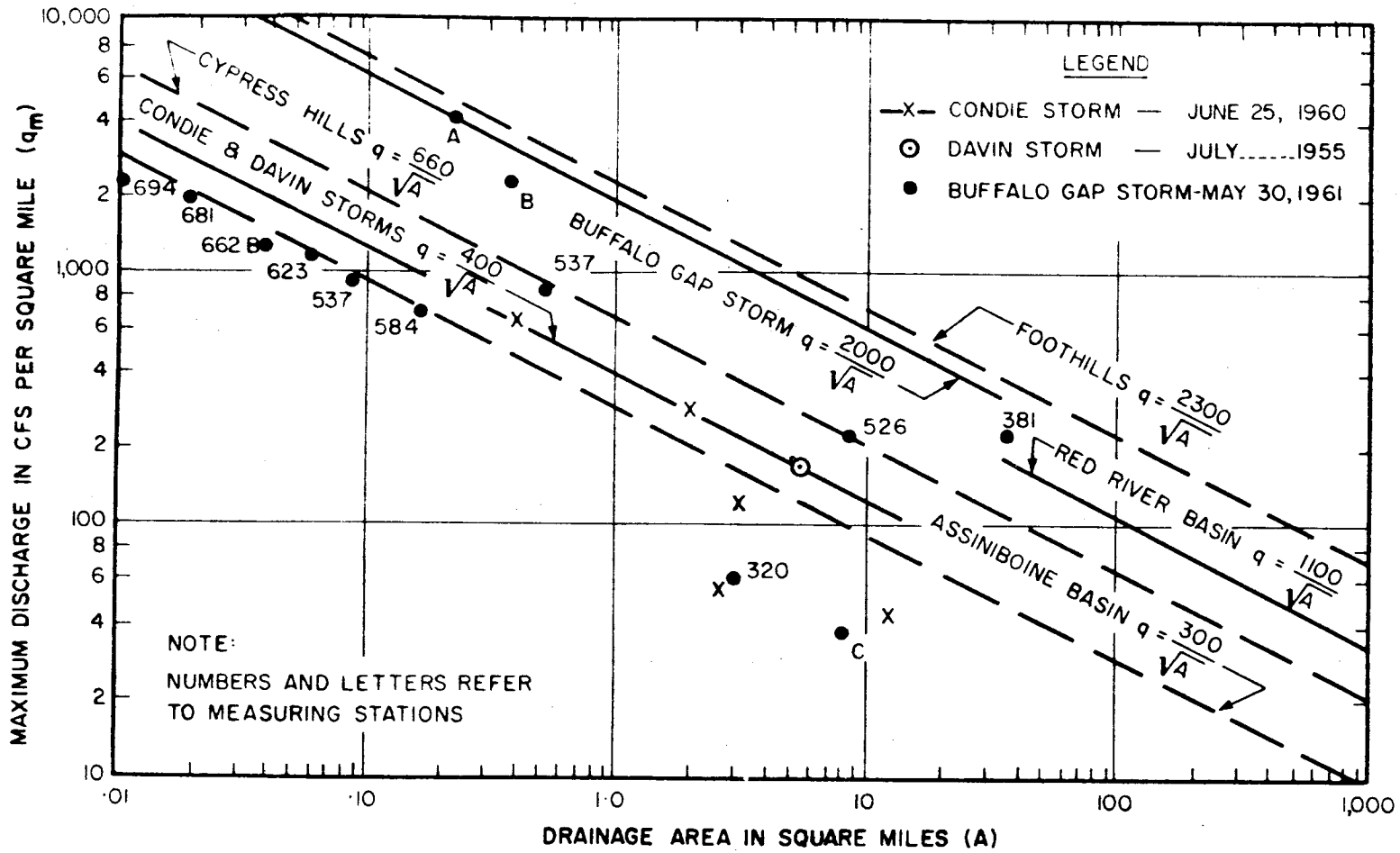


Figure 2·28B: Envelope curves of extreme floods per unit drainage area on the Canadian Praries (from Gray and Wigham, 1970)

Strahler (1964) concluded that controls other than basin planform dominate the hydrologic characteristics of the basin. Although it is likely that flood hydrographs reflect extremes in basin shape such as that depicted in Figure 2.29c, there is a need for further research to confirm this expectation.

In contrast to the basin shape factors, basin relief and slope have a demonstrated close relationship with the form of the hydrograph. The general relationship, shown schemmatically in Figure 2.29B, is that, as basin slopes increase, runoff becomes more rapid and the flood peak increases while the basin lag and time of rise decreases, and the recession curve steepens. These changes are, of course, consistent with the popular impression of flash flooding in mountainous terrain. As in so many natural systems, it is virtually impossible to untangle the interdependence problems in this case. Certainly part of the reason for the strong dependence of hydrograph characteristics on basin slope is the fact that slopes vary in a systematic way with changes in drainage area and stream length, distance from the watershed, elevation, channel slope, in soil depth and infiltration rates, etc. In this sense slope also behaves as a surrogate variable for many other factors influencing the form of the flood hydrograph and it forms an important component in many equations designed to predict flood peaks. For example, Wong (1963) found fom a factor analysis of a number of variables that stream length and average basin-slope combined to provide the best estimate of mean annual flood in many New England drainage basins. This and other flood predicting equations based on basin slopes, are shown in Table 2.7 (equations 2.55 and 2.58 to 2.60).

The aspect and roughness of a drainage basin surface will also influence the character of the flood hydrograph. Aspect is sometimes important because it influences the radiation balance within the basin. For example, the rate of snowmelt and evapotranspiration on shaded slopes is considerably less than those on slopes in receipt of direct solar radiation. The orientation of the basin with respect to prevailing winds will also influence local airflow, and thus patterns of snow accumulation and evapotranspiration rates.

The roughness of the basin surface will also likely control the hydrograph form through its influence on overland flow rates and on surface storage. Although we can deduce that basic lag and time of rise of the flood hydrograph will increase as basin surface roughness becomes greater, the deduction has never been tested because of the difficulty of isolating the influence of this property from all others.

We have seen in Part I that stream order is related to stream length and to drainage area. It follows from the preceding discussion that flood hydrograph characteristics will also vary with stream order. For example, we would expect a first order stream to have a hydrograph which is strongly peaked with a relatively small basin lag and time of rise; the hydrograph of a higher order stream will have a large flood peak but a more rounded general form.

Stream order is not used in predictive models of flood hydrograph characteristics because this ordinal scaling of basin size contains less information than the interval scales of direct length, area, and slope

Source	Equation	
Jarvis (1942)	$Q = 100pA_d^{0.5}$	(2.52)
Forsaith (1949)	$Q_{tr} = C(32.3A_d^{0.5}t_r^{0.44})$	(2.53)
Leopold and Miller (1956)	$Q_{2.33} = 12A_d^{0.79}$	(2.54)
Kirpich (1940)	$t_c = 0.0078L_o^{0.77}S_o^{-0.39}$	(2.55)
Snyder (1938)	$t_L = C_t(LL_{ca})^{0.30}$	(2.56)
Snyder (1938)	$t_b = 3 + 3(t_L/24)$	(2.57)
Wong (1963)	$Q_{2.33} = 0.096L^{1.29}S_a^{0.97}$	(2.58)
Nash and Shaw (1966)	$\bar{Q} = 0.074A_d^{0.75}S_a$	(2.59)
Hickok et al (1959)	$t_L = 23\ (L_{csa} + W_{sa})/S_a\ D_d^{\ 0.65}$	(2.60)
Carlston (1963)	$Q_{2.33}/A_d = 1.3D_d^2$	(2.61)
Manning (1891)	$Q = 1.49R_h^{2/3}S^{1/2}A/n$	(2.62)
Potter (1953)	$\log Q_{10} = 0.17 \log A_d - 0.55 \log T + 0.93 \log P + 0.45 \log S_f - 1.4$	(2.63)
Rodda (1963)	$\log Q_{2.33} = 0.77 \log A_d + 2.92 \log R_{2.33} + \log D_d + 1.08$	(2.64)

Q=discharge(ft^3/sec); Q_{tr}=peak flood with t_r recurrence interval (ft^3/sec); $Q_{2.33}$=mean annual flood (ft^3/sec); Q_{10}=peak flood with 10 year recurrence interval (ft^3/sec); $\bar{Q}$=mean annual maximum discharge (ft^3/sec); A_d=drainage area (miles^2,acres in equation 2.63); t_b=hydrograph time base (hrs.); t_c=time of rise (min.); t_r=recurrence interval (years); t_L=basin lag-time (hrs); L_o= maximum length of travel of water (ft); C_t=coefficient (ranging from 1.8 to 2.2); L=length of main stream from divide to outlet (miles); L_{ca}=distance from outlet to a point nearest the centre of the basin (miles); S=slope; S_o=slope equal to H/L where H is the height difference (ft) between the basin outlet and the most remote point on the divide; S_a=average slope (as % for source area in equation 2.60); L_{csa}=length from the basin outlet to the centre of gravity of the source area (ft), source area is that half of the basin with the highest average slope; W_{sa}=average width of source area (ft); D_d=drainage density (miles/miles^2, ft/acre in equation 2.60);R_h= hydraulic radius (cross-sectional area, A/channel perimeter); n= Manning's roughness factor; T=topographic factor; P=rainfall intensity factor; S_f=rainfall frequency factor; $R_{2.33}$= mean annual daily rainfall (inches).

Table 2.7: Equations relating flood characteristics to basin properties.

measurements. But two other parameters of the drainage network do contain information not included in length and area measurements of basins: the bifurcation ratio and drainage density. The former is essentially conditioned by the geological structure of the basin and the latter is largely dependent on soil type and on climate (see Melton, 1957).

The relation between flood hydrograph form and the bifurcation ratio (R_b) of the drainage net is shown schemmatically in Figure 2.29c. Clearly, bifurcation ratio and basin shape in general will be closely related. Low bifurcation ratios (and rotund basins) will tend to deliver surface runoff from all parts of the basin to the mouth at about the same time. The resulting hydrograph will be strongly peaked. High bifurcation ratios (and elongated basins) will deliver surface runoff to the basin mouth over a much longer time base. As we noted in the discussion of basin shape, however, these responses will probably be evident only among rather extreme values of R_b.

The fastest natural means of moving water from its point of impact as basin precipitation to the basin mouth, is through the network of open channels. If other factors are constant, the more rapidly water can find its way to a channel, the faster will be its rate of delivery to the basin mouth. It follows that, if these other factors remain constant, as drainage density increases so will the magnitude of the flood peak. This relationship has been demonstrated in a number of basins in North America and in the U.K. (see Figure 2.30A). Furthermore, Hickok et al (1959) has shown that basin lag-time is inversely related to drainage density (see equation 2.60), and Carlston (1963) has shown that flood discharge per unit drainage area increases as drainage density increases (see equation 2.61).

Finally, the within-channel conditions will influence the form of the flood hydrograph by determining the rate of discharge through the channel. We will consider in Part III the factors governing the flowrate in open channels. For the moment we should note that one of the most commonly used expressions for predicting velocity and discharge, the Manning formula (see equation 2.62 in Table 2.7), indicates that channel roughness is a very important control. As the roughness of the channel boundary increases and other factors remain constant, discharge will decline. Peaked flood hydrographs with short basin lags and times of rise are associated with steep, smooth, and hydraulically efficient river channels. This is a case where the distinction between discharge and stage hydrographs becomes important; stage may remain constant but discharge (and velocity) may vary in accordance with the roughness of the channel.

