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# A SUMMARY OF METHODS FOR THE COLLECTION AND ANALYSIS OF BASIC HYDROLOGIC DATA FOR ARID REGIONS

By

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# A SUMMARY OF METHODS FOR THE COLLECTION AND ANALYSIS OF BASIC HYDROLOGIC DATA FOR ARID REGIONS

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## ABSTRACT

This report summarizes and discusses current methods of collecting and analyzing the data required for a study of the basic hydrology of arid regions. The fundamental principles behind these methods are no different than those that apply to studies of humid regions, but in arid regions the infrequent occurrence of precipitation, the great variability of the many hydrologic elements, and the inaccessibility of most basins usually make it economically infeasible to use conventional levels of instrumentation. Because of these economic considerations hydrologic studies in arid regions have been commonly of the reconnaissance type; the more costly detailed studies are generally restricted to experimental basins and to those basins that now have major economic significance. A thorough search of the literature and personal communication with workers in the field of arid-land hydrology provided the basis for this summary of methods used in both reconnaissance and detailed hydrologic studies. The conclusions reached from a consideration of previously reported methods are interspersed in this report where appropriate.

# INTRODUCTION

The use of conventional methods of studying the basic hydrology of humid areas is generally not practical for areas in an arid environment. This is particularly true if hydrologic knowledge is sought for an entire region rather than for a few individual basins in the region. The same basic principles govern the study in both arid and humid environments, but in the arid regions the great variability of the many hydrologic elements and the inaccessibility of most basins usually make it economically infeasible to study the regional hydrology through the use of conventional instrumentation that is as complete for the purpose as that used in studies of humid regions. Furthermore the infrequent occurrence of storms makes it infeasible to study the regional hydrology by reasonably complete conventional instrumentation of a few small areas for a few years and successive transfers of the instruments to other small areas. The number of years required to complete the regional study by this successive sampling method would be prohibitively long.

Over the years hydrologists have developed and improvised investigational techniques for use in arid-land studies, but up to now there has been no single reference source to acquaint one with those methodologies. This paper represents an attempt to fill the need for such a reference.

## Purpose and Scope

The purpose of this report is to summarize and discuss current methods of collecting and analyzing the data required for a study of the basic hydrology of arid regions. As a state-of-the-art study, it entailed a thorough search of the literature and personal communication with workers in the field of arid-land hydrology. Although the principles discussed in this report have general applicability, most of the discussion deals with the Basin and Range physiographic province (fig. 1), as delineated by Fenneman (1931).

At this point it is necessary to define the term "arid" and to delineate the facets of the hydrology with which this report is concerned. A precise definition of arid will not be attempted. For the purpose of this report, an area is considered arid if most of the area receives considerably less precipitation than the potential evapotranspiration. The indefinite adjective "considerably less" is used in the definition because there is nothing definite about the determination of potential evapotranspiration; the various methods of computing its value may give results that differ significantly from each other. The definition of arid adopted for this study does not exclude areas, which though generally dry, are drained by streams that originate in humid or subhumid highlands.

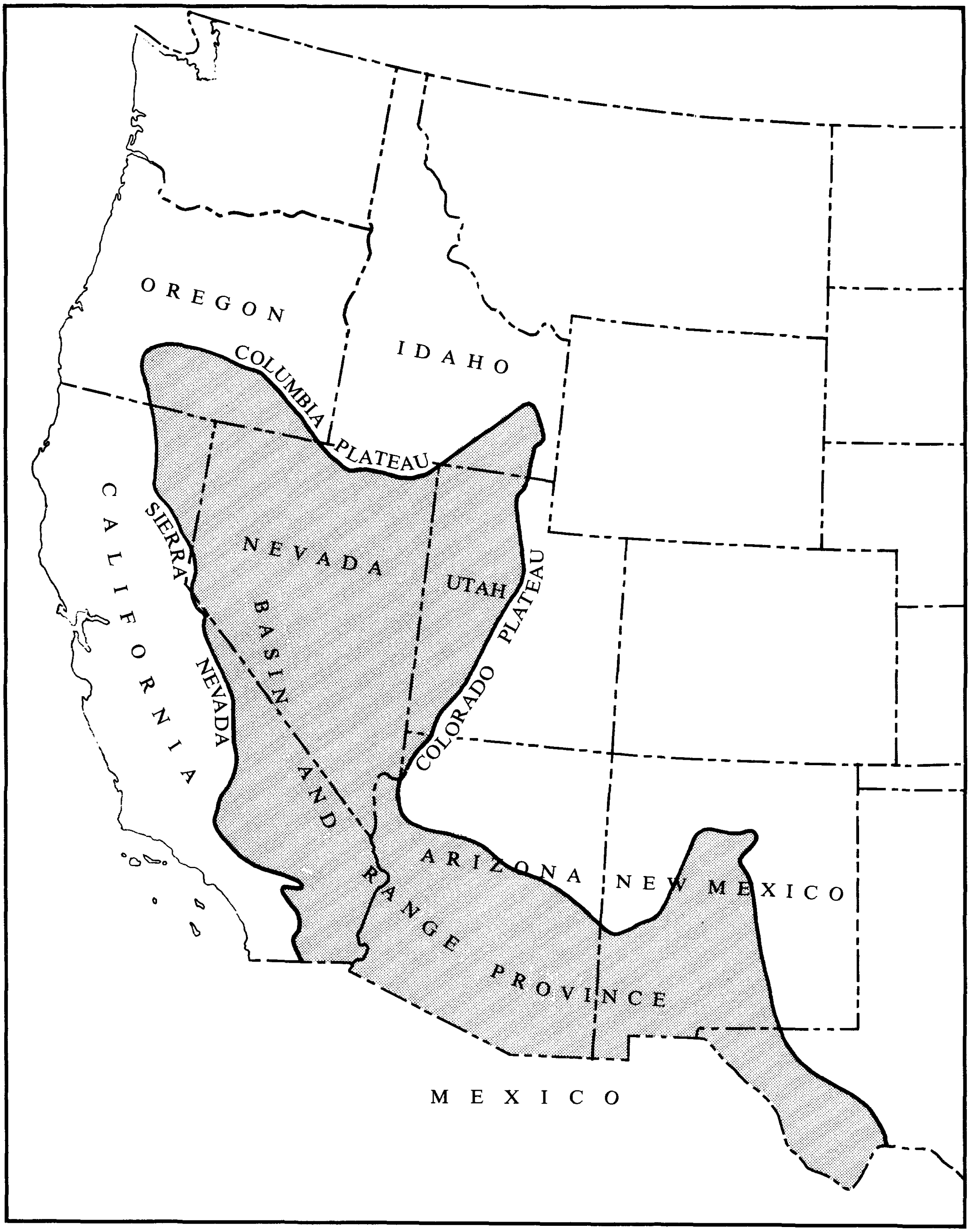


FIGURE 1.--Location of Basin and Range province in the United States. [After Fenneman, 1931.]

The facets of the hydrology with which this report is concerned, include the following:

1. A description of the hydrologic regime of the region. This involves the temporal and areal distribution of such hydrologic elements as: precipitation, soil moisture, evapotranspiration, evaporation from open-water surfaces, streamflow (including floods), and ground water; and such associated characteristics as water temperature, dissolved solids, and sediment transport.

2. A general inventory of the water resources of a region, including a hydrologic budget. This refers to a seasonal, annual, or long-term accounting of the surface and subsurface discharge, water withdrawal and consumptive use, import and export of water, and surface and subsurface storage. This information, together with an accounting of the precipitation, provide the data for a hydrologic budget.

3. Sufficient information for general study of both interrelations among various hydrologic elements and relations of various hydrologic elements to the environment. Basic data are needed to identify such relations as: (a) Short-term or annual relation of runoff to precipitation, (b) quantitative relations among dissolved constituents in surface or ground waters, (c) relation of base-flow or underflow to ground-water levels, (d) relation of streamflow characteristics to physiographic and climatic parameters, (e) relation of sediment production to geologic and climatic parameters, and (f) relation of chemical quality to geologic parameters.

This report is not concerned with the collection of research-type data that the authors consider to be too highly specialized for inclusion in a basic-data program. For example, the report discusses a simple program for monitoring changes in channel morphology in the vicinity of stream-gage sites, but does not consider the collection of specialized data for use in complex studies of fluvial morphology. Vegetation is discussed solely from a standpoint of water loss by evapotranspiration; no consideration is given to the collection of tree-ring data for detecting climatic trends, or the study of trees and their local environment for the purpose of deducing facts concerning flood activity or land aggradation and degradation. Nor is consideration given to the collection of data for the study of the effect of such changes in land use as reduction in grazing, removal of vegetation, and urbanization. One should not expect to find in this report a discussion of the measurement of natural tritium to estimate the age, movement, and quantity of water stored in ground-water basins. These are but a few of the more obvious examples, but they illustrate the types of data whose collection is beyond the scope of a basic-data program.

The various elements of the hydrologic budget are discussed separately in the body of this report. However, because these elements are interrelated it is often difficult to isolate them completely for separate discussion of techniques. Hopefully, any confusion that may arise from the artificial separation of hydrologic elements will be dispelled in a unifying discussion of reconnaissance-type and detailed basin studies at the end of the report. In the discussion of basin studies no effort is made to define the minimum level of data collection that is consistent with: (1) The particular environment of a study area, (2) the purpose of investigating the area, and (3) the immediacy of need for results of the investigation. The study of such guidelines, admittedly urgent, is beyond the scope of this report.

The discussion of methods is aimed at the reader who has had some experience with hydrologic investigations and the report, therefore, is not intended to be a primer. Standard techniques that are well known are mentioned but are not described--for example, the use of simple stage-discharge relations to determine surface runoff, or the construction of frequency or duration curves. Recent developments that are applicable to studies of arid-land hydrology, but which have not had wide dissemination in the literature, are discussed briefly and bibiographic references are furnished. Modifications of existing techniques that have not been previously discussed in the literature are described in more detail.

## Acknowledgments

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# DESCRIPTION OF THE BASIN AND RANGE PROVINCE

## Physical Setting

Because so many of the methodologies to be discussed refer to areas in the Basin and Range province, a general description of the region is a helpful adjunct to the report. The literature abounds with such descriptions (for example, Fenneman, 1931, or Zierer, 1956) and the account that follows is extracted from those descriptions.

The Basin and Range province includes that part of Southwestern United States that is bounded on the west by the Sierra Nevada, on the east by the Colorado Plateau, and on the north by the Columbia Plateau (fig. 1). The landforms of this vast arid region are surprisingly uniform--nearly parallel mountain ranges separated from each other by broad, smooth-floored, elongate alluvial basins. A few of these basins are tributary to rivers that eventually flow into the ocean, but most of the basins are closed; that is, their drainage is internal and does not reach the sea. Faulting has created most of the ranges, uplifting them as blocks, high above the adjacent depressed troughs. All the component rocks--igneous, sedimentary, and metamorphic alike--have been cut by those displacements with apparent disregard for preexisting geologic structure.