Many of the factors listed in Table 2.6 have been incorporated in multiple regression equations to provide comprehensive regional prediction formulae. These formulae are generally only applicable within the region in which they were developed, testifying to the extreme complexity of the basin hydrologic cycle. Two examples are given in Table 2.7 by equations 2.63 and 2.64

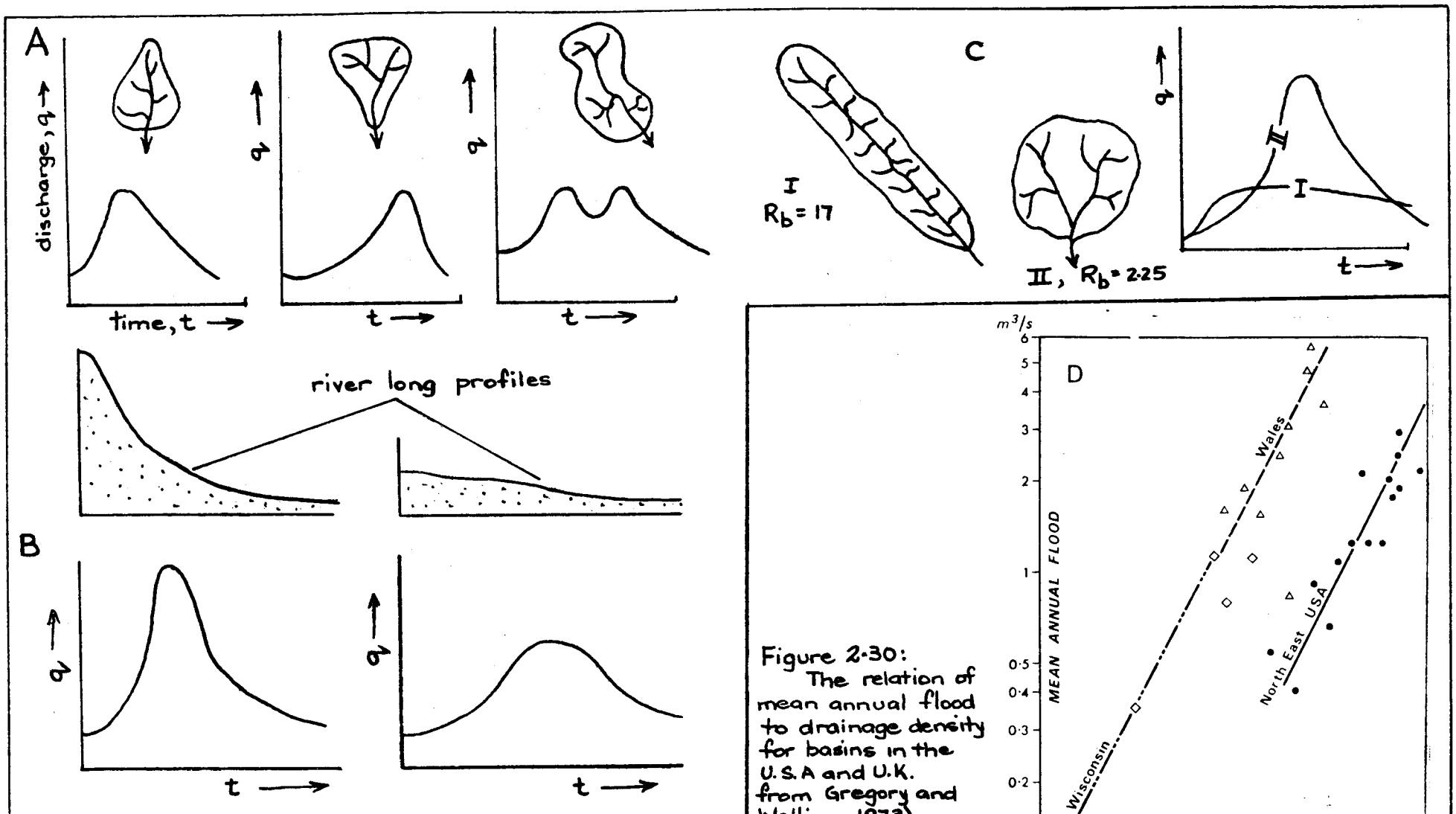


Figure 2·29: Flood hydrograph response to variation in basin shape (A), basin relief (B) and bifurcation ratio (C); respectively after De Wiest (1965), Gregory and Walling (1973) and Strahler (1964).

Figure 2·30: The relation of mean annual flood to drainage density for basins in the U.S.A and U.K. from Gregory and Walling, 1973)

2.9: General models of basin hydrology

The ultimate goal of the many lines of research into the specific hydrologic processes described in the previous sections is the synthesis which will provide a complete picture of the operation of the basin hydrologic cycle. Drainage basin hydrology is so complex, however, that it is difficult to predict the behaviour of any element in response to changes in another without making the unrealistic assumption that the relationship among all other factors will remain unchanged. It is extremely difficult, at an intuitive level, to cope with the complex chain of events set in motion by a simple change in some basin property; the adjustments take place through a complicated process of responses, feedbacks and readjustments, of many interrelated factors. The development of general models in hydrology is an attempt to overcome these problems.

For our purposes, we may recognise three basic types of models: physical, analogue, and digital models (Ward, 1971). Physical models are scaled-down versions of actual or typical basins: the laboratory watershed (for example, see Amorocho and Hart, 1965; Chery, 1966; Eagleson, 1969; Black, 1970; Roberts & Klingeman, 1970). In such models basin inputs and basin properties can be varied and the system response observed. The principal limitations of physical models are the facts that they simplify reality by ignoring detail at certain scales, and that it is impossible to maintain complete dynamic similarity. For example, few physical models faithfully represent basin details such as vegetation structure and soil horizonation. Furthermore, although the physical dimensions of a basin's planform may be modelled at a scale of say, 1/10,000, it is rarely possible, for example, to reduce the grain size of surface material by this ratio without making the model surface completely impermeable. On the other hand, increasing the grain size to simulate reasonable infiltration rates, results in a surface roughness equivalent to that of a boulder-strewn hillside ! Many other properties of soil and flowing water simply will not scale to an appropriate level without violating the principle of similitude.

The increasing popularity of analogue and digital models in recent years has accompanied the development and widespread use of electronic computers. Although analogue models are strictly any mechanical or electrical device that possesses functional characteristics equivalent to those of the system being modelled (Gregory and Walling, 1973), most modern applications take the form of electronic computer models (see Chow, 1964; Tinlin, 1969; Riley & Narayana, 1969). The most widely exploited analogue is that between the flows of electricity and water, in which analogue pairs are potential difference (volts) and hydraulic head (metres), electrical resistance (ohms) and flow resistance, current supply (amperes) and discharge (m^3/sec). The equations describing Ohms' law of an electrical current, steady uniform flow in an open channel, and Darcy's law for flow through porous media, are structurally identical.

Digital models are designed to be operated by general purpose digital computers capable of processing vast amounts of data and cycling repetitive mathematical operations in very short periods of time. Because of the general accessibility to this type of computer,

the digital model is now the most commonly used variety of general model. Indeed most digital models can be operated by the new generation of programmable hand-held calculators. They are essentially a series of mathematical expressions in which the solution to the first provides the input data for the solution of the second, and so on, until the series is solved. The model may be a simple linear series or it may involve many feed-back loops and interactive components. Most digital models have a deterministic framework based on fundamental physical principles (such as continuity and conservation of energy or momentum), although some may contain stochastic elements, and almost all contain calibrated parameters. Examples of stochastic elements in a digital model are the use of random number generators to make certain decisions such as providing the location of a storm centre over a drainage basin, or selecting which trees in a maturing forest will die in a long-term evatranspiration model. Calibration of model parameters is achieved by examining actual input/output data and treating the parameters as coefficients of proportionality. Thus the digital model of this type, although general in structure, can only be applied to specific regions after it has been tested (referred to as "parameter optimisation") in that area or in one similar to it. Once it has been calibrated, however, it becomes a useful tool of prediction and a means of examining the dynamics of the basin hydrology. The simplest way to characterise these general models of basin hydrology, is to consider an example. The example is not difficult to choose because one of the most widely applied digital simulations of runoff, during the last decade, is the Crawford & Linsley (1966) Stanford Watershed Model IV.

2.9 (a) The Stanford Watershed Model IV (S.W.M. IV)

The conceptual framework of this model is illustrated by the flowchart in Figure 2.31. The main inputs to the model are precipitation and potential evapotranspiration; if snowfall is significant the snowmelt subroutine requires inputs of precipitation, maximum and minimum daily temperatures, and incoming shortwave radiation. Snowmelt is calculated from a set of equations similar to those discussed in section 2.3 (d). All of the total precipitation (rainfall and snowmelt) is initially held as interception storage until a specified interception capacity is filled. Interception storage is discounted by evapotranspiration at the potential rate so that interception actually continues during a storm. Once the interception storage is exceeded, the precipitation excess, after allowing for direct runoff to channels from adjacent impervious areas, represents the water input at ground level.

Crawford & Linsley (1966) modelled the infiltration process in terms of upper and lower storage zones representing variable soil-moisture profiles and groundwater conditions. The infiltration capacity, dependent on the lower zone soil-moisture conditions (see Figure 2.32A), determines how precipitation is divided between infiltration and surface detention. The infiltrating water is then transferred to interflow (throughflow) storage and or to lower zone storage and groundwater storage by the empirical functions graphed respectively in Figures 2.32, B and D. The increment to surface detention is divided between overland flow and upper zone storage (depression

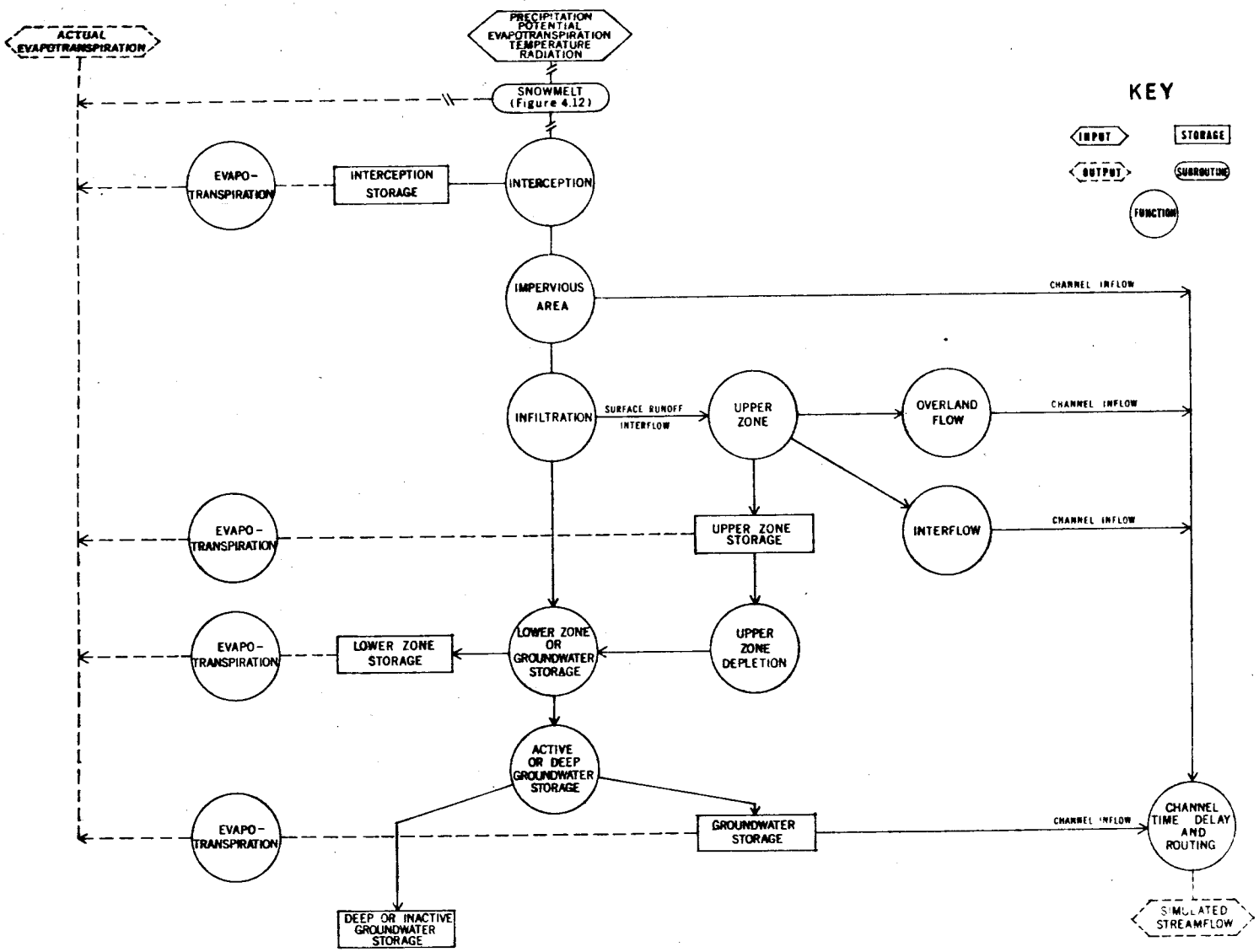


Figure 2.31: The Stanford Watershed Model IV flowchart (from Crawford and Linsley, 1966)

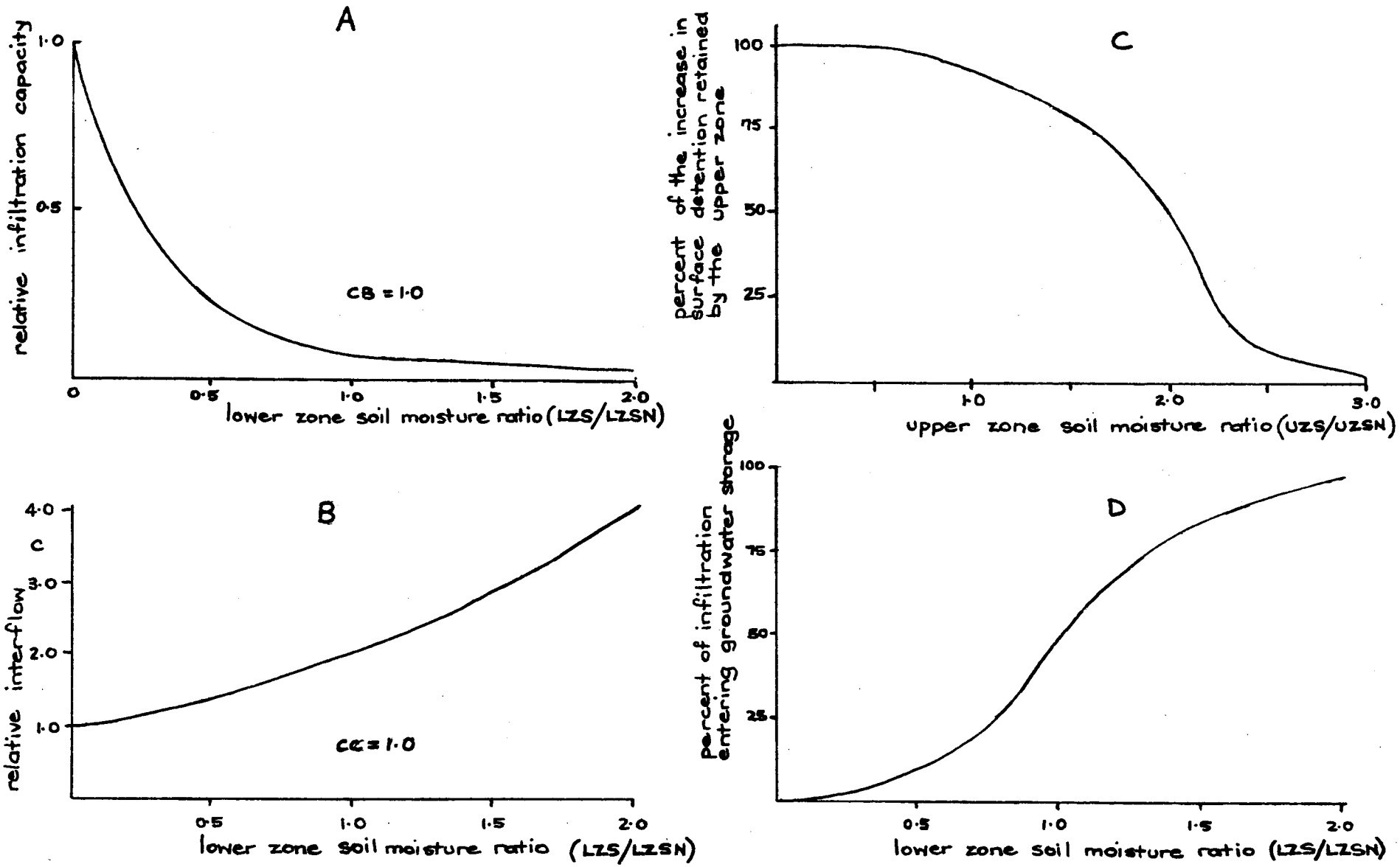


Figure 2·32: The Stanford Watershed Model IV control functions (after Crawford and Linsley, 1966)

storage) in accordance with the empirical function shown in Figure 2.32C. The rate of moisture loss from the upper zone storage by percolation to the lower zone and to groundwater storage is included in the groundwater storage function in Figure 2.32D. Deep or inactive groundwater storage is governed by a simple assigned groundwater bypass fraction. Potential evapotranspiration is satisfied, in order, by evapotranspiration from interception storage and from upper zone storage (both at the potential rate), and finally from the lower zone storage at a rate equal to $(PE - PE^2/2r)$ in which r is the product of the lower zone soil-moisture ratio and an input parameter. Finally, inputs from the overland flow component (based on the Manning equation; see Part III), from interflow (based on the recession curve function in equation 2.48, where the recession constant is a variable input parameter), and from groundwater storage (based on Darcy's law in equation 2.49, and on the recession curve function), are added to the direct runoff from impervious areas, to yield a total channel input.

The usual procedure adopted in the S.W.M. IV is to "sum" the input to channel reaches from basin segments in order to determine the total basin runoff. These reach inputs are "translated" to the basin mouth to form a single hydrograph which is then further modified by a flood routing transformation. This final processing, designed to simulate the flood-wave altenuation and increased time lag associated with friction and storage losses as the water moves through the channel system, yields the flood hydrograph in its final form.

The complete set of input parameters controlling the functioning of the very flexible S.W.M. IV, are listed in Table 2.8; several examples of obviously successful hydrograph simulation by the model are shown in Figure 2.33.

2.9 (b) Flood routing models

The final calculations of the flood hydrograph form in the S.W.M. IV are based on one type of a number of flood routing models. Flood routing is the prediction of hydrograph form in a reach of channel based on a known hydrograph at some upstream point. Most flood routing procedures may be assigned to one of two somewhat related categories: open channel (or streamflow) routing models, or reservoir routing models.

The movement of a flood wave through an open channel is a very complex phenomenon in which flow depths and velocities vary in space and time. Because the equations describing such flows have so far defied exact solution, engineers faced with solving practical problems have developed a variety of approximate numerical solutions. It will be useful to briefly consider an example of each of the two basic types of flood routing models that are commonly used in current engineering practice.

All flood routing models are based on some knowledge of the river reach under consideration. The most common type of data base used is the observed inflow and outflow hydrographs. If these are not available, they are estimated, perhaps from precipitation records and unit hydrograph theory.

Parameters estimated from the physical character of the basin

1. Percentage of impervious area
2. Overland flow slope
3. Overland flow length
4. Manning n for overland flow

Parameters requiring optimisation

1. Interception storage capacity
2. Infiltration index
3. Interflow index
4. Nominal lower zone storage
5. Nominal upper zone storage
6. Interflow recession constant
7. Groundwater recession constant
8. Lower soil evaporation factor
9. Groundwater bypass fraction

Table 2.8: Input parameters governing the operation of the Stanford Watershed Model IV (based on Wood and Sutherland, 1970)

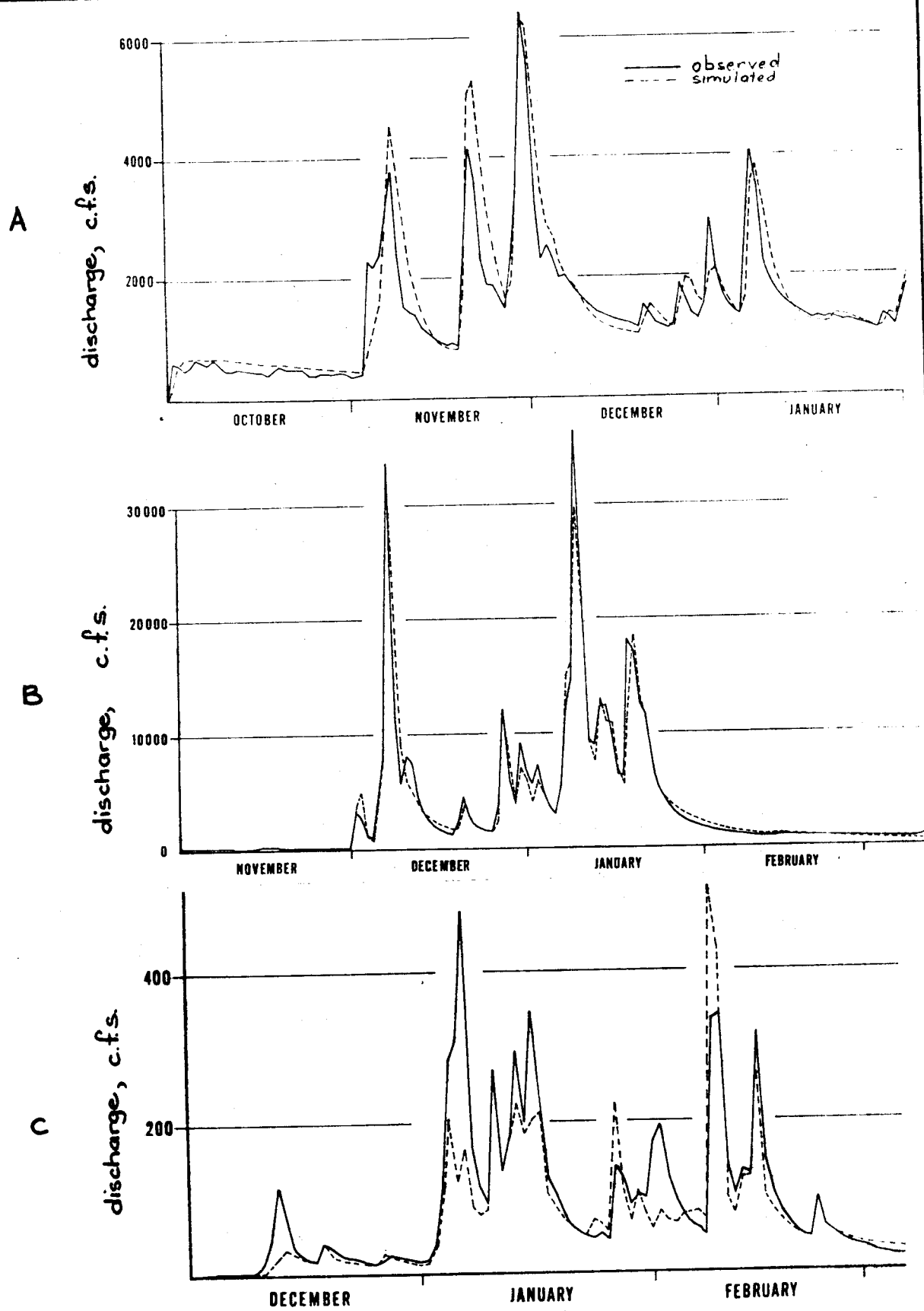


Figure 2.33: Hydrograph simulation by the Stanford Watershed Model IV (from Crawford and Linsley, 1966)
A: French Broad River at Blantyre, 1949
B: Russian River near Healdsburg, 1953
C: Middle Fork Beargrass Creek at Cannons Lane, 1950

The inflow hydrograph is the hydrograph of the total flow into a channel reach in a given period of time; the hydrograph of outflow is that for the total flow leaving the reach in the same period of time. As a flood wave enters the reach the discharge will exceed that leaving it and the amount of water within the reach (storage) increases. Similarly, during the waning flood, storage decreases as outflow exceeds inflow. The difference between the inflow and outflow hydrographs constitutes the storage curve. Storage plotted against outflow, produces a discharge storage curve which forms a hysterisis loop reflecting the changing balance of inflow and outflow during the passage of the flood. The locus of corresponding abscissa midpoints of the loops is the average storage curve. These relationships are shown in Figure 2.34.