A typical Basin and Range mountain is therefore steep sided. Its soil mantle is thin and numerous steep canyons are incised in its flanks. Between the mountains and the valley floors are extensive areas of intermediate slope. These sloping areas may be erosion surfaces, or pediments cut on rocks and thinly mantled by unconsolidated rock debris; more commonly they are formed by coalescing alluvial fans constructed from water-transported rock debris. The upper ends of the fans are composed of coarser material than the lower ends. The fans range in width from several feet to several miles, and range in thickness from a few feet to many hundreds of feet. Rarely is there a single channel on the alluvial fan, and streamflow debouching from the canyon mouth flows down the fan in several distributary channels. The distributaries are generally incised at the upper end of the fan and gradually disappear as they approach the lower end of the fan.

There is generally too little ground-water storage in the soil and fractured mantle rock of the mountains to sustain streamflow for any great length of time after the rains cease, and consequently most of the streams are ephemeral.

The canyon streambeds usually are nearly impermeable, and as the streams flow down the canyons, runoff from tributary channels increases the discharge. However, after the streams leave the canyons and flow over the alluvial fans, they undergo a loss in discharge. The rock debris and gravel composing the fans are usually highly permeable, and even those streams that flow perennially in the canyons become ephemeral after they reach the fans. Davis and De Wiest (1966, p. 396), in discussing the permeability of alluvial fans, state:

> "Apex areas of alluvial fans are composed of older material or are dominated by mudflows and other poorly sorted material from flash floods so that the average permeability is considerably lower than deposits farther down on the alluvial fan. The deposits in an intermediate position tend to be worked by streamflow and have higher permeabilities despite smaller grain sizes. The distal (lower) parts of the alluvial fans tend to interfinger with playa deposits that are predominantly fine-grained material."

The streamflow resulting from minor storms will usually disappear entirely into the alluvial fan before reaching the flat desert floor. After intense storms, however, the runoff may be great enough to reach the desert floor, despite losses into the alluvial fan. The desert-floor sediments are often fairly permeable, but in some areas soil layers of low permeability that are close to the surface will cause ponding of the runoff, and form a playa lake. Such lakes, of course, are ephemeral; the water evaporates in a short time; and the playa remains bare until the next flood event.

The streamflow that seeps into the alluvial fan and some that seeps into the desert floor, recharges the underlying ground-water body and is usually the only major source of local water supply in desert areas. The precipitation that falls on the alluvial fan or desert floor is almost entirely soaked up in the upper layers of the ground and is lost later by evaporation or evapotranspiration; only a minor amount penetrates to the ground-water body below. Davis and De Wiest (1966, p. 424) illustrate this fact concerning precipitation on the desert floor with the following example:

> "For example, a soil that has a specific retention of 15 percent and is depleted of moisture to a depth of 2 feet during the summer heat will require 3.6 inches of rain merely to make up for the soil-moisture deficiency. If the rain occurs at several different times during the year, intervening periods of dry weather will cause the loss of water from the soil so that amounts much in excess of 3.6 inches will be needed to start recharge."

## Climate

The climate of the region is marked by summers that are hot and winters that range from warm to cold, depending on the altitude and latitude. Aridity, of course, is the most common characteristic of the climate. The low humidity and absence of clouds in the desert cause a large daily fluctuation in air temperature. Commonly, about 90 percent of the available solar radiation is absorbed by the ground during the day, and the ground in turn rapidly heats the lower layers of the atmosphere. At night, about 90 percent of the heat radiated by the ground is lost to outer space and rapid cooling occurs. By comparison, in a humid area only about 40 percent of the solar heat is absorbed by the ground during the day, and only about 50 percent of the heat radiated by the ground at night is lost to outer space.

Precipitation has two seasonal sources. Winter storms--October through April in the northern part of the region and November through April in the southern part--originate over the Pacific Ocean, as a result of the interaction between polar Pacific and tropical Pacific airmasses. These storms move eastward over the land where they cover widespread areas and usually last for several days.

Summer storms--May through September or October, depending on the latitude--are primarily thunderstorms and convective showers which form in moist tropical air that normally enters the region from the Gulf of Mexico. However, it is not too unusual for late summer storms to be associated with moisture drawn over the land from disturbances centered in the tropical Pacific Ocean. Sellers (1960) explains that the meteorological conditions associated with convective midsummer storms, are associated, in turn, with conditions that are unfavorable for late summer storms of tropical Pacific origin. The converse is also true, and therefore it is uncommon in a single year to have both heavy midsummer convective rains and heavy late-summer tropical storms. On the other hand, heavy late-summer tropical storms and heavy winter storms are generally associated with each other.

The intensity of summer thunderstorms is usually much greater than that of winter storms, but summer storms do not last as long as winter storms, nor are they as widespread in area. As a matter of fact, even during periods of widespread thunderstorm activity, the rainfall is usually very "spotty"; the many small, scattered areas that receive heavy rainfall are generally smaller than 10 square miles each and usually range from 1 to 5 square miles. In the more arid areas a single summer storm may bring all the rainfall occurring within that month, and occasionally within the entire year. The closer the desert area is to the Pacific Coast, the greater the likelihood that winter storms will predominate in supplying precipitation to the area; farther inland from the Pacific Coast, summer storms are predominant. This is particularly true of the southern part of the region.

## Hydrologic Classification of Valleys

Snyder (1962) proposed a classification of valleys in the Great Basin section of the Basin and Range province; that is, that part of the province whose drainage does not reach the sea. Snyder's classification is based on both topographic and hydrologic considerations. Those valleys that are completely surrounded by land higher than the valley floor, as commonly found in Nevada, are considered topographically closed. Those valleys that have graded connections with adjacent valleys are considered topographically open. Open valleys are commonly found in southern Arizona, where isolated mountains are completely surrounded by valleys. The topographic characteristics of the valleys, however, are not necessarily indicative of the hydrologic characteristics. In the low-lying part, the valley fill may be saturated with water to the ground surface or it may be entirely drained, depending upon the position of the controlling outlet. The playa on the valley floor may be dry or wet depending on the depth to water in the valley. Table 1 presents Snyder's classification and figure 2 illustrates some of the valley types. It should be realized that Snyder's classification of valleys is not all-inclusive; the hydrogeology of many desert valleys is much more complex than that described here.

TABLE 1.--*Hydrologic classification of valleys in the Great Basin*

[After Snyder, 1962]

| Topographically closed | Topographically open |
|---|---|
| **Drained** | |
| Alluvium, unsaturated near land surface, but water table may occur at depth. | Alluvium, unsaturated near land surface. |
| Playa, dry during dry season. | Playa; if present, usually dry during dry season. |
| Water escapes through subsurface outlet. | Water moves out of valley near base of valley fill (fig. 2C) or above a barrier (fig. 2D). |
| Water usually of good quality (fig. 2B). | Water usually of good quality (fig. 2C); quality may vary if there is a barrier (fig. 2D). |
| **Undrained** | |
| Alluvium, saturated.<br>Playa, wet during dry seasons.<br>Water lost by evapotranspiration.<br>Quality of water will range from good in areas farthest from playa to poor in the valley center close to the playa. | Not known in Great Basin. |

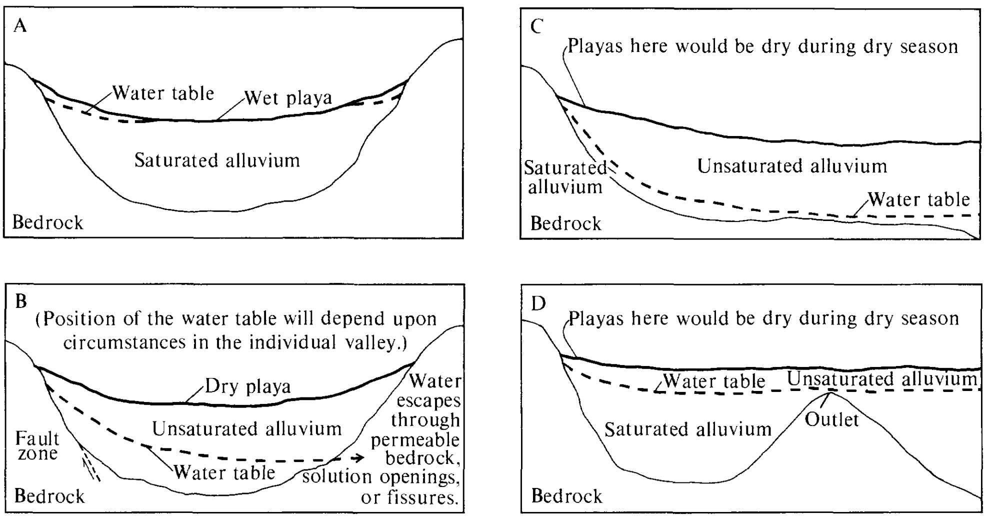



FIGURE 2.--Valley types found in the Great Basin.
(After Snyder, 1962.)

## CURRENT METHODS OF STUDYING THE ELEMENTS OF THE HYDROLOGIC BUDGET

This section of the report discusses current methods of collecting data and analyzing the various elements of the hydrologic budget. It is emphasized that the collection of recorded or observed data at sampling sites is not the end product of a basic hydrologic-data program. As mentioned in the introduction to this report, the aims of such a program include preparation of a hydrologic budget for the basin or subregion studied, estimates of the time and areal distribution of the various hydrologic elements, and the accumulation of sufficient information to permit general study of both interrelations among the various hydrologic elements and relations of the various hydrologic elements to the environment.

The instrumentation and methodologies used to accomplish the aims of a basic hydrologic-data program for a humid area, although continually being improved, are relatively well established. Because of the relative homogeneity of the hydrologic elements in a humid area, point observations of an element usually provide reliable indexes of the basinwide values of that element. This homogeneity is lacking in arid areas, and therefore study of the hydrology is more complicated. Precipitation, for example, not only occurs infrequently but has great variability. Furthermore, the sparse development of arid regions makes most of the area poorly accessible by surface travel, and hydrologic observation sites are therefore few in number and are generally limited to the more accessible and populous areas.

In view of the many difficulties, it generally is not economically feasible to make detailed hydrologic studies of arid areas, except for those areas that now have economic significance or those areas that are established as experimental basins for research in arid-land hydrology. Most hydrologic studies of arid areas therefore have been of the reconnaissance type for the purpose of obtaining long-term average values of the elements in the hydrologic budget, so that the potential water supply can be evaluated with regard to both quantity and quality.

### Precipitation

Precipitation is the most basic element in the hydrologic budget, and the first step in most hydrologic studies, whether detailed or of the reconnaissance type, is to relate mean annual precipitation to altitude. This relation is then used to construct an isohyetal map of mean annual precipitation (Dawdy and Langbein, 1960; Peck and Brown, 1962) or is used with a hypsometric (altitude versus drainage area) curve of the basin to determine mean annual basinwide precipitation (Crippen, 1965).