Thus we can say that, in a given time interval, the difference between inflow (I) and outflow (O) is equal to the change in storage (ΔST), or

$$I - O = \Delta ST \qquad (2.65)$$

The selection of a time interval (Δt) or routing period, is a very important step in this analysis. It is the time interval at which the ordinates of the hydrograph used in routing are represented. It must be short enough to adequately represent the form of the hydrograph. Discharge variation is assumed to be linear over the routing period so that equation 2.65 can be expressed in finite time intervals as

$$\tfrac{1}{2}(I_1 + I_2)\Delta t - \tfrac{1}{2}(O_1 + O_2)\Delta t = ST_2 - ST_1 \qquad (2.66)$$

or

$$\tfrac{1}{2}(I_1 + I_2)\Delta t + ST_1 - \tfrac{1}{2}O_1\Delta t = ST_2 + \tfrac{1}{2}O_2\Delta t \qquad (2.67)$$

where the subscripts refer to the routing periods. The values on the left of equation 2.67 are known from observed hydrographs and can be used to calculate $ST_2 + \frac{1}{2}O_2\Delta t$.

Although equation 2.67 can be solved using the average discharge-storage curve (see Puls, 1928), the solution ignores the variable water slope that occurs during the passage of a flood wave (the hyster sis storage effect) and consequently predicts a poor approximation of channel routing (Lawler, 1964). A far more realistic view of the routing process considers the channel storage as two parts: prism storage and wedge storage. The former is the water below an arbitrary line drawn parallel to the bed, and the wedge storage is that above it (see Figure 2.35). Prism storage is a simple function of outflow, KO, and wedge storage is represented by a function of the difference in inflow and outflow, KX (I-O). Thus, total storage is

$$ST = KO + KX(I - O) \qquad (2.68)$$

in which K is an empirical storage coefficient (a measure of travel time through the reach) and X is an input parameter to account for the relative effect on storage of inflow and outflow. Equation 2.68 is known as the Muskingum equation, named for the Conservancy District in the Connecticut River Valley where it was first developed (U.S.C.E., 1936; Carter and Godfrey, 1960).

Equation 2.68 can be rewritten in terms of routing periods as

$$ST_2 - ST_1 = K\left[X(I_2 - I_1) + (1 - X)(O_2 - O_1)\right] \qquad (2.69)$$

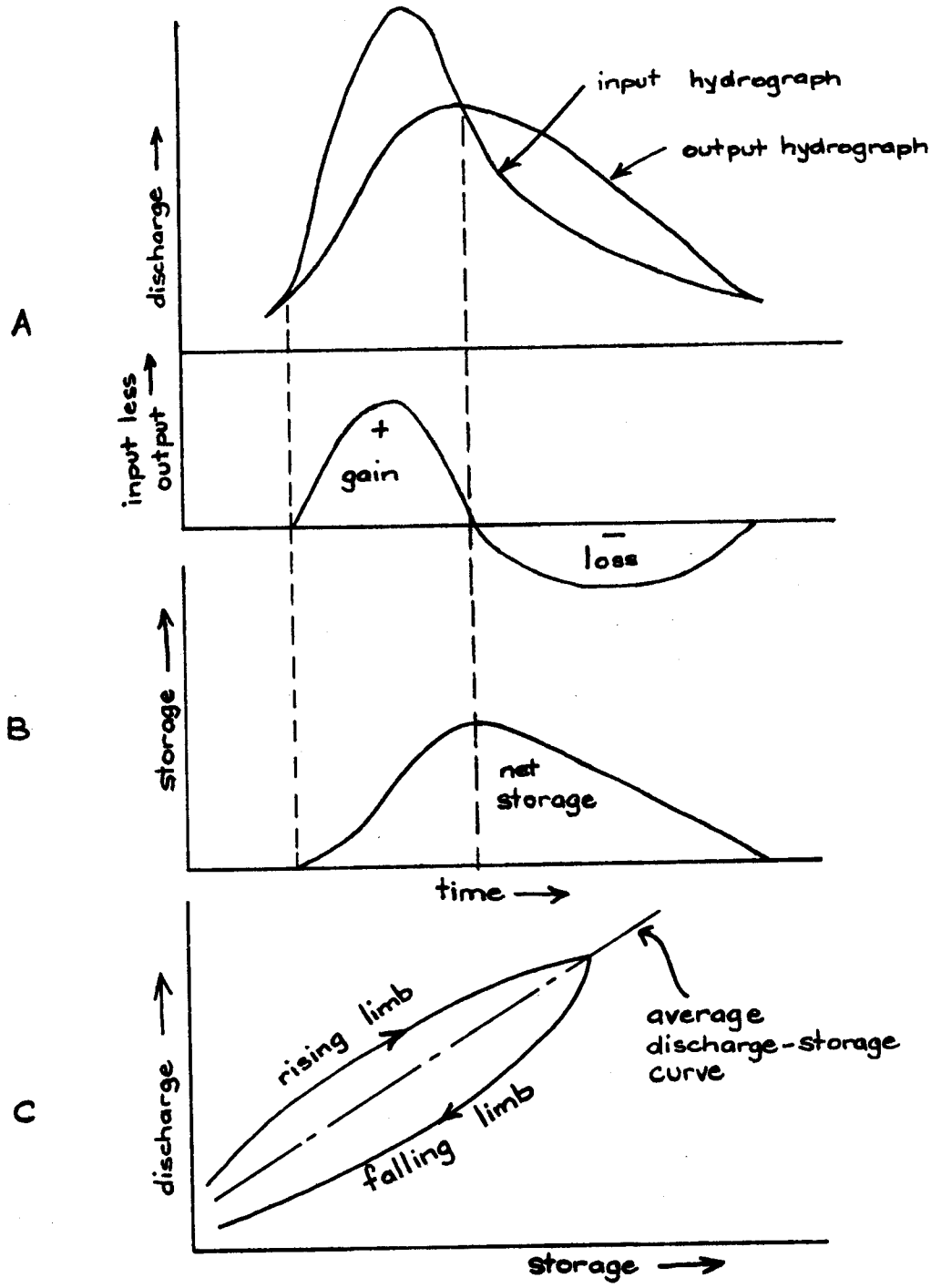


Figure 2·34: Relations among inflow and outflow hydrographs and channel storage.
A: Inflow and outflow hydrographs
B: Storage curve
C: Discharge-storage curve.

Equation .69 provides the additional relationship between O_2 and ST_2 needed to solve equation 2.67; the latter can now be written as

$$\tfrac{1}{2}(I_1+I_2) - \tfrac{1}{2}(O_1+O_2) = \frac{K}{\Delta t}\left[XI_2+(1-X)O_2\right] - \frac{K}{\Delta t}\left[XI_1+(1-X)O_1\right] \qquad (2.70)$$

Simplifying equation 2.70,

$$O_2 = C_1I_2 + C_2I_1 + C_3I_1 \qquad (2.71)$$

in which

$$C_1 = \frac{\Delta t - 2KX}{2K(1 - X) + \Delta t} \qquad (2.72)$$

$$C_2 = \frac{\Delta t + 2KX}{2K(1 - X) + \Delta t} \qquad (2.73)$$

$$C_3 = \frac{2K(1 - X) - \Delta t}{2K(1 - X) + \Delta t} \qquad (2.74)$$

The value of X can be determined by the parameter optimisation procedure outlined in the dicussion of the Stanford Watershed Model. Using a known pair of inflow and outflow hydrographs, the constant K is taken as the reciprocal of the slope of the average discharge-storage curve (see Figure 2.34C) and X is the value which yields the flood hydrograph which best matches the actual hydrograph. A second and more common approach to the determination of X involves solving equation 2.70 for K, where

$$K = \frac{\tfrac{1}{2}\Delta t\left[(I_2 + I_1) - (O_2 + O_1)\right]}{X(I_2 - I_1) + (1 - X)(O_2 - O_1)} \qquad (2.75)$$

Successive values of the numerator (the storage increment) and the denominator (the weighted flow increment) are computed from a known pair of inflow and outflow hydrographs for a range of X values. The computed values of accumulated numerator and denominator are then plotted, the latter against the former, producing hysteresis loops such as those shown in Figure 2.36. The assumed value of X yielding the loop closest to a straight line is taken as the correct value and that of K is given by the reciprocal slope of the straight line. A detailed example of flood routing by this Muskingum method is provided in the illustrative problems in section 2.12.

Flood routing through a large reservoir is far simpler than open-channel routing because outflow can be expressed as a single-valued function of storage (i.e., there is no hysteresis effect caused by a pronounced storage wedge). Routing through a reservoir can be accomplished by substituting in the left-hand side of equation 2.67 the appropriate values derived from the inflow hydrograph to be routed. The required value of O_2 can then be determined from the relationship between O_2 and $ST + \tfrac{1}{2}O_2\Delta t$. Figure 2.37 shows the curves $ST - \tfrac{1}{2}O\Delta t$ and $ST + \tfrac{1}{2}O\Delta t$ with a mean reservoir-storage curve derived in the same way as that shown in Figure 2.34; the outer curves are derived by respective subtraction from, and addition to, the abscissa

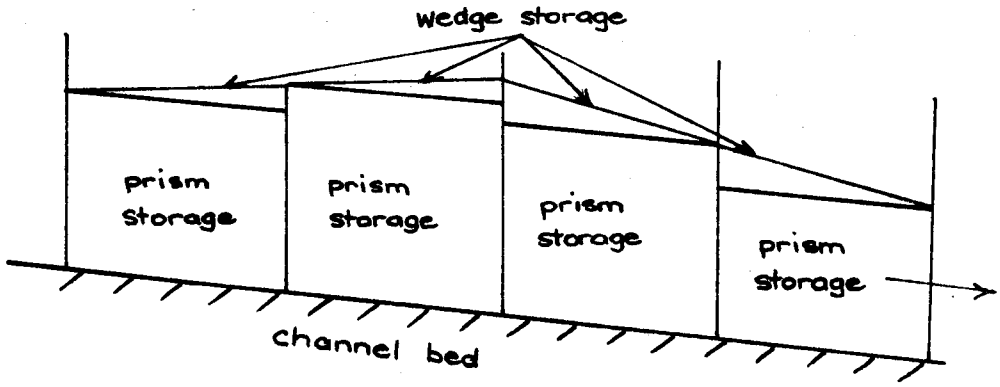


Figure 2·35: The concept of wedge and prism storage

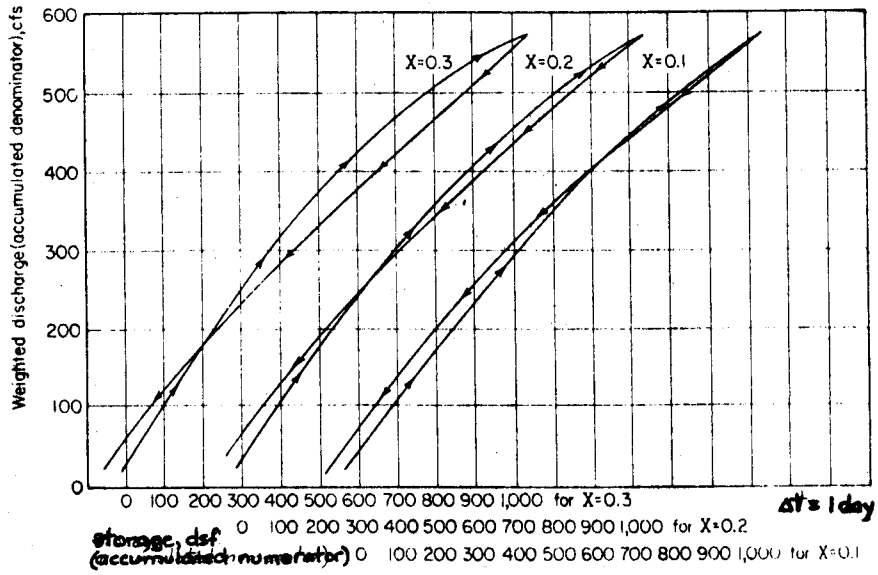


Figure 2·36:

Hysterisis loops of the discharge-storage relation. (see text)

After Lawler(1964)

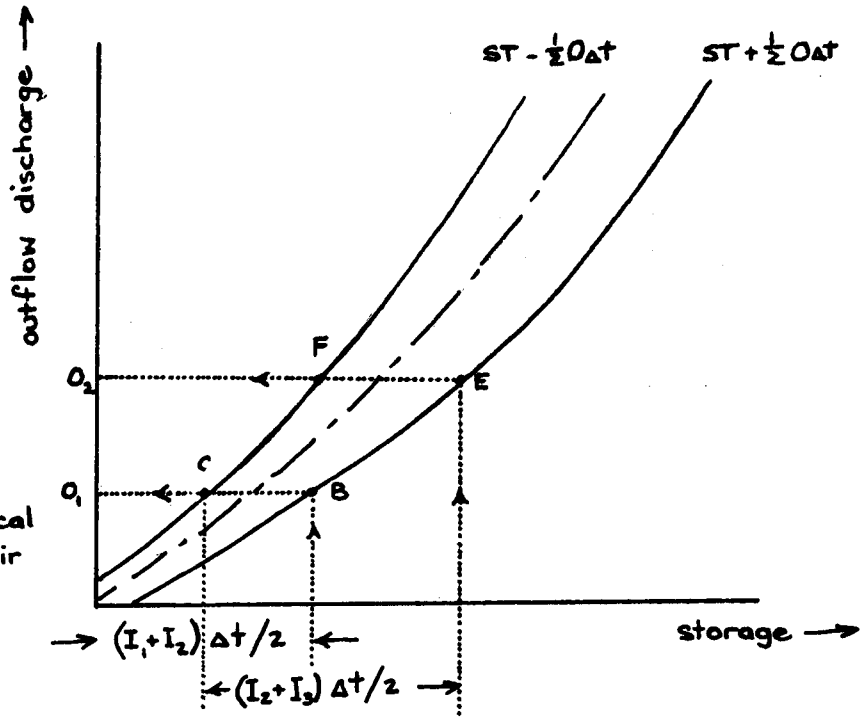


Figure 2·37: Graphical solution of reservoir routing (see text)

of the storage curve, the quantity $\frac{1}{2}O\Delta t$. The routing procedure is shown graphically in Figure 2.37. The storage increment derived from the known inflow hydrograph for routing period 1, $(I_1 + I_2)\Delta t/2$, is entered on the absicssa and the intersection with the ST + $O\Delta t/2$ curve (at B) is read as an outflow discharge on the ordinate. This discharge value (O_1) is the predicted outflow for routing period 1. Point C on the ST-$O\Delta t/2$ curve in Figure 2.37 corresponds to the storage at the end of the first routing period. The storage increment for the second routing period, $(I_2+I_3)\Delta t/2$, is added to the abscissa at C to give a new intersection with the ST $+O\Delta t/2$ curve at E, corresponding with a second period outflow, O_2. The storage at the end of period 2 is given by point F on the ST - $O\Delta t/2$ curve. This procedure is repeated for the remaining periods until the outflow hydrograph is completed. Details of the equivalent numerical routing procedure are given in section 2.12.

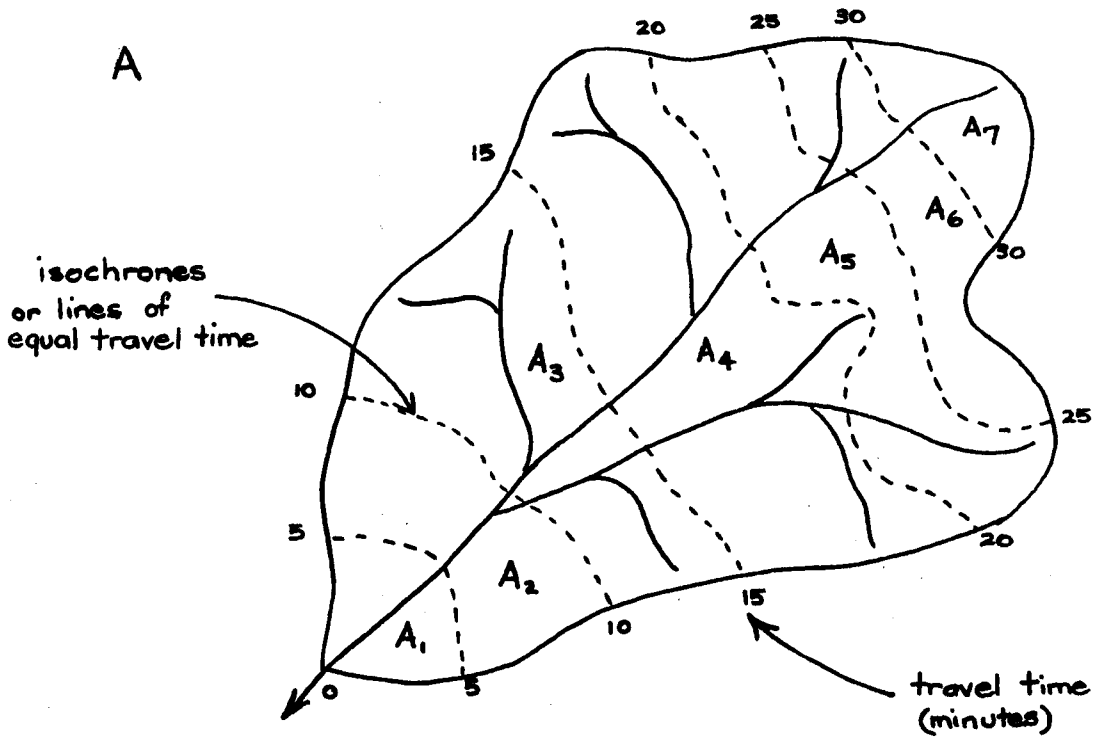
Although reservoir routing is generally of less concern to the fluvial geomorphologist than it is to the civil engineer, it has been used to generate synthetic streamflow hydrographs. Perhaps the most notable contribution to the theory of synthetic hydrographs based on flood routing techniques, is that of Clark (1945).

The theory is based on the fact that surface storm runoff is modified by two factors: the translation or travel time of the surface runoff as overland and channel flow, and storage effects. Although both factors occur similtaneously in nature, Clark argued that they could be treated sequentially. The translation effects are modelled by assuming that, if there is no storage modifications, then the runoff hydrograph will closely resemble the time-area curve of the basin. The time-area curve is obtained by plotting the area of basin segments against their distance (relative travel time) from the basin outlet (see Figure 2.38). The basin segments are selected so that rain falling within any given one will have sensibly equal travel distances. The conversion of travel distance to absolute travel time is accomplished by hydrograph analysis (of basin lag), or on the basis of emperical formulae (see section 2.8c). The hydrograph so formed is the theoretical instantaneous hydrograph from a basin with no storage. It is then routed through reservoir storage equivalent to that in the channel reach in order to simulate storage attenuation of the flood wave.

Although Clark (1945) was the first to show how the area-time curve of the basin, and reservoir routing, could be combined to predict basin outflow hydrographs, there have been many subsequent applications of this type. Indeed, the translation and flood routing functions in the Stanford Watershed Model IV are little modified versions of Clark's original procedures.

Before leaving this topic we should note that there are, of course, many methods of flood routing other than the two examples I have selected as illustrations of the general concept. Information about these other flood routing models can be found in most texts in engineering hydraulics and hydrology (for example, see Gilcrest, 1950; Chow, 1959; Lawler, 1964). Somewhat more advanced models are presented by several authors in the proceedings of the "Rivers 76" Symposium of the American Society of Civil Engineers.

A



time class	relative area	
0 - 5	0·04	A_1
5 - 10	0·13	A_2
10 - 15	0·18	A_3
15 - 20	0·26	A_4
20 - 25	0·18	A_5
25 - 30	0·15	A_6
30 - 35	0·08	A_7

B.

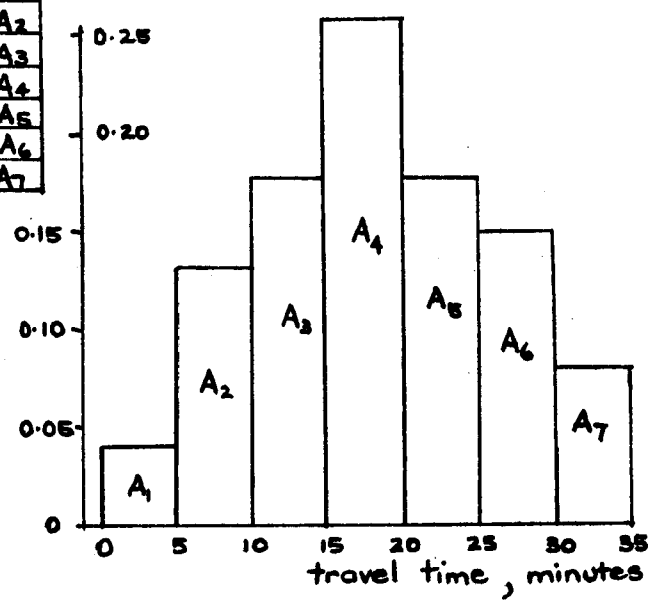


Figure 2·38: A simple translation model and synthetic hydrograph
A: An isochronal map of a basin
B: Time-area curve based on the isochronal map.

2.10: River regimes - the long-period hydrograph

At the start of this chapter (in section 2.2) I set the scene by placing the discussion in the context of the flood hydrograph. Most of this discussion, however, also applies at the longer time scale of the weekly, monthly, or annual hydrograph; I will leave you to make the extension of ideas.

The geomorphologist concerned with average river-channel responses over long periods of time may find the long term record much more useful than the character of individual flood events. A common way of representing this longer record is to display it as a seasonal march of average streamflow. The annual hydrograph, averaged over many years is referred to as the river regime. It is usually based on mean monthly discharge, although daily flows may be utilised where data are stored on cards or on magnetic tape and thus can be readily processed by machine.

Climate exerts a direct control on river regime through the seasonal march of the water balance (precipitation minus evapotranspiration). Because river regimes are regionally dominated by climate, they can be classified and mapped in a manner similar to that used to generalise climatic variation. This generalisation ranges from the simple mapping of representative regime hydrographs, as in Figure 2.39, to the construction and mapping of flow regime classification systems such as that briefly described below.

2.10 (a) Regime hydrograph classification

A recent example of regime bydrograph classification was provided by Beckinsale (1969). He adapted the nomenclature of the Koppen climatic classification to construct the river regime classification outlined in Table 2.9 and mapped on a world scale in Figure 2.41; examples of the main river regime types are shown in Figure 2.40.

Beckinsale essentially divided river regimes into those normally found in the moist tropics (A), in deserts (B), in temperate humid areas (C), in cold moist climates (D), and in high altitude zones influenced by snow and ice (H). He further classified them into the minor divisions in Table 2.9 on the basis of the length and amount of low flow, and of whether a single or double flow peak occurs in the winter or summer season.

Broad classifications of this type (also see Guilcher, 1965) are useful at a regional level of appreciation but more detailed hydrologic information is usually required by the fluvial geomorphologist working at the scale of a channel reach. Although more detailed river regime classifications are available for some areas (for example, see Ledger, 1964), they lack the precise quantitative data often required for research purposes. This type of data is usually obtained from the statistical analysis of the annual hydrographs of record.