Where separate isohyetal maps are constructed for the period of summer storms (May through September or October, depending on the latitude) and the period of winter storms (October or November through April), the winter-storm data for high-altitude precipitation stations can sometimes be augmented by data from snow courses that are sufficiently high to have little winter snowmelt. The water equivalent of the snowpack at these courses is usually measured April 1 of each year. The use of snow-course data requires sufficient high-altitude precipitation stations to establish a relation between the water equivalent of the snowpack on April 1 and the winter precipitation. In preparing seasonal isohyetal maps, the winter period is usually considered to be October through April and the summer period May through September, to provide consistency with the widely used inclusive dates of the water year, October 1 to September 30. This modification of the seasonal dates has little effect on the results of a precipitation study.

The relation of precipitation to altitude may vary widely over large areas and consequently it is necessary to determine individual precipitation-altitude relations for individual geographic zones. Peck and Brown (1962), for example, subdivided the State of Utah into 20 geographic zones. Even within each zone the data for individual precipitation stations will depart from the average curve of relation for the zone. These departures or anomalies are usually related to physiographic features. A preliminary step to constructing an isohyetal map is to plot these anomalies on a topographic map and draw lines of equal anomaly. The precipitation for any point in the study area is then determined by use of both the precipitation-altitude relations and the anomaly contours.

A more sophisticated method of computing point precipitation is to correlate the available records of precipitation with additional details of the terrain in the vicinity of the precipitation gage. Russler and Spreen (1947) used such physiographic features as exposure, orientation, rise, zone of environment, and altitude in a graphical coaxial correlation with mean annual precipitation. Application of this method is extremely laborious and because of the paucity of precipitation stations and the great variability of precipitation in arid regions, it is questionable whether the refinement added by this method justifies the additional work required. However, regardless of whether the Russler-Spreen type of precipitation relation is used or whether the altitude-anomaly method is used, it seems desirable that separate relations be used for winter-storm periods and summer-storm periods. The storm paths for the two periods usually differ, winter storms coming generally from the west and summer storms from the south, and consequently differing relations between precipitation and such physiographic features as orientation and exposure would be expected. Isohyetal maps for the two periods can then be combined into an isohyetal map for the entire year. Hely and Peck (1964) in their study of precipitation in the Salton Sea area of the Lower Colorado River Basin used separate winter and summer periods in applying the altitude-anomaly method to each of eight geographic zones in the area.

The areal distribution of native vegetation types may be an indicator of the areal distribution of mean annual precipitation. Hardman, in 1949, used such information as an adjunct to the sparse precipitation data that were available to him, to prepare an isohyetal map of Nevada that was later updated in 1965. Hardman (Hardman and Mason, 1949, p. 54-55) states:

> "In preparing the precipitation map of Nevada the data from over one hundred Weather Bureau stations were correlated with data on elevation, general topography, latitude, and types of vegetation. These vegetative types cover broad areas in Nevada. The extent of their areal development seems to have been governed in large measure by the amount of moisture normally received, and some types seem to be fair indicators of the average annual depth of precipitation . . . . The vegetative types were most useful in refining and extending precipitation zone boundary lines where historic weather records were incomplete or lacking. The indicator types of native vegetation used in this study and their apparent moisture relationships are discussed below.
>
> "The southern desert shrub type seems to have its upper limit at approximately 5 inches of average annual precipitation. The northern desert shrub type joins the southern desert shrub type on the northern boundary of the latter type and is dominant over most of the area which receives more than 5 and less than approximately 8 inches of rainfall.
>
> "Northward and on higher elevations the sagebrush vegetative type replaces the northern desert shrub. It is fairly well identified with areas which receive more than 8 inches and less than 12 inches of moisture annually. Stands of Piñon-Juniper occupy some areas in this rainfall belt and often extend upward into areas which receive more than 12 inches of moisture. When used in conjunction with other criteria, this type is a usable, but not an ideal indicator of moisture supplies. Higher on the mountains, above the sagebrush, is a vegetative type known as mountain brush. It seems to be best developed where the precipitation averages approximately 12 to 15 inches.
>
> "Grass and grassland herbs are a major component of the vegetation in the 15- to 20-inch precipitation zone. In Nevada, when the precipitation averages more than 20 inches, which is near the top of the higher mountain ranges, the vegetation becomes alpine in character."

There is virtually unanimous agreement that precipitation increases locally with altitude during general winter storms, although Williams and Peck (1962) show that the rate of increase of precipitation with altitude varies with the synoptic situation. No such agreement exists, however, with regard to intense summer storms. For example, Osborn and Reynolds (1963) state that orographic effects are absent during intense summer storms at the Alamogordo Creek Experimental Watershed near Santa Rosa, N. Mex. On the other hand, Peck (oral commun., 1966) states that summer storms at the higher altitudes usually are of equal or lesser intensity than those at the lower altitudes, but because the duration of the high-altitude storms is greater, total volume of summer precipitation increases with altitude. The orographic effect, however, is not nearly as pronounced in summer as it is in winter.

Croft and Marston (1950) are not in agreement with Peck with regard to the intensity of summer thunderstorms. In their discussion of summer rainfall characteristics in northern Utah, they show that the intensity of rainfall bursts at mountain stations is greater than that for valley stations. Brancato (1942), however, in discussing thunderstorm rains in the Sevier River basin of Utah, states,

> ". . . three to four times as many thunderstorms occurred on the middle and upper windward slopes of the mountains as on the relatively flat and lower portions of the basin. However, contrary to published and popular accounts, the thunderstorms produced the greatest amount of precipitation at the lower elevations and not on the mountain slopes. The most favorable condition for the production of heavy rain is the presence of an airmass with a sufficient amount of available energy and the greatest possible amount of moisture. Orographic lifting is very effective as a mechanism to release the latent energy in an airmass but as the air is lifted over progressively higher terrain the total amount of available precipitable water above any given area becomes progressively smaller. Over the lower level areas therefore there is always more potential rain waiting to be released, but the opportunity for realization occurs less frequently than at the higher elevations which are favored by orographic lifting . . . .
>
> "With regard to variation of thunderstorm rainfall with elevation, the records cited as well as records over other portions of the country show that although sloping terrain will increase the frequency of thunderstorms, the amount of precipitable water in the atmosphere above a unit area decreases with elevation, so that over a long period of time a station located at a lower elevation is likely to experience the most intense thunderstorm."

Dorroh (1946) states that "data collected thus far concerning precipitation and runoff in the Southwest (Utah, Colorado, New Mexico, and Arizona) materially substantiates this (Brancato's) hypothesis, although it must be realized that exceptions can be found." Dorroh (1950) referring to these exceptions as storms of the "lift-convective" type further states ". . . it is more than probable that the location of this particular area [the northern Utah area discussed above by Croft and Marston] with respect to incoming moist air--and its abruptly rising topography--results in storms approaching those we have referred to as the 'lift-convective' type. These storms are becoming notorious as a source of high intensity rainfall."

From the foregoing quotes it would seem that intimate knowledge of the precipitation pattern in a specific area is needed in deciding if orographic influence is a factor to be considered when constructing isohyets of summer precipitation. In some areas, particularly in those that receive almost all their precipitation in the summer, the pattern of vegetation may offer a clue as to whether or not orographic influences predominate. One occasionally hears conjecture from highway maintenance men to the effect that one basin is more susceptible to summer convective storms than an adjacent basin of similar topography. In support of this conjecture the theory has been advanced that this effect is a result of differences in the albedo and thermal diffusivity of the surficial material of the two basins, or of such differences in areas upwind from the basins. These differences would cause the ground-surface temperatures of one area to be consistently higher than those of another area. The stronger thermal updrafts over the warmer of the two areas would contribute to greater instability of the overlying air, thereby increasing the probability for convectional precipitation to occur. Meyer (1966) obtained apparent corroboration of that theory in a test made in the Harquahala Plains area of Arizona. Meyer, however, used rainfall data for only a single year and his results, therefore, are not conclusive. His report ends with the following statement: "Finally, the fact is that the relationship between rock type and [areal] rainfall distribution, which is apparently evident from the field data, might simply be the result of pure chance, in that the [areal] rainfall distribution was not observed for more than one rainy season and the [areal] rainfall distribution for another rainy season may not be similar."

Information concerning mean annual precipitation is needed for evaluating the potential water supply of an area, but even more urgent, perhaps, is the need for regional precipitation-frequency data for use in the design of highway, railway, and local drainage facilities. Not only do we need depth-duration-frequency relations at a point, but also depth-area-duration relations. On a national scale the U.S. Weather Bureau (USWB) (1961, 1964) has attempted to fill this need with the publication of Technical Papers 40 and 49, which present these relations in sets of graphs and tables for the 48 conterminous states. The data in these two reports are highly generalized for the Basin and Range province because of the sparse coverage of the area by long-term precipitation stations, the abrupt and frequent variation in topography, and the "spottiness" of intense summer precipitation.

Osborn (1964), on the basis of his experience with experimental basins in Arizona and New Mexico, finds that the depth-area-duration relations in the USWB technical papers can easily be misused as criteria for drainage design in the Southwest. Peak discharge rates that result from summer storms of high intensity and short duration are much greater, on the smaller watersheds at least, than those that result from winter storms of comparatively long duration. The curves given in the technical papers do not differentiate between summer and winter storms and they show that depth increases with duration. In other words, the graphs for short-duration precipitation in the technical papers are associated with summer storms, and those for durations of more than about 6 hours are generally associated with winter storms. It is therefore likely that the peak discharge on a small watershed resulting from a 25-year 1-hour storm will be greater than that resulting from a 100-year 24-hour storm, because the latter storm is probably a winter storm of relatively low intensity.

The combining of winter storms and summer storms in the technical papers also results in depth-area relations that may not be strictly applicable for design purposes, because the flood-producing summer storms cover areas that are relatively small in comparison with the areas covered by general winter storms. Fletcher (1960), in discussing the characteristics of precipitation in the rangelands of the Southwest, reiterates that most runoff results from intense short-duration bursts of convective rainfall and states that 60 percent of the individual storm cells have diameters smaller than 1½ miles and that only 3 percent of all storms cause runoff. Hodge (1963, p. 124) corroborates this, stating, "Some recent observations of thunderstorm rainfall from several dense rain-gage networks set up by the U.S. Agricultural Research Service in Arizona, Idaho, and New Mexico indicate that as many as 12 storm centers may occur over an area of 100 square miles, but 80 percent of these storms may cover less than 4.5 square miles. Gage density varies from one per 4 square miles to four per square mile." Osborn (1964) concludes that, "For runoff designs involving large watersheds, two probability estimates may be needed--the probability of storms of certain intensities and sizes falling on tributary watersheds of finite sizes and the probability of storms developing over a multitributary system in such patterns as to produce important volumes and peaks of runoff." The prospect of fulfilling this need in the near future is not bright.

A method sometimes used in frequency studies of summer storms is the station-year method (Clarke-Hafstad, 1942). This is a means of extending the precipitation record by assuming that the records for a number of stations in a region may be considered as a composite record for a period equal to the number of years involved. For example, if 15 years of record are available at each of 20 stations, they would be combined to give a single record equivalent to one of 300 years at any site in the region. It is not necessary that all the station records cover a concurrent period or be of equal length, but it would seem advisable that each station have a record sufficiently long to define the mean for the rainfall duration (2 hours, 6 hours, etc.) under consideration. The rainfall at each station can then be expressed as a percentage of its mean, thereby putting all stations on a common basis for combining records. The following two criteria must be met if the station-year method is to be used:

1. The stations used must be mutually independent--that is, the stations must be far enough apart so that no single storm covers more than one station.
2. The stations used must all be in a meteorologically homogeneous region--that is, the rainfall-frequency curves for all points in the region must be nearly identical, when the rainfalls for each site are expressed as percentages of their respective means.

The discussion in the preceding paragraphs indicated that it is difficult to delineate reliably areas of meteorologic homogeneity in the Basin and Range province. Then too, the assumption of total independence of station records is often suspect. For those reasons the station-year method is not widely used. Furthermore, random variation in storm activity in arid regions usually results in there being alternate periods of relatively wet and dry years. The consistent use of short-term records for one or the other of these periods may bias a station-year distribution. The Geological Survey, however, is engaged at present in developing a procedure to define the degree to which records may be extended in time by means of a modified station-year method (Matalas, oral commun., 1969). In that procedure the assumption of total independence of station records is being replaced by a computed degree of independence based on statistical principles. The method, when developed, should be a valuable tool in frequency studies for areas of probable meteorological homogeneity that have precipitation records long enough to have sampled both wet and dry periods.

It was previously mentioned, or at least implied, that the only adequate networks of precipitation gages in arid regions are located in experimental basins and in some basins of economic importance. Elsewhere, the networks are sparse and gages are usually located at easily accessible sites. This generally means that precipitation data are lacking at the high altitudes where the bulk of the runoff originates. To supplement the sparse networks, storage precipitation gages (nonrecording) are sometimes installed at remote locations where information is needed. These gages require servicing no oftener than every 2 to 6 months. They are usually charged with oil and antifreeze solutions to retard evaporation and prevent freezing. Storage gages located in areas of heavy snowfall must be set with their orifices higher than the greatest snowpack expected.

Another action sometimes taken to obtain supplementary precipitation data is the "bucket survey" that is made immediately after intense rains occur in a basin. Special reconnaissance teams are sent into the field to collect all possible information from vessels or containers that may fortuitously have caught the heavier rainfall amounts. The absence of rainfall in some of the vessels helps delineate the boundaries of the storm-affected area. Vessels such as stock-watering troughs, old bathtubs (hopefully with their plugs well-seated), and milk buckets are investigated. It is necessary to exercise great care in measuring the depth of precipitation in these containers of varying cross-sectional area, and the measurements should be supplemented by judicious questioning of any who may have witnessed the storm.

In some specialized areal studies radar is used to delineate storm boundaries and estimate rainfall rates. Of the several types of radar in use today (1966), the Weather Bureau WSR-57 best meets the needs of the meterologist and hydrologist. The maximum range within which this instrument can detect all significant rainfall is about 125 miles, but intense rainfall centers can be detected beyond this range. The capabilities and limitations of the instrument are discussed by Wilson (1964) who states:

> "Quantitative data collected with the WSR-57 radar at Atlantic City, N. J. from five rainstorms and two snow-storms are compared with precipitation data from 60 recording rain gages within 100 mi. of the radar. Hourly rainfall amounts of from 0.01-0.02 inch are detected by the radar in at least 95 percent of the cases at all radar ranges out to 70 mi. Hourly amounts of from 0.04-0.05 inch are detected in at least 95 percent of the cases at all ranges out to 100 mi.
>
> "The relationship between radar echo intensity and rainfall rate varies from storm to storm. <u>Although the radar appears to have excellent potential for determination of area-average rainfall, reflectivity measurements provide only coarse estimates of point rainfall intensity</u>. The radar estimates of hourly rainfall averages, over a 750 sq mi area within 60 mi of the radar, are within the confidence limits of the average of 10 gage measurements, when a best-fitting radar-rainfall relationship is used for each storm. Use of one grand average relationship for all storms provides estimates of the average areal rainfall whose accuracy corresponds to those of a single rain gage located near the area center. . . .
>
> "Although the study shows that radar is a promising means for measuring areal average rainfall, further work is required to clarify the radars' ability to measure rainfall over areas of varying size and under varying synoptic situation, and to apply the findings to operational procedures. A dense network of rain gages within range of a WSR-57 will be valuable aid to such studies."

Some agencies--for example, the San Bernardino County Flood Control District in California--use aerial photographs to supplement their precipitation gage networks in delineating storm boundaries. If an area is flown shortly after rainfall has ceased there is usually a marked difference in color between wetted and unwetted soils. It is probable that even better results would be obtained by this method if infrared photography or imagery were used.

It was mentioned earlier that snow-course data may be used to estimate high-altitude precipitation. This, however, is not the primary purpose of snow courses. In high-altitude basins winter precipitation is stored as snow in mountain snowpacks for subsequent release as runoff in the spring when the pack melts. In those high-altitude basins that have economic importance, snow courses are often established for measuring the maximum accumulation of snow and its water equivalent. Experience has shown that significant snowmelt usually does not begin until about April 1, and it is on or near that date that measurements are usually made. Additional measurements are also made in some basins February 1, March 1, and May 1. The measurements not only supply data needed for study of the hydrologic budget of a basin, but they are used to predict the volume of snowmelt runoff in the months following the date of the snow survey. A publication by the U.S. Department of Agriculture (1940) gives detailed information on snow surveying.

## Soil Moisture

Soil moisture refers to all water contained in the unsaturated zone between the land surface and the ground-water table. Much of this water is returned to the atmosphere by evapotranspiration, a part is retained by the soil against the pull of gravity, and the remainder (if any) moves downward and reaches the water table. The physical properties of the materials composing the unsaturated zone control the water fluxes into, through and within, and out of that zone. The principal flux is the downward movement of water at the land surface, or infiltration. Water applied to the land surface, as by precipitation, in excess of the infiltration rate largely becomes surface runoff. Within the unsaturated zone upward and downward movement of water determines the residence time of water within the root zone and thereby influences evapotranspiration and ground-water recharge. Principles of soil-moisture occurrence are reviewed and summarized by Remson and Randolph (1962) and Stallman (1964).

Only in highly detailed basin studies is soil moisture measured. Reports by Johnson (1962) and the American Society of Civil Engineers (1964) discuss and evaluate the various methods of measuring soil moisture and describe the equipment needed. The authors of those two reports conclude that the neutron-probe method, despite some limitations, is the best method of making repeated measurements of soil moisture in place. That method is now widely used. Calibration of the neutron meter is important because, as Cotecchia and others (1968) warn, the meter reading for soil of a given water content may vary with the chemical composition of the soil.

Numerous sampling sites are needed to study the soil-moisture regime of a basin. The heterogeneity of the soil material and our lack of complete understanding of the hydrodynamics of soil-water systems are complicating factors in designing a network of sampling sites. Usually the sites are selected to sample the various soil complexes in the basin, but in the absence of obvious differences in soil, sampling may be random or by elevation zone. After a period of operation, study of the hydrologic budget of the instrumented basin will usually indicate a method of weighting the soil-moisture observations, and unneeded sites can be eliminated or additional sites can be added as needed.

Frequent soil-moisture observations over a long period of months may indicate a practical method of predicting the depletion of soil-moisture storage with time. Collings and Myrick (1965, p. 173) have had a measure of success in developing a recession curve of moisture stored in the upper 6 inches of the soil mantle at their project site in Arizona. Their recession curve is somewhat similar in form to the recession curve of base flow that is so familiar to surface-water hydrologists. It is likely that in a climate less equable than that of much of Arizona, climatic factors or an index obtained from atmometer or evaporation-pan observations would have to be considered in making soil-moisture predictions.

Usually the soil-moisture element of the hydrologic cycle is not measured but is deduced from measurements or inferred knowledge of the basinwide precipitation and streamflow. This can be done successfully in a mountainous basin where it is apparent that there is little or no underflow leaving the basin. If such is the case and the streamflow is ephemeral, it is assumed that all precipitation that does not become streamflow is held as soil moisture for subsequent evaporation or evapotranspiration. Minor losses, such as those attributable to the interception of precipitation by vegetation and the evaporation of precipitation held in depression storage, do not warrant separate attention in any studies other than those of a research nature. If there is no underflow but the streamflow is intermittent or perennial, the accretion to the ground-water body whose effluent supports the "fair-weather" streamflow, can be computed from the daily record of streamflow recession. The details of this computation are described in most standard hydrology texts (for example, Linsley, Kohler, and Paulhus, 1949, p. 396). The difference between precipitation and the sum of streamflow and the ground-water accretion can then be assumed to represent the soil moisture that is retained for subsequent evaporation or evapotranspiration. The above-mentioned methods can also be applied in areas where underflow occurs, if the magnitude of the underflow can be computed or estimated.

In ungaged mountainous basins of the type described above, where all the infiltrated precipitation is held as soil moisture, knowledge of the infiltration capacity of the soil can be used to determine soil moisture. We cannot determine absolute values of the infiltration capacity, but we can obtain index values at selected sampling points in the basin by means of infiltrometer tests.

Several types of infiltrometers are described in the literature. Wilm (1943) and Dortignac (1951) have described sprinkler type infiltrometers, but this type of infiltrometer is not hand portable and therefore cannot readily be used in areas inaccessible to surface vehicles. Johnson (1963) has described ring-type infiltrometers which are hand portable, but which disturb the soil that is tested. McQueen (1963), however, has developed a rainfall-simulator infiltrometer that not only is hand portable but does not disturb the soil, and its use is therefore recommended for studying the infiltration of rainfall in desert areas. Because infiltration figures obtained by infiltrometer are only index values, the instrument should be calibrated by making infiltrometer tests in small basins of known infiltration capacity; that is, in basins for which rainfall and runoff data are available. In using the infiltrometer at a site, two tests should be made--the first on dry soil after which the test site is mulched; the second should be made 3 days later to obtain values for the soil when the initial moisture content is at field capacity. By combining precipitation information with a knowledge of the infiltration capacity--deduced from infiltrometer tests and a calibration curve relating infiltrometer data to infiltration capacity--soil moisture can be estimated roughly for an ungaged basin.

The discussion in the three previous paragraphs refers to soil moisture in a mountainous basin upstream from the desert floor. When storm runoff in a stream channel is of sufficient magnitude to reach the desert floor, the stream usually loses water by seepage as it progresses downstream, and in a basin that is topographically closed and hydrologically undrained, ponding on the desert floor may occur. If ponding occurs, the underlying soil will become saturated. If the water table was originally very high, seepage will have raised the water table to the ground surface; if the water table was originally low, the added subsurface water will be perched above a soil layer of low permeability near the ground surface. In any case, all of the ponded water and most of the subsurface water added by the storm runoff will subsequently be evaporated or transpired by vegetation.