2.10 (b) Descriptive statistics of streamflows

It is beyond the scope of this discussion to examine the fundamentals of descriptive and inferential statistics and the reader who is unfamiliar with these concepts should consult a general text such as that by Dixon and Massey (1968) or an analysis oriented to hydrologic data, such as that by Chow (1964).

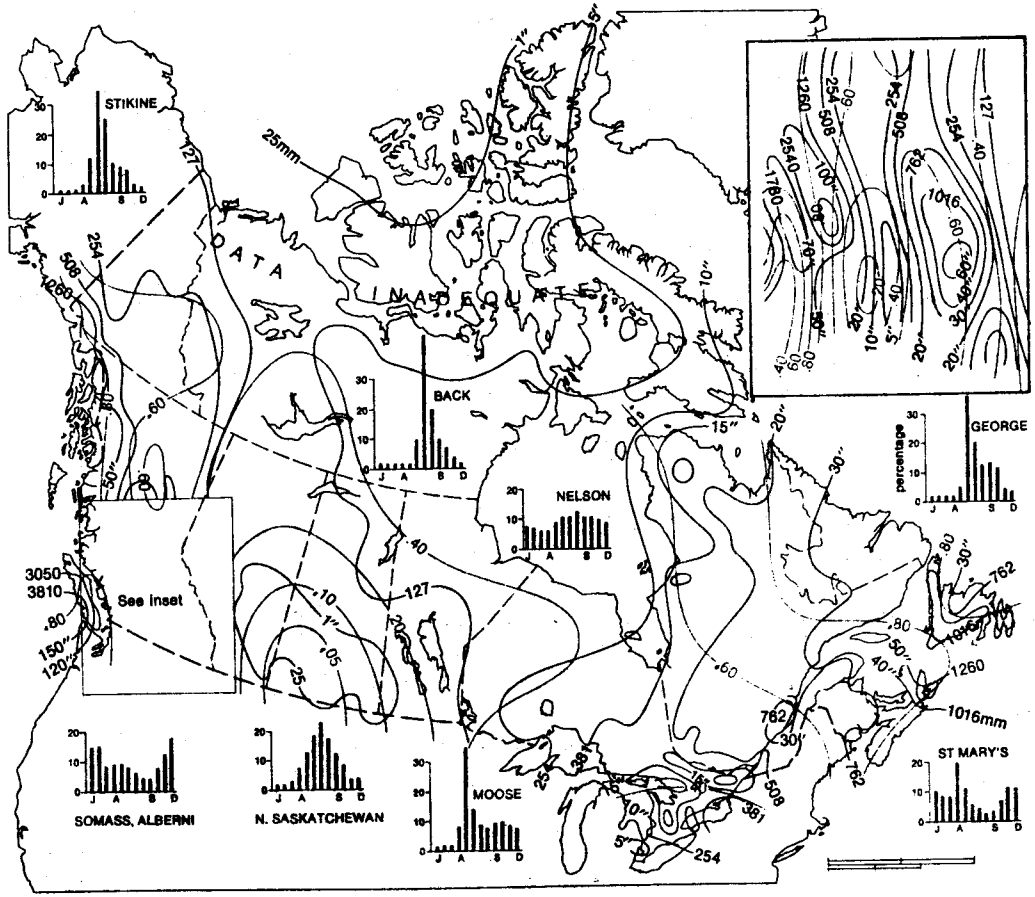


Figure 2·39: Selected river regimes, mean annual runoff (in inches and millimetres) and runoff ratio (values less than 1·0) in Canada. (from Hare and Thomas, 1974)

MAJOR REGIME TYPES	MINOR DIVISIONS	DESCRIPTION
Megathermal flow regimes A	AF	Equatorial double maximum.
	AM	Tropical strong single maximum with a short low-water period.
	AW	Tropical single maximum with a long low-water period.
Mesothermal flow regimes C	CFa	Warm subtropical double maxima.
	CWa	Warm subtropical with strong summer maximum and winter minimum.
	CS	Strong summer minimum.
	CFa/b	All year flow with slight warm season minimum.
	CFaT	All year flow with spring maximum and winter minimum (T = Texas regional type
Microthermal flow regimes D	DFa/DWa	Summer pluvial maximum; winter nival minimum.
	DWb/c	Strong summer pluvial maximum; long winter nival minimum.
	DWd	Strong summer pluvio-nival maximum; prolonged cold season minimum.
	DFa/b	Moderate pluvio-nival or nivo-pluvial maximum; slight summer maximum.
	DFb/c	Strong nival spring maximum, secondary autumn pluvial maximum
	DFc	Violent nival spring maximum, strong winter minimum
Mountain flow regimes H	HN	Regimes dominated by nival or highland snow.
	HG	Regimes dominated by gletscher or highland ice.

Table 2.9: A classification of river regimes according to Beckinsale (1969)

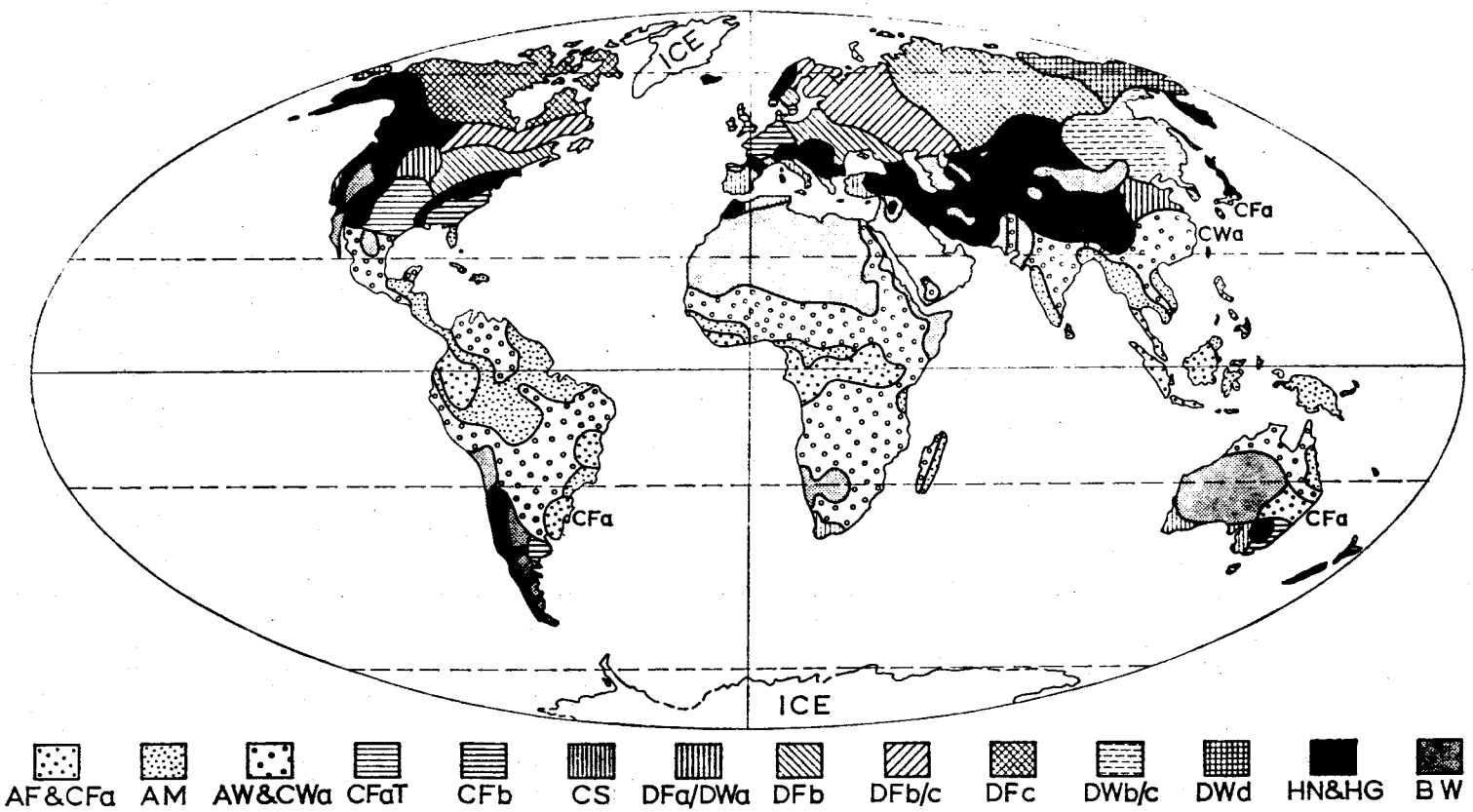
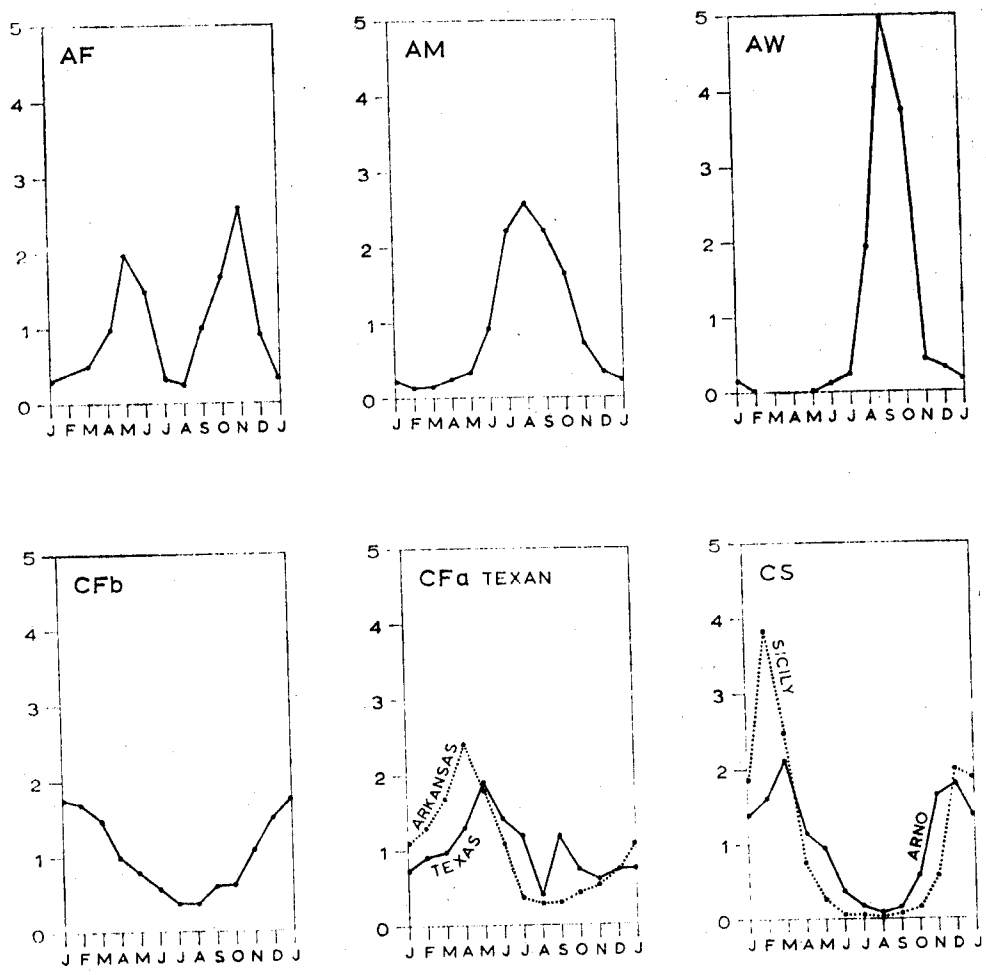
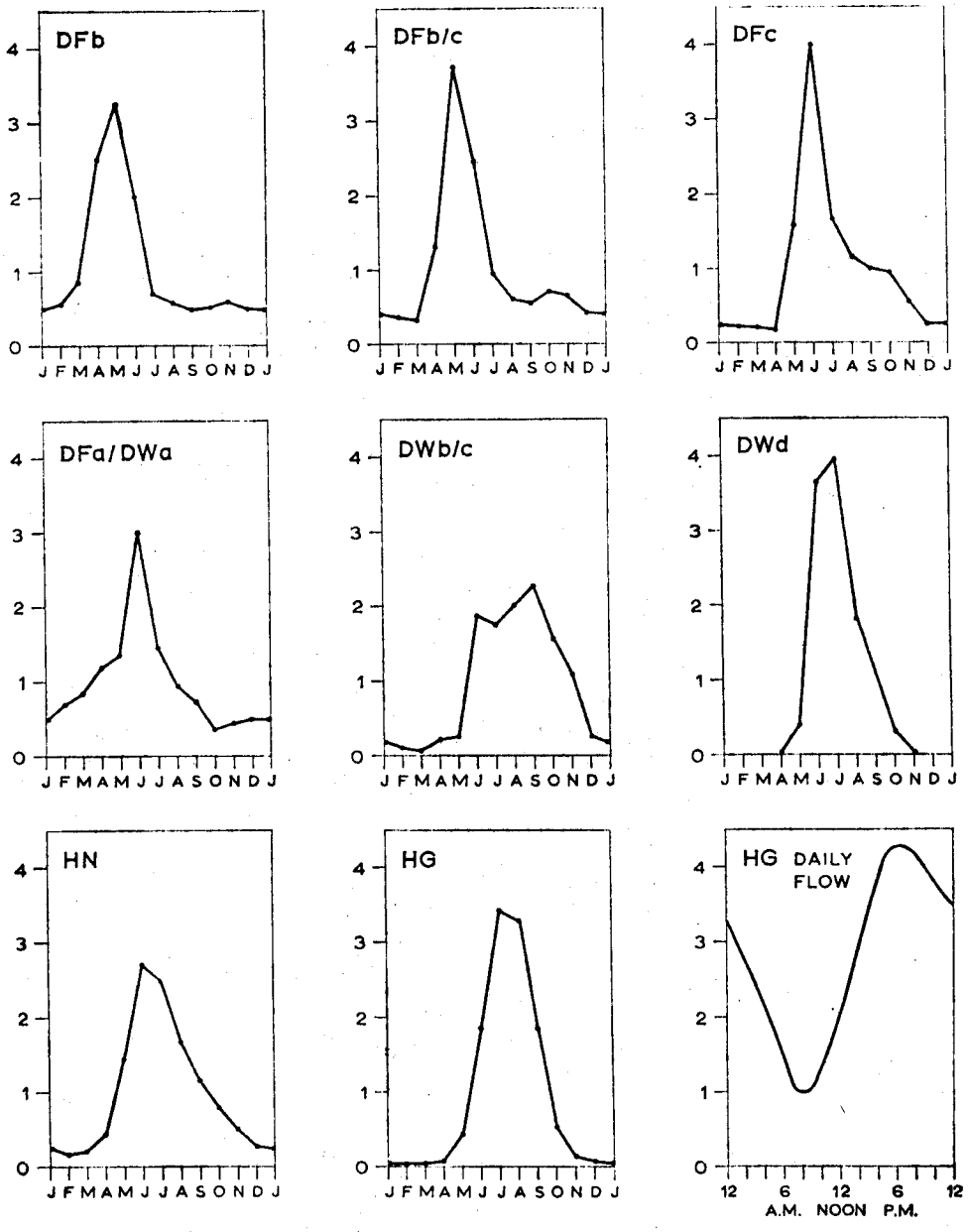