Evaluation of the evaporation of the ponded water will be discussed later under the heading "Evaporation from open-water surfaces;" this section of the report is concerned only with the evaluation of the subsurface component of the storm runoff. In a reconnaissance study there is little likelihood that this subsurface water that seeps into the desert floor will be identified as an item in the hydrologic budget separate from the remainder of the storm runoff. In a detailed basin study one or more wells on the desert floor would be used to monitor changes in the water table or piezometric head. One or more shallow test pits would be dug for visual inspection of the soil to determine the location of an impermeable layer, and if such a layer were found, a ring infiltrometer driven into the impermeable layer would be used to determine the infiltration rate of ponded water and the porosity (water-holding capacity) of the soil above this layer. The fluctuation of soil moisture in the upper part of the soil, when it is no longer saturated, might also be monitored using a neutron probe, but there may be some question as to whether or not this refinement is warranted.

## Evapotranspiration

In computing a hydrologic budget for humid regions, evapotranspiration is commonly considered the unmeasured residual that results when all other hydrologic elements are evaluated. This procedure is used in humid regions because evapotranspiration is generally the most difficult element to compute or measure directly. In arid regions, however, where all elements are difficult to evaluate accurately, it is helpful to have some means of estimating evapotranspiration as a check on the evaluation of the other elements of the hydrologic budget.

An excellent summary of methods of measuring or estimating evapotranspiration is found in a report by the Pacific Southwest Inter-Agency Committee (1966). Several of the methods, however, such as the energy-budget or mass-transfer (vapor-flux) method, require extensive instrumentation and study that goes beyond the scope of a basic data network and therefore those methods will not be discussed here. From a practical standpoint we may divide our problem into three parts: (1) Evapotranspiration by vegetation (mesophytes and xerophytes) that draws water entirely from that compnent of the precipitation that is stored as soil moisture; (2) evapotranspiration by irrigated crops where soil moisture is systematically replenished by water applied in accordance with crop needs; and (3) evapotranspiration by phreatophytes that not only consume soil moisture, but draw water from the underlying ground-water body. The preceding section of this report presented methods of computing soil moisture, and for arid areas where the vegetation draws all its water from soil moisture supplied by precipitation alone, it was pointed out that the quantity of soil moisture in storage at any time is considered equivalent to the subsequent evapotranspiration. We are left then with only the problems of computing water use by phreatophytes and by irrigated crops.

### Evapotranspiration By Phreatophytes

Water use by phreatophytes is the subject of many reports, notably those by Blaney (1961), Gatewood and others (1950), and Robinson (1958, 1964). Three methods of estimating water use are practical from the standpoint of a basic-data network--(1) transfer of results from evapotranspiration tank studies to nearby natural areas; (2) study of the diurnal fluctuation of stream stage or water table; and (3) empirical methods involving either (a) an index of evapotranspiration obtained from atmometer or evaporation-pan observations, or (b) such formulas as used in the Blaney-Criddle, Thornthwaite, Penman, Weather Bureau (modified Penman), or Lowry-Johnson methods. Four additional methods for computing water use by phreatophytes have been described by Gatewood and others (1950). Those procedures--the seepage-run, inflow-outflow, chloride-increase, and slope-seepage methods--require the collection of more data than are generally obtained in the operation of a basic-data network, and will not be discussed here.

Results of evapotranspiration-tank studies must be used with caution if they are to be transferred to a nearby natural area. Erroneous conclusions may be drawn if the microclimate of the tank differs from that of the natural area--the so-called oasis effect is too well known to require explanation. Error may also be introduced if the depth to water table in the tank differs appreciably from that in the natural area, or if there is appreciable difference in density and vigor of plant growth or foliage. Examples of the use of evapotranspiration-tank data in areal hydrologic studies have been reported by Gatewood and others (1950) and by Dylla and Muckel (1964). For his project area in Arizona, Gatewood used determinations of water consumed by phreatophytes grown in tanks under conditions that duplicated those found in the project area. Dylla and Muckel, in their study of water use by grasses in the flood plain of the Humboldt River, made use of the ratio of water consumed by tank-grown grasses to dry weight of the grass produced. This ratio was then applied to the known volume of hay produced by those meadows. Other evapotranspiration-tank studies of note have been reported by McDonald and Hughes (1968) and by Van Hylckama (1963, 1968).

The diurnal fluctuation of the stage of a stream during fair-weather periods provides a measure of the water loss by evaporation from the stream surface and transpiration by riparian vegetation. The fluctuations will generally be detectable only during periods of low stage, because it is usually at that time only that the stage-discharge relation of a stream is sensitive enough for the stage to respond noticeably to small changes in discharge. Dunford and Fletcher (1947) have shown that almost all fluctuation in stage was eliminated from a stream draining a 22-acre watershed by removing all streambank vegetation in a belt extending 15 feet above the water-surface level. The actual daily water loss resulting from evaporation and riparian transpiration can be roughly estimated as the difference in daily discharge as indicated by the observed fluctuating discharge hydrograph and the daily discharge as indicated by a straight-line hydrograph connecting the diurnal peaks (fig. 3). This method of estimating water loss has been reported by Croft (1948).

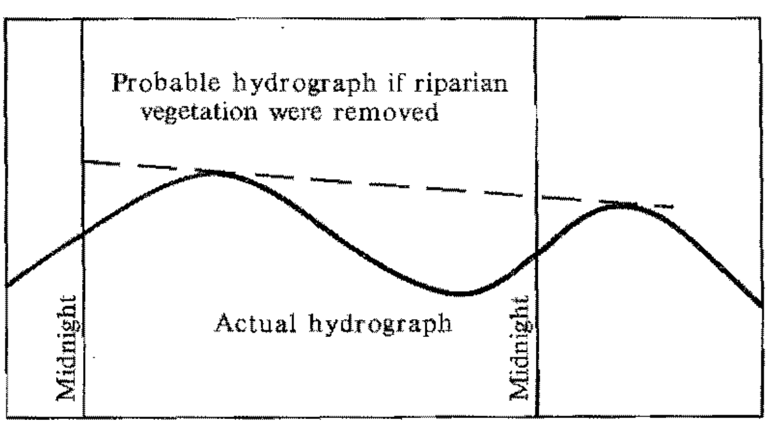



FIGURE 3.--Typical base-flow discharge hydrograph of stream whose discharge is depleted by transpiration of riparian vegetation.

The preceding method of estimating evapotranspiration by phreatophytes is applicable only for basins drained by perennial streams with narrow flood plains, as, for example canyon streams. For basins drained by intermittent streams with wide flood plains, it is necessary to study the diurnal fluctuations of the water table. The effect of transpiration by phreatophytes on the water table has been demonstrated in the past by several investigators, and more recently by Gary (1962). Gary showed that water-table fluctuations were significantly reduced after tamarisk and arrowweed were removed from an area, 25 feet in diameter, around a well site. The study of water-table fluctuations as a means of estimating evapotranspiration--the "transpiration-well method"--was popularized by the investigations of White (1932) and has been applied by several subsequent investigators, such as Gatewood and others (1950) and Robertson and Torrell (1963). White, in his work in Escalante, Utah, developed the formula,

$$q = y(24r \pm s)$$

where $q$ is the depth of water, in inches, used by the phreatophytes,
$y$ is the specific yield of the water-bearing material,
$r$ is the hourly rate of rise of the water table, in inches, from midnight to 0400 hours

and

$s$ is the net fall or rise of the water table, in inches, during the 24-hour period from midnight to midnight

The terms of the equation are illustrated in figure 4. The hours between midnight and 0400 hours were selected for determing the rate, $r$, because transpiration and evaporation are minimal during those hours. The effect of various ground-water conditions on the hydrograph of a well whose water level is affected by transpiration by phreatophytes is shown in figure 5.

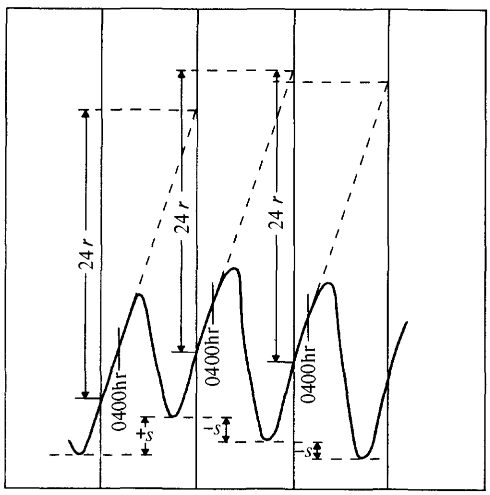



FIGURE 4.--Typical hydrograph of well whose water level is affected by transpiration of phreatophytes. (After White, 1932.)

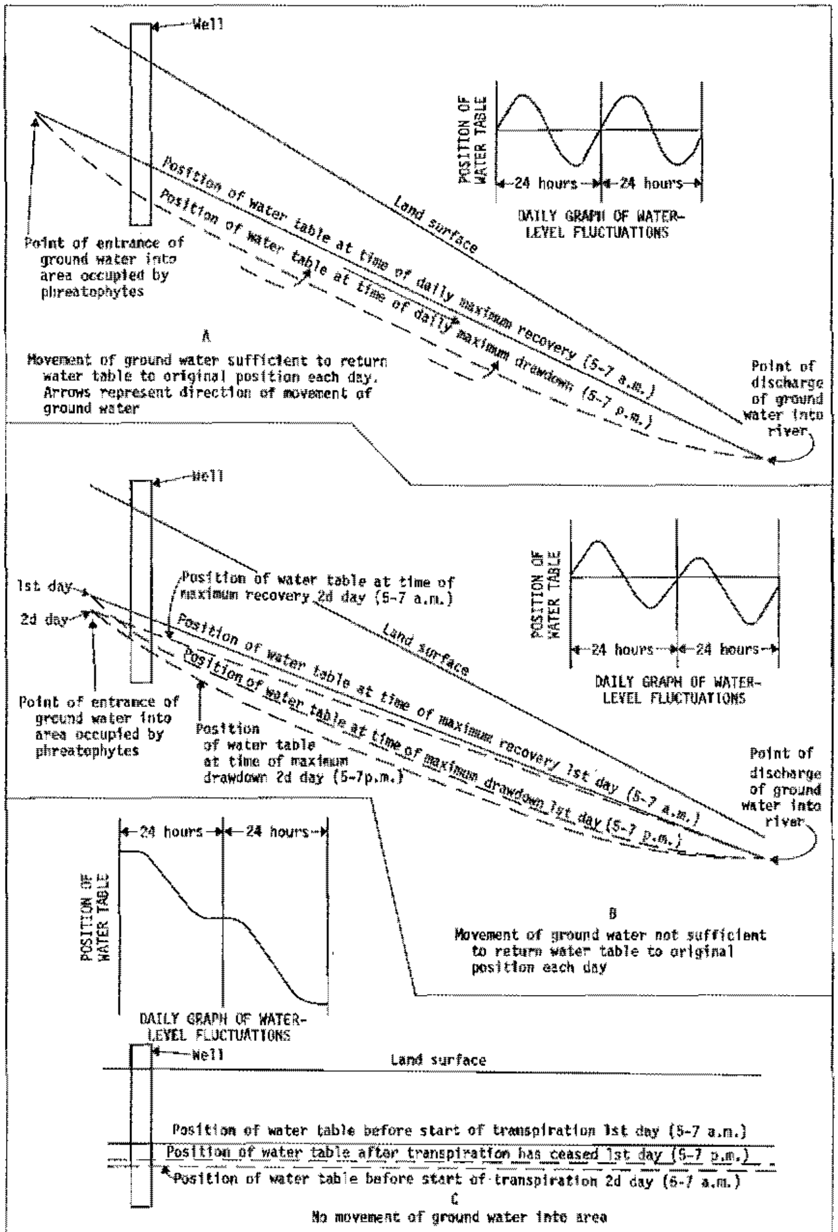



FIGURE 5.--Idealized sketch of section perpendicular to river channel, showing changes in position of water table as a result of daily transpiration (after Gatewood and others, 1950)

There are some shortcomings in the transpiration-well method. Gatewood and others (1950) point out that transpiration at night, which commonly is considered to be negligible in applying the method, may be significant and therefore require evaluation. Stallman (1961) indicates that changes in the water table that result from transpiration discharge may be dependent on the relative conductivities of the saturated and unsaturated zones as well as on changes in storage. Van Hylckama (1968) cautions that diurnal atmospheric pressure changes may also cause water-level fluctuations in a well and thereby bias the interpretation of the observed fluctuations. However, despite these shortcomings the transpiration-well method remains a valuable tool for making estimates of evapotranspiration by phreatophytes, providing, of course, that the specific yield of the water-bearing material has been determined or can be estimated. Troxell (1933), in his study of transpiration by phreatophytes in the Santa Ana River basin in California, had no knowledge of specific yield in the area, but he devised a modification of the transpiration-well method to compensate for this lack of information. Troxell's method not only has limited use but also has some theoretical weaknesses, and therefore is not described here.

Of the empirical methods of computing evapotranspiration by phreatophytes, the Blaney-Criddle formula is the most popular. Atmometer and evaporation-pan data are seldom used because the coefficients to be applied to those data are generally available only for irrigated crops. The other evapotranspiration formulas, such as the Thornthwaite, Penman, and Lowry-Johnson formulas, are not believed to give as good results as the Blaney-Criddle method, because they do not differentiate between vegetal species. Consequently methods other than the Blaney-Criddle will not be discussed here.

The Blaney-Criddle method is based on the assumption that with ample moisture available, evapotranspiration is affected primarily by temperature, duration of daylight, and vegetal species. For a complete description of the method, the reader is referred to a report by Blaney and Criddle (1962). In brief the Blaney-Criddle equation for evapotranspiration is:

$$U=K\sum\left(\frac{T\times p}{100}\right)$$

where $U$ is evapotranspiration, in inches, during the growing period,
$K$ is an empirical consumptive-use coefficient that is primarily dependent on the vegetal species,
$p$ is the monthly percentage of total daytime hours in the year,
and
$T$ is the mean monthly temperature, in degrees Fahrenheit.

Table 2 gives values of $p$ between lat 24° and 50°N. In using the equation, the monthly products of $T$ and $p$ are added for all months in the growing period.

TABLE 2.--*Monthly percentage of daytime hours of the year for lat 24° to 50°N*

[From Blaney and Criddle (1962, p. 43)]

| Month | Percentages for indicated degrees of latitude | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 24 | 26 | 28 | 30 | 32 | 34 | 36 | 38 | 40 | 42 | 44 | 46 | 48 | 50 |
| January | 7.58 | 7.49 | 7.40 | 7.30 | 7.20 | 7.10 | 6.99 | 6.87 | 6.73 | 6.60 | 6.45 | 6.30 | 6.13 | 5.98 |
| February | 7.17 | 7.12 | 7.07 | 7.03 | 6.97 | 6.91 | 6.86 | 6.79 | 6.73 | 6.66 | 6.59 | 6.50 | 6.42 | 6.32 |
| March | 8.40 | 8.40 | 8.39 | 8.38 | 8.37 | 8.36 | 8.35 | 8.34 | 8.30 | 8.28 | 8.25 | 8.24 | 8.22 | 8.25 |
| April | 8.60 | 8.64 | 8.68 | 8.72 | 8.75 | 8.80 | 8.85 | 8.90 | 8.92 | 8.97 | 9.04 | 9.09 | 9.15 | 9.25 |
| May | 9.30 | 9.37 | 9.46 | 9.53 | 9.63 | 9.72 | 9.81 | 9.92 | 9.99 | 10.10 | 10.22 | 10.37 | 10.50 | 10.69 |
| June | 9.19 | 9.30 | 9.38 | 9.49 | 9.60 | 9.70 | 9.83 | 9.95 | 10.08 | 10.21 | 10.38 | 10.54 | 10.72 | 10.93 |
| July | 9.41 | 9.49 | 9.58 | 9.67 | 9.77 | 9.88 | 9.99 | 10.10 | 10.24 | 10.37 | 10.50 | 10.66 | 10.83 | 10.99 |
| August | 9.05 | 9.10 | 9.16 | 9.22 | 9.28 | 9.33 | 9.40 | 9.47 | 9.56 | 9.64 | 9.73 | 9.82 | 9.92 | 10.00 |
| September | 8.31 | 8.32 | 8.32 | 8.34 | 8.34 | 8.36 | 8.36 | 8.38 | 8.41 | 8.42 | 8.43 | 8.44 | 8.45 | 8.44 |
| October | 8.10 | 8.06 | 8.02 | 7.99 | 7.93 | 7.90 | 7.85 | 7.80 | 7.78 | 7.73 | 7.67 | 7.61 | 7.56 | 7.43 |
| November | 7.43 | 7.36 | 7.27 | 7.19 | 7.11 | 7.02 | 6.92 | 6.82 | 6.73 | 6.63 | 6.51 | 6.38 | 6.24 | 6.07 |
| December | 7.46 | 7.35 | 7.27 | 7.14 | 7.05 | 6.92 | 6.79 | 6.66 | 6.53 | 6.39 | 6.23 | 6.05 | 5.86 | 5.65 |
| Total | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 |

A major difficulty presented by the Blaney-Criddle formula is the selection of the proper value of the all-important coefficient, *K*. This coefficient depends not only on the vegetal species, but also on the depth to the water table and on the vigor and density of growth. In addition, *K* has a regional variation because mean monthly temperature is only an index to the many climatic factors that affect evapotranspiration. In those parts of the arid Southwest, however, where the use of water by native phreatophytes is a significant factor in the hydrologic budget, the variation in *K* attributable to climatic factors is less important than the variation attributable to vegetal species, density of growth, and depth to the water table. Rantz (1968) after examining the available literature, prepared a graph, reproduced here as figure 6, which gives values of *K* for the growing season, for dense growths of various phreatophytes, and shows the variation of *K* with depth to water table. A *K* value of 1.30 is recommended for dense growths of hydrophytes, which are plants, such as tule and sedge, that live with roots wholly or partly submerged in water or in saturated soil that is intermittently submerged. Factors for adjusting *K* values for the effect of density of growth of both phreatophytes and hydrophytes are given in the following tabulation. These factors were derived from a report by Blaney (1954b, table 3).

| *Growth* | *Factor by which to multiply* K *value for density of growth* |
|---|---|
| Dense---------------- | 1.00 |
| Medium--------------- | .85 |
| Light---------------- | .70 |

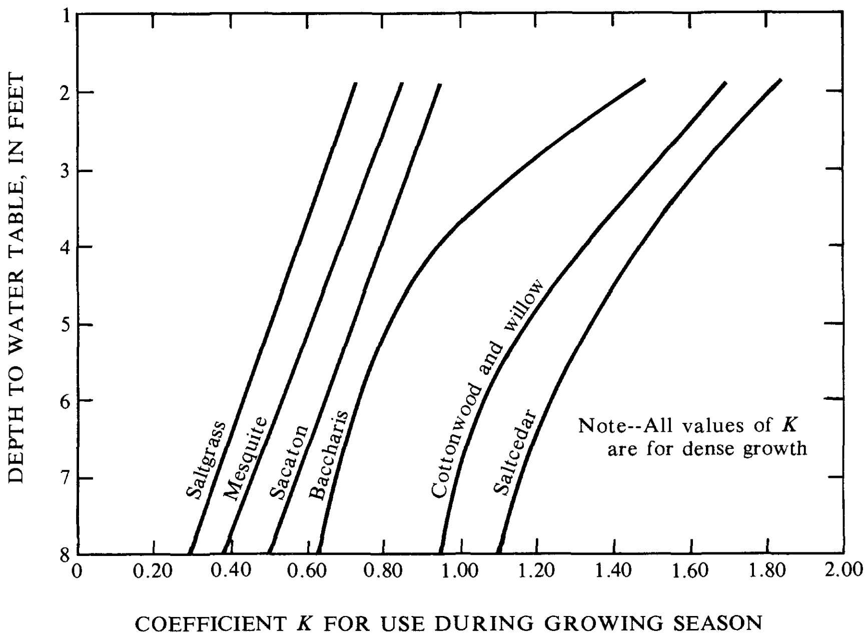



FIGURE 6.--Graph for estimating value of Blaney-Criddle coefficient $K$ in determination of water use by phreatophytes in Southwestern United States. (To be used only in absence of quantitative data at a site.) (After Rantz, 1968.)

Rantz states that subjective reasoning was used in constructing his figure 6 from the welter of conflicting data on evapotranspiration by phreatophytes. He advocates use of the curves only in the absence of quantitative evapotranspiration data, at sites where the time and expense required for a quantitative study are not warranted. Three additional comments concerning the use of figure 6 and the table of adjustment factors are pertinent.

First, the data on which figure 6 was based included no depths to water table greater than 8 feet. From the shape of the curves it is evident that the curves cannot be extrapolated downward indefinitely, but unfortunately the authors know of nothing in the literature, other than a statement by Van Hylckama (1963), that might be helpful in estimating values of $K$ where depths exceed 8 feet. Van Hylckama states that saltcedar at Buckeye, Ariz., does not grow or develop where the depth to water is 18 feet or more.

Second, density of growth as used in figure 6 and in the table of adjustment factors, represents a combination of two elements--thickness of foliage canopy and areal density of cover--and is expressed qualitatively as dense, medium, and light. While no greater refinement in defining growth characteristics was warranted for the subjectively produced figure 6 and accompanying table, the reader should be aware that quantitative methods of measuring the combination of vertical and areal density of cover have been standardized (Horton and others, 1964).