Figure 2·41: World distribution of characteristic river regimes (from Beckinsale, 1969; symbols are explained in the text



AF. Lobaye R., a northern tributary of the Congo; AM, Lower Irrawaddy; AW, Pendjari R., a tributary of the Volta (See Fig. 10.1.2); CFb, Thames R., England; CFa Texan Buffalo R., Arkansas and Guadalupe R., Texas; CS. Arno R. and Imera Meridionale, Sicily.

Figure 2.40 A: Characteristic river regimes controlled mainly by rainfall and warm season evaporation All graphs show the ratio of mean monthly discharge to mean annual discharge (from Beckinsale, 1969)



DFb Dnepr R. at Kremenchug; DFb/c Volga R. at Kuybyshev; DFc Yenisey at Igarka; DFa/DWa Republican R. near Bloomington, Nebraska; DWb/c Amur R. at Komsomol'sk; DWd Indigarka R.; HN Reuss R. at Andermatt; HG Massa R. at Massaboden. HG DAILY FLOW, South Cascade Glacier stream during a fine warm spell (After Meier and Tangborn, 1961).

Figure 2·40B: Characteristic river regimes controlled mainly by cold season snowfall and warm season rainfall. All graphs show the ratio of mean monthly discharge to mean annual discharge (from Beckinsale, 1969)

Nevertheless, we should be aware that, because the flow of water through natural river channels is usually quite variable, it is amenable to statistical manipulation. Discharges for measurement periods of a day, a month, or a year, or some other interval, may be displayed graphically as a simple frequency distribution (see Figure 2.42A). Basic parameters of central tendency include daily, monthly, and mean annual discharge, although the annual value is most commonly used as a single measure of flow scale. As with all statistical analyses, care must be taken to ensure that the sample size is large enough to provide a reliable estimate of the population mean. That is, where flow records are short, the computed mean annual discharge may diverge considerably from the long-term mean. In most cases it would seem that at least 10 years, and preferably more than 20 years, of record are generally needed to ensure a stable mean (see Jeppson et al, 1968).

Single measures of dispersion may take the form of the standard deviation of flows but are more often simply cited as minimum and maximum discharges for the year, or for some other period. In cold climates it may also be appropriate to know when and for how long the river freezes over and the average time of ice break-up. Similarly, in desert areas the number of days with or without channel flow may be of particular geomorphic significance.

A far more common method of representing flow variability, however, is to transform the frequency histogram of discharges into a cumulative frequency distribution called a flow duration curve (see Searcy, 1959; also see Figure 2.42B). It displays the frequency or duration with which discharges of given magnitude are equalled or exceeded. As you can see from Figure 2.42B, however, one of the disadvantages of this type of curve is that it is difficult to discern very much detail at the extremes of the distribution. For this reason the discharge is often plotted on a logarithmic scale, making the distribution of low discharges more obvious (see Figure 2.42C). Furthermore, the distribution of both extremes are more easily interpreted if the time scale is expressed on a normal probability scale (see Figure 2.42D). This last transformation tends to make the flow duration curve conform to a straight line. Just as it is useful in other contexts to characterise a frequency distribution in terms of shape parameters such as skewness and kurtosis, there have also been several attempts to develop similar indices based on the flow duration curve (for example, see Lane and Lei, 1950, and Hall, 1967).

The flow duration curve is, of course, a probability function. We can say for example, that 10, 50, or 90 per cent of the time, discharge in a river will, on average, equal or exceed the flow magnitude given by the flow duration curve. Similarly, from the flow histogram (Figure 2.42A) we can clearly see that very small and very large (low frequency) discharges are both unlikely events, occurring perhaps less than a few per cent of the time on average. This is an important observation because it has led geomorphologists to believe for many years that the discharges most responsible for the shaping of river channels are those intermediate flows with the highest probabilities of occurrence. We will see in later chapters that this type of reasoning can be very misleading.

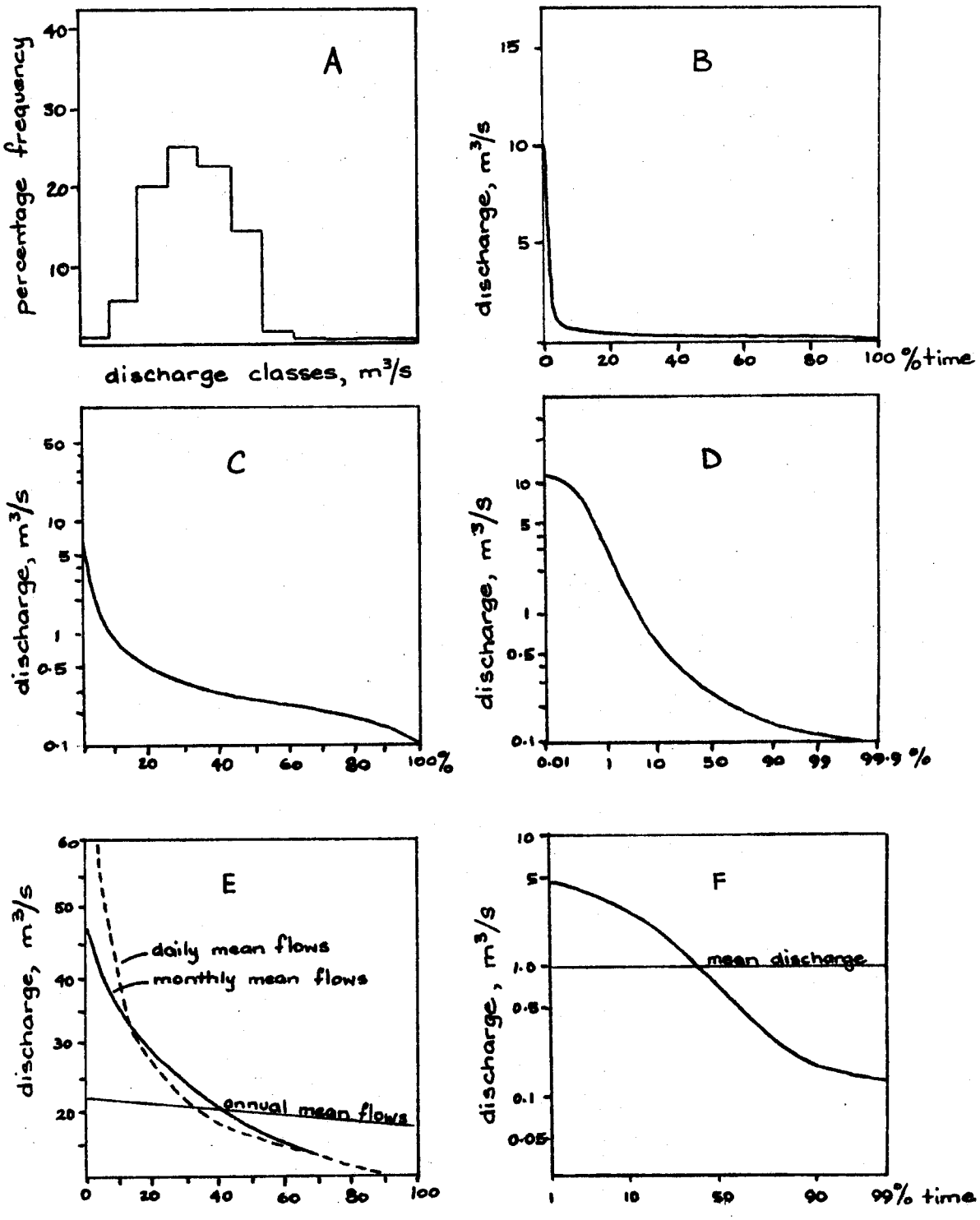


Figure 2.42: Various ways of representing flow duration data (see text) Adapted from Gregory and Walling (1973)

2.10 (c) The statistics of flood discharges

The term "flood" is a useful but very subjective one which is used in a variety of ways by different interest groups. The popular concept of the term, and that adopted by many professional groups, is as much a function of social, as it is of physical, criteria. For example, Dalrymple (1960) has defined a flood as "any relatively high flow that overtops the natural or artificial banks in any reach of a stream". Dalrymple, a hydraulic engineer with the U.S. Geological Survey, would be one of the first to support the view that this type of definition is of limited use in the scientific study and application of hydrologic principles. For example, engineers concerned with the construction of riverside structures, irrigation systems and dams must base their designs on the probability of occurrence of floods (and droughts) of some given magnitude. For this reason the frequency distribution of discharge (and rainfall) extremes has received considerable attention by theoretical statisticians. Much of this theory is of relevance to the geomorphologist, not only because it provides a convenient way of summarising the extreme-discharge record, but because much about a river's behaviour can often be causally related to these extreme events.