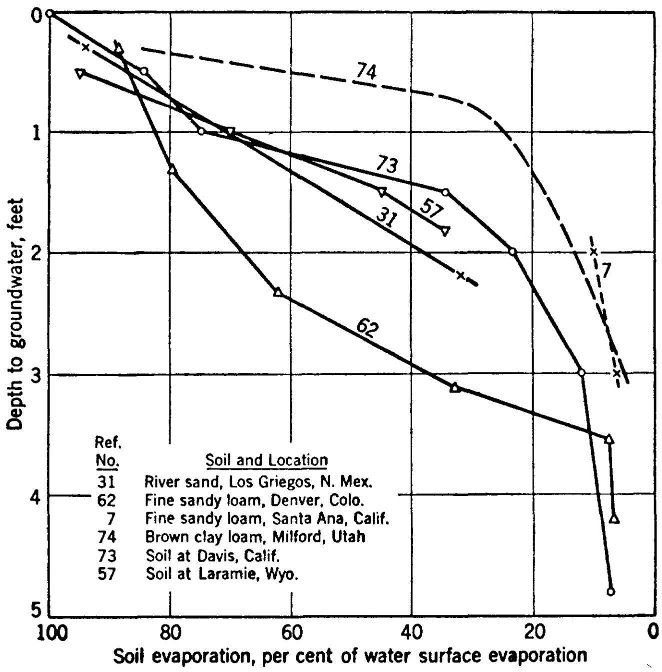
Third, the transpiration rate of phreatophytes decreases with increasing salinity of the moisture supply (Van Hylckama, 1966). Quantitative values of general applicability in defining the effect of salinity are not available.

## Evaporation from bare soil

The importance of being able to estimate reliably evapotranspiration by phreatophytes is apparent when it is realized, for example, that in a desert basin that is topographically closed and hydrologically undrained, evapotranspiration may represent the entire discharge, or water yield, of the basin. However, if the water table is high in the lower part of such a basin, appreciable evaporation of ground water through large areas of bare soil may occur. The amount of evaporation will depend on soil type and on depth to the water table, as well as on climatic conditions. Many measurements of ground-water loss through evaporation from bare soil have been made in Western United States. Houk (1951) derives the following general conclusions from these investigations:

1. When the water table is at the ground surface, the rate of evaporation from sand or sandy loam is approximately equal to the rate of evaporation from a free water surface.
2. Rates of evaporation from bare soil decrease rapidly with increasing depth to the water table, particularly in the first 2 feet of depth. This is especially true of sandy soils.
3. In most soils, rates of evaporation from ground-water supplies become extremely low when the water table falls to a depth of more than 4 feet. However, even an extremely low rate of evaporation may cause an appreciable loss of water if the evaporation is effective over a large area.

These conclusions are based largely on the data in figure 7. The figure, however, can only be used as a guide because evaporation rates can be seriously affected if the soils contain appreciable quantities of salts. Tests at the Utah Agricultural Experiment Station, reported by Harris and Robinson (1916), showed that rates of evaporation from Greenville loam were decreased about 50 percent when 5 to 7 percent of sodium chloride was mixed with the soil. Van Hylckama (1966) also warns of the effect of salinity on evaporation from bare soil, but his data are insufficient to provide quantitative information of general applicability. To use figure 7 it is necessary that the evaporation from an open-water surface be known; the determination of that evaporation is discussed in a following section (p. 31).

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74. White, W. N., 1932, A method of estimating ground-water supplies based on discharge by plants and evaporation from soil--Results of investigations in Escalante Valley, Utah: U.S. Geol. Survey Water-Supply Paper 659-A, 105 p.

FIGURE 7.--Variation of evaporation from bare soil with soil type and depth to water table. (After Houk, 1951).

Ripple and others (1970) recently devised a procedure in which climatological data are combined with equations of water transfer through soil to estimate the steady-state evaporation from bare soils under high water-table conditions. Data required to apply the method include soil-water retention curves, depth to water table, air temperature and humidity, and wind velocity. The effects of thermal transfer and salt accumulation are neglected.

## Evapotranspiration By Irrigated Crops

The quantity of water used by irrigated crops is usually computed by using evaporation as an index of evapotranspiration, or by using empirical formulas such as those mentioned earlier in connection with water use by phreatophytes. Where evaporation is used as an index to evapotranspiration, the evaporation data used are usually either that measured in a U.S. Weather Bureau class-A pan, or the "net evaporation" for a Livingston atmometer as given by the difference between measured evaporation from the black and white spheres of the atmometer. The relation between evaporation and evapotranspiration for each plant species is expressed as a simple ration--either a pan coefficient or an atmometer coefficient--for each month or for each 10-percent increment of the growing season. The coefficients for any particular plant species also vary areally in response to differences in environmental setting. The evaporation-index methods of computing evapotranspiration are discussed in reports by Shannon (1968) and Hargreaves (1968).

The two most commonly used empirical formulas for computing water use by irrigated crops are the previously described Blaney-Criddle formula and the Jensen-Haise formula (Jensen and Haise, 1963). It will be recalled that the Blaney-Criddle formula required the following data for each month of the growing season: (1) A monthly percentage of daytime hours of the year (p) based on latitude, (2) mean monthly air temperature (t), and (3) a monthly crop coefficient (k). Values of p, t, and k for various crops and locations in the United States are given in a report by Blaney and Criddle (1962). The monthly percentage of daytime hours of the year for any latitude is an index of the solar radiation received; the monthly air temperature is an index of the sensible heat flux; and the crop coefficient varies monthly to reflect the stage of plant development and varies areally as a result of areal differences in the climatic and soil environment.

The Jensen-Haise formula uses solar radiation as the principal index of evapotranspiration. A report by Jensen and Haise (1963) presents ratios of evapotranspiration ($E_t$) to solar radiation ($R_s$) for the four regions into which the authors divided Western United States--Columbia Basin area, Northern Plains States, Southern Plains States, and Southwest States. The authors state that the ratios ($E_t/R_s$) "represent the combined effects of reflectance or albedo, and relative effects of effective thermal radiation, sensible heat flux to the soil and air, plus other minor components at various stages of crop growth." For orchards, ratios are given for each 10 percent of the growing season, or for each month where the growing season lasts all year, as in parts of the Southwest. For field crops, the authors divided the growing season into two periods. The first period starts with planting and ends at the growth stage when evaporating and transpiring surfaces are no longer the limiting factor in the vaporization of water. The second period begins with the aforementioned growth stage and lasts until maturity of the crop. For the first period, ratios are given for each 20-percent increment of the period; for the second period, ratios are given for each successive 10-day interval after the beginning of the period. (Field crops are usually harvested within 70 days after the start of the second period.)

To compute the evapotranspiration for a given crop, on a given date, in a given location, the appropriate ratio from tables in the Jensen-Haise report is multiplied by the historical mean daily solar radiation received on that date. Before performing the multiplication the daily solar radiation, which is recorded in cal per sq cm (calories per square centimeter), is transformed to equivalent inches of water evaporated per day (1 inch of water = 1500 cal per sq cm). Tables of solar radiation are summarized in the report for 20 locations in Western United States, and procedures are given for estimating solar radiation for other areas where only limited climatic data are available.

## Evaporation from Open-Water Surfaces

Open-water bodies most commonly found in arid regions fall in three categories--water-supply reservoirs on large streams that originate in humid highlands, small stock ponds, and storm runoff temporarily ponded on desert floors.

The evaporation from large water-supply reservoirs is generally studied in detail using methods that are well established and well documented in the literature. Any of the following four methods may be used.

1. Water Budget.--This method requires the measurement of precipitation on the lake surface, inflow from streams or effluent ground water, outflow to streams or influent ground water, and the change in volume of stored surface water. Evaporation is the residual that results from the algebraic summation of these elements. Often there is the added complication of bank storage, which is extremely difficult to evaluate. Langbein (1954) discusses the effect of bank storage on the effective capacity of Lake Mead. Harbeck and Kennon (1954) demonstrate the use of the water budget in computing evaporation from Lake Hefner in Oklahoma.

2. Energy budget.--This method requires the measurement of incoming and reflected long- and short-wave radiation, net energy advected into the water body (including that of precipitation but excluding that of evaporated water), and the increase in energy storage of the body of water. The residual obtained by the algebraic summation of these elements represents the sum of the three remaining elements in the energy budget--energy utilized by evaporation ($Q_e$) plus energy advected by evaporated water ($Q_w$) plus net energy conducted from the body of the water to the atmosphere as sensible heat ($Q_h$). $Q_e$, the element sought, is then computed by expressing $Q_w$ and $Q_h$ in terms of $Q_e$. The relation of $Q_w$ to $Q_e$ is dependent solely on the water temperature, and the relation of $Q_h$ to $Q_e$ is dependent on the Bowen ratio. The use of this method at Lake Hefner is described by Anderson (1954) and its use at Lake Mead is described by Harbeck and others (1958).

   One might digress at this point to mention that while the energy budget gives good results for computing the evaporation from an open-water surface, it is usually unsatisfactory for computing evapotranspiration from a vegetated surface because of the difficulty in calculating the highly significant Bowen ratio (Leppanen, 1961).

3. Mass transfer.--This method, which is probably preferable to the preceding methods, is based on the aerodynamics of vapor flux. Its use in the Lake Hefner study is described by Marciano and Harbeck (1954). The empirical equation used is

$$E=Nu(e_o-e_a),$$

where $E$ is evaporation,
$N$ is a coefficient,
$u$ is average wind speed,
$e_o$ is vapor pressure of saturated air at water-surface temperature,
and
$e_a$ is vapor pressure of ambient air.

3. Mass transfer.--Continued
   Wind speed is measured directly, and vapor pressures are determined from measurements of air and water temperature and relative humidity. There are two methods of evaluating the coefficient, $N$. One method requires that the reservoir be instrumented with the equipment needed for either the energy-budget or water-budget method. This equipment is operated simultaneously with the equipment needed in the mass-transfer method during a calibration period lasting a year or two. $N$ is then computed from the evaporation as determined by either the energy budget or water budget, and thereafter this value of $N$ is used routinely in the mass-transfer equation. The second, but less reliable, method of obtaining $N$ was derived by Harbeck (1962) and is simply the use of the equation

$$N=0.00338/A^{0.05}$$

   where $A$ is the area of the reservoir surface, in acres.

4. Combination of mass-transfer and energy-budget methods.--This method, developed primarily by Penman (1948), is widely used. The equations are quite complicated, but they have been reduced to a family of curves by Kohler and others (1955, fig. 6), and the graphical solution of the equations is relatively simple. The elements that must be measured to apply the method are air temperature, relative humidity, wind speed, and incoming solar radiation. The method is basically weak because it ignores energy storage and advected energy.