The return period or recurrence interval of a flood is the average time interval in which the given discharge will be equalled or exceeded once. Methods of computing the recurrence interval of floods vary (see Jarvis et al, 1936; Chow, 1964) depending on the type of theoretical frequency distribution of maximum values and the measure of frequency (plotting position) adopted. The most widely applied method, however, is that used by the U.S. Geological Survey (Dalrymple, 1960) based largely on the work of Fisher and Tippett (1928), Gumbel (1941; 1945) and Powell (1943). In this method the recurrence interval (T_r) of a flood in a series is defined as:

$$T_r = \frac{n + 1}{m} \qquad (2.76)$$

where n is the number of years of record and m is the simple rank of the flood in the series. Floods are defined as the maximum discharge in each year (the annual series) or as all discharge peaks above some reference discharge (the partial series). By these definitions, a "flood" could conceivably be a relatively small discharge well below the banktops of the channel. The principal objection to the annual series is that it excludes secondary flow peaks in a given year that may exceed the maximum flood in many other years. The partial series is an attempt to overcome this deficiency although the definition of separate flood peaks can be rather subjective; discharge peaks although separated by a recession curve, may be parts of a singl flood wave. For this reason, and because maximum peak-discharge data are generally more readily available, the annual series is more widely used than the partial series.

Table 2.10 and Figure 2.43A illustrate the calculation and graphing of the flood-frequency curve for the Fraser River in British Columbia. The graph is plotted on Gumbel Type I graph paper which has an abscissa scale designed to transform the data to a straight-line plot. In many cases, however, flood frequency plots as a

two-segment distribution consisting of a low slope, high-frequency curve, and a steeper low-frequency curve (see Figure 2.43B). This kink or dog-leg in the curve (see Potter et al, 1968) indicates that the assumptions of the theoretical distribution of extreme values have been violated. The most likely reason for this is that the floods included in the one series belong to two separate populations corresponding with two general types of weather patterns, although Gregory and Walling (1973) suggest that basin properties may also be a contributing factor. The dog-leg is also evident in many Canadian flood-frequency curves where the reason is probably related to floods caused by snowmelt on the one hand, and by late summer rains on the other (M.C. Church, University of British Columbia, personal communication).

When the plotted data do closely conform to a straight line (as they do in Figure 2.43A), the mean annual flood corresponds with that discharge with a recurrence interval of 2.33 years ($q_{2.33}$). The flood which on average is more likely to occur than any other, the most probable annual flood, has a recurrence interval of 1.58 years on the annual series. Both of these parameters are commonly used to express the scale of channel-forming flows in rivers.

Care must be taken not to misinterpret this type of frequency data. For example, in Figure 2.43A a discharge of 1,355 m^3/sec will occur at an average frequency of once every 25 years. There is a temptation to falsely conclude that, if a flood of this magnitude occurs, the same discharge will not occur for another 25 years. In fact, a flood of this magnitude may occur during several years in succession. This type of magnitude-frequency analysis is not a short-term predictive tool; it is merely an estimate of the <u>average</u> recurrence interval of flood magnitude in the long run.

In areas that are homogeneous with respect to flood producing factors, individual streams regardless of their size will have flood frequency curves of about the same slope. Thus a regional flood frequency curve can be constructed if the discharge is expressed in dimensionless terms (see Dalrymple, 1960; Cole, 1966); this may be achieved by dividing all flood values at each gauging station by the corresponding mean annual flood. Station records can be tested for hydrologic homogeneity by computing the mean of the dimensionless 10 year flood ($q_{10}/q_{2.33}$) for all stations and in turn applying this mean value to estimate q_{10} at each station. If these estimates of q_{10} fall within the 95 per cent confidence limits in Figure 2.43C they are assumed to be derived from a hydrologically homogeneous region. Thus it is valid to base a regional flood-frequency curve on the station records. Station values which fall beyond the confidence limits are unlikely to do so by mere chance (a probability of less than 5 per cent) and are assumed to be part of another region requiring a separate regional flood-frequency curve. This test of homogeneity was developed by Langbein (reported in Dalrymple, 1960) and the confidence limits are based on the theoretical probability density of the Type I extreme-value distribution used by Gumbel (1941).

It can be seen in Figure 2.43C that the 95 per cent confidence band for the 10-year flood contracts as the length of record increases.

YEAR	PEAK DISCHARGE m^3/sec	RANKED DISCHARGE	RANK ORDER	RECURRENCE INTERVAL, years
1950	12,601	12,997	1	27.00
1951	8,127	12,601	2	13.50
1952	8,410	11,582	3	9.00
1953	7,334	11,497	4	6.75
1954	9,146	10,902	5	5.40
1955	11,497	10,901	6	4.50
1956	9,854	10,534	7	3.86
1957	10,534	9,854	8	3.38
1958	9,854	9,854	8	3.38
1959	8,552	9,599	10	2.70
1960	9,458	9,458	11	2.45
1961	9,599	9,146	12	2.25
1962	8,297	8,892	13	2.08
1963	7,759	8,693	14	1.93
1964	11,582	8,634	15	1.80
1965	8,637	8,552	16	1.69
1966	7,957	8,552	17	1.69
1967	10,901	8,410	18	1.50
1968	8,891	8,297	19	1.42
1969	7,872	8,127	20	1.35
1970	8,693	8,014	21	1.29
1971	8,552	7,957	22	1.23
1972	12,997	7,872	23	1.17
1973	8,014	7,759	24	1.13
1974	10,902	7,702	25	1.08
1975	7,702	7,334	26	1.04

Table 2.10 : Annual floods and recurrence intervals for the Fraser River at Hope, British Columbia, 1950 - 1975.

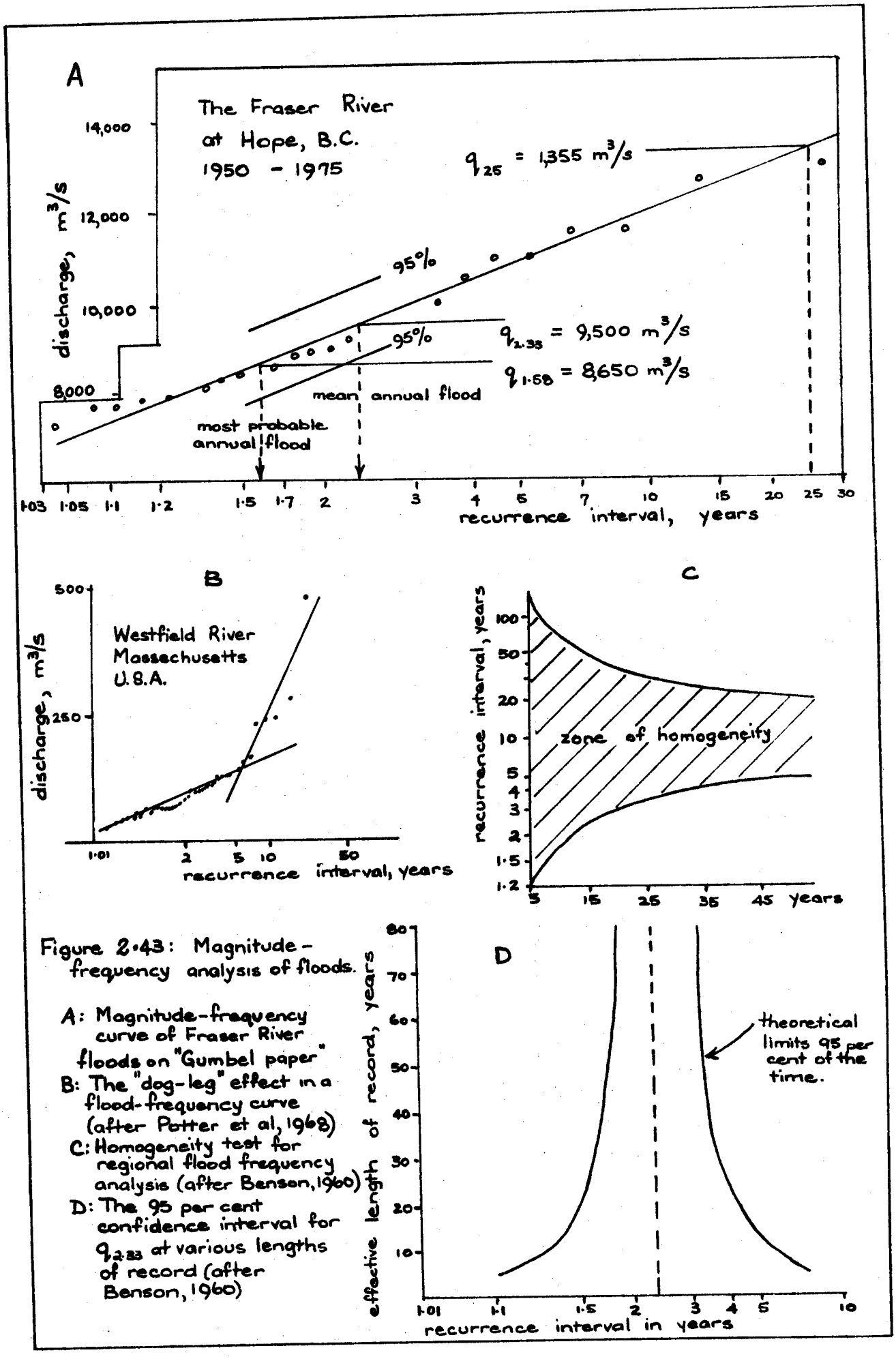


Figure 2·43: Magnitude-frequency analysis of floods.

A: Magnitude-frequency curve of Fraser River floods on "Gumbel paper"

B: The "dog-leg" effect in a flood-frequency curve (after Potter et al, 1968)

C: Homogeneity test for regional flood frequency analysis (after Benson, 1960)

D: The 95 per cent confidence interval for $q_{2.33}$ at various lengths of record (after Benson, 1960)

The reason for this is that, as the length of record increases, the influence of short-term fluctuations in flood magnitudes resulting from unusual climatic conditions decreases. In other words, the longer the record the more confident we can be that the computed magnitude of the 10-year flood approaches that of the theoretical or "true" long-term 10-year flood.

Fluvial geomorphologists have been concerned in recent years with more frequently occurring floods as a means of characterising flow scales in rivers (for example, see Leopold, Wolman and Miller, 1964). In this context Figure 2.43D is of particular relevance; it shows the 75 per cent confidence limits of the mean annual flood plotted against the length of discharge records. We can see that, for a five year record, the actual mean annual flood may have an apparent recurrence interval ranging anywhere between 1.15 to 7.5 years, 95 per cent of the time. The corresponding range for a 25 year record is from 1.55 to 3.80 years. The 95 per cent confidence band for the mean annual flood for the Fraser River is shown in Figure 2.43A. These limits indicate that 95 per cent of the time, the actual magnitude of the mean annual flood might, because of chance variation, occur anywhere between 8,700 and 10,300 m^3/sec, or $\pm$ 8.4 per cent about the computed mean.

Benson (1960) calculated from a 1000-year synthetic discharge record that it requires at least 12 years of record to yield a mean annual flood value within 25 per cent of the correct value 95 per cent of the time. In order to reduce the error to 10 per cent, the length of record must be increased to 40 years for the mean annual flood, and to 90 years for the 10-year flood.

Although the above examples can only be a guide to the accuracy of a particular flood-frequency curve, they clearly indicate that it is dangerous to treat too precisely the flood magnitudes computed for short lengths of record ! Furthermore, unless the length of record exceeds 50 years, it would seem pointless to base an argument on small differences in computed flood discharges such as that between $q_{1.58}$ and $q_{2.33}$; in most cases the two discharges will not be significantly different from each other in a statistical sense.

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