The four methods described above permit lake evaporation to be determined for short periods of time, as for example, weekly or monthly evaporation. If the evaporation amounts desired are merely totals for the entire year or for the 6-month period, May through October and November through April, the only instrumentation needed is one or more evaporation pans installed near the reservoir, and a water-stage recorder to indicate lake levels so that the area of the open-water surface at any time is known. Regional pan coefficients for the entire year and coefficients to convert annual evaporation to evaporation for the period May through October are given by Kohler and others (1959,pls. 3 and 4). These coefficients, when applied to annual pan evaporation, will give annual and 6-month totals for the lake that should be accurate within 15 to 20 percent. The annual pan coefficient when used with pan evaporation for short periods of time will invariably give erroneous results. Total annual lake evaporation is a conservative element in the sense that the annual evaporation in any year usually will not vary more than 10 percent from its mean value.

As mentioned in the preceding section, the evaporation from an open-water surface must be known if figure 7 is to be used for estimating evaporation from bare soil. The annual evaporation from an open-water surface may be obtained from plate 2 of the previously cited report by Kohler and others (1959), or it may be obtained by installing an evaporation pan in the area and applying the annual pan coefficient from plate 3 of the Kohler report to the annual pan evaporation.

The contours on plate 2 of the Kohler report are highly generalized because of the small scale of the map, and because there are insufficient evaporation stations and first-order Weather Bureau stations in the mountainous West to permit significant refinement of the isopleths shown. Therefore annual evaporation read from plate 2 is only a rough estimate. It is known that evaporation decreases with increased altitude in response to the decreasing air temperature, but with increasing altitude such factors as increasing wind speed and decreasing vapor pressure of the air tend to compensate for the decreasing air temperature. Consequently the decrease in evaporation with altitude is not as great as one would expect from a consideration of air temperature alone. A study of the effect of altitude on evaporation in the Wasatch Mountains of Utah has been reported by Peck (1967).

The plates in the Kohler report are generally of no help in estimating the evaporation from stock ponds, because the water usually remains in storage for short periods only. However, in any hydrologic study involving stock ponds a water budget for the reservoir, however crude, is kept. A crude budget may involve no more than the installation of a crest-stage gage on the reservoir and on the spillway, and a precipitation gage and reservoir staff gage that are read weekly. This instrumentation will provide data concerning precipitation, surface-water inflow, spill, and change in reservoir contents. The residual that results from the algebraic summation of these measured or partly estimated elements of the water budget is water loss, or the sum of reservoir evaporation and seepage. Langbein and others (1951) describe such an investigation in Arizona and demonstrate a method of separating evaporation from seepage (fig. 8). Monthly water loss is plotted against monthly pan evaporation at a nearby site, a practice that is permissible when the stock tanks are shallow and do not retain water for appreciably long periods of time. The y-intercept of a straight line fitted to the plotted data represents minimum seepage. Seepage is assumed to increase with temperature and, therefore, with monthly evaporation. The increase in rate of seepage is made proportional to the change in viscosity of the water as deduced from air temperatures at some nearby Weather Bureau station. This method of determining evaporation is admittedly crude but is adequate for reconnaissance studies.

Where the hydrology of stock ponds is studied in detail the reservoirs are generally instrumented so that the mass-transfer method of computing evaporation can be applied. With evaporation known, seepage from the reservoir is usually computed as the residual item in the water budget. Periods of negligible surface inflow or outflow to the reservoir are usually used in the computation, and the computed seepage is adjusted for change in water viscosity as indicated by air temperatures. Notable examples of detailed studies of stock pond hydrology are those by Culler (1961a), Gilbert and others (1964), and Kennon (1966). In the study by Kennon, evaporation was determined by the energy-budget method.

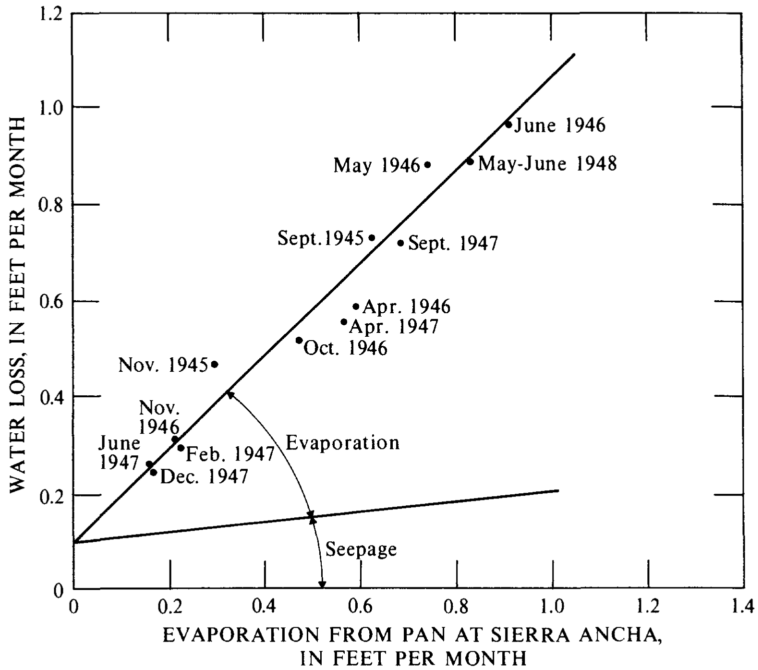


FIGURE 8.--Evaporation and seepage from Postoffice Tank, Ariz. (After Langbein and others, 1951.)

As for storm runoff that ponds on the desert floor, it usually does so because soil layers of low permeability are present at shallow depths or because of a high water table. Under those conditions all storm runoff that reaches the desert floor, including that portion of the water that saturates the upper soil layers, will be evaporated in a relatively short time. In a reconnaissance study it will only be necessary, therefore, to estimate the average annual volume of ponded water to determine the amount of evaporation. This is not easily done, particularly in areas that are not traversed by roads, and therefore are not viewed regularly, as for example, by road maintenance men. If ponding is a fairly regularly occurrence, there may be evidence of its occurrence in the form of salt incrustations on nonporous objects above the ground surface and the depth and volume of water involved may therefore be estimated. Salt incrustations on the desert floor itself or on porous objects such as wooden stakes can be misleading, however, because they can form in the absence of surface water, as a result of evaporation from a high water table. In short, if ponding is not a highly significant element in the disposition of storm runoff, open-water evaporation probably cannot be evaluated in a reconnaissance study. If ponding is a significant element there may be visual evidence of its occurrence and open-water evaporation may appear in the hydrologic budget for an undrained basin as the difference between water yield of the basin (to be discussed later in this report) and outflow from the basin as represented by evapotranspiration from the soil and ground-water body.

In a detailed basin study, one or more crest-stage gages would be installed on the desert floor to indicate the extent and volume of the ponded storm runoff. The approximate rate of evaporation would be deduced from periodic observations of the ponded water surface or from observations of the nearest evaporation pan. The results obtained by this procedure will not achieve a high degree of accuracy, but maintaining more elaborate instrumentation over a period of years for this phase of hydrologic study is generally not warranted. The disposition of the water that seeps below ground, either to be held as soil moisture or to reach a high water table, has already been discussed under the headings, "Soil Moisture" or "Evapotranspiration."

## Water Yield

One of the prime objectives of most basic-data studies is to determine the water yield of a basin. Water yield, sometimes referred to as recoverable water or the water crop, is defined as the sum of runoff (streamflow adjusted for man-made diversions and storage, if any) and underflow (ground-water outflow that bypasses the surface outlet channels and leaves the basin by an underground route). Because streamflow and ground-water storage and movement are interrelated it is difficult to completely isolate them for separate discussion of the investigational techniques used for each. For those situations where sharp demarcation exists between surface- and ground-water aspects of the study of water yield, investigational techniques for each are discussed in later sections titled "Surface Water" and "Ground Water," respectively. In this section of the report, water yield is discussed for those situations where separation of the two components is impractical, as it is in most reconnaissance studies.

In most reconnaissance studies average annual water yield is sought. The basin being investigated will usually have either no streamflow records or records of insufficient length to provide information concerning the long-term yield. Frequently the determination of yield is confined to that part of the drainage basin that lies upstream from the site where the stream debouches from the canyon and begins to lose significant amounts of water through the streambed. The runoff at that site commonly represents the basin yield because there usually is little contribution to the yield from area downstream from the site and little underflow upstream from the site. The average annual water yield is commonly estimated on the basis of average annual basinwide precipitation and an inferred rainfall-runoff relation based on records for basins in a similar environment. Estimates of evapotranspiration loss may provide a rough check on the reliability of the results obtained. The pages that follow describe 12 techniques that have been used to estimate average annual water yield. The methods discussed are valid only where there is no underflow and the term "runoff," where used, is therefore equivalent to total yield. For convenience the individual methods are usually identified in this paper by the name of the author of a formal report in which the method is described. Most of the methods have evolved over the years and their development generally is not attributable to any one investigator.

## Streamflow-Correlation Method

If short-term streamflow records are available for the basin under investigation, they may correlate with concurrent long-term records for nearby streams and thereby provide a means of estimating long-term mean annual runoff. In the absence of streamflow records in the study basin, miscellaneous discharge measurements may be obtained and similar correlation attempted with concurrent discharges for streams with long-term records.

Riggs (written comm., 1968) has described an alternative method for use with ungaged sites, the first step of which is to estimate the mean monthly and annual discharge for a single year at the ungaged site. In a second step this single annual mean discharge is used to obtain the long-term mean. Riggs' method requires that one miscellaneous measurement per month--preferably in the middle of the month--be made at the ungaged site during one complete water year. Then, on the assumption that the ratio of concurrent daily mean flows of two streams at about the middle of the month equals the ratio of their mean discharges for that month, the ratios of the measured discharges to the concurrent discharges at a nearby long-term station are computed. The monthly mean discharges at the long-term station are multiplied by the corresponding ratios to obtain the monthly mean discharges for the ungaged site. The computed monthly discharges are then averaged to give the annual mean discharge for that single year. The next step is to estimate the long-term mean from that single annual mean. That is done by using records for all nearby stations that have long-term records. The mean discharges at those stations for the single year under consideration are correlated with their long-term mean discharges. The resulting relation is used with the single annual estimate for the ungaged site to obtain its long-term mean discharge.

The correlation methods described above usually yield satisfactory results in investigations in humid regions, but are much less likely to do so in arid-land investigations. The limitations of the correlation technique are discussed at the end of this section of the report.

## Crippen Method

Crippen (1965), using long-term hydrologic data for southern California, obtained a relation between average annual potential evapotranspiration ($E$) and altitude, where $E$ is represented by annual pan evaporation data converted to equivalent lake evaporation by use of the appropriate pan coefficient. Separate relations were obtained for desert and coastal environments. From an isohyetal map of average annual precipitation he obtained a relation between average annual precipitation ($P$) and altitude. A hypsometric (altitude versus drainage area) curve for each basin was then used to obtain basinwide values of $P$ and $E$. These values of $P$ and $E$ were used with long-term records of runoff ($R$) to construct a graph of $P/E$ versus $R/E$ (fig. 9).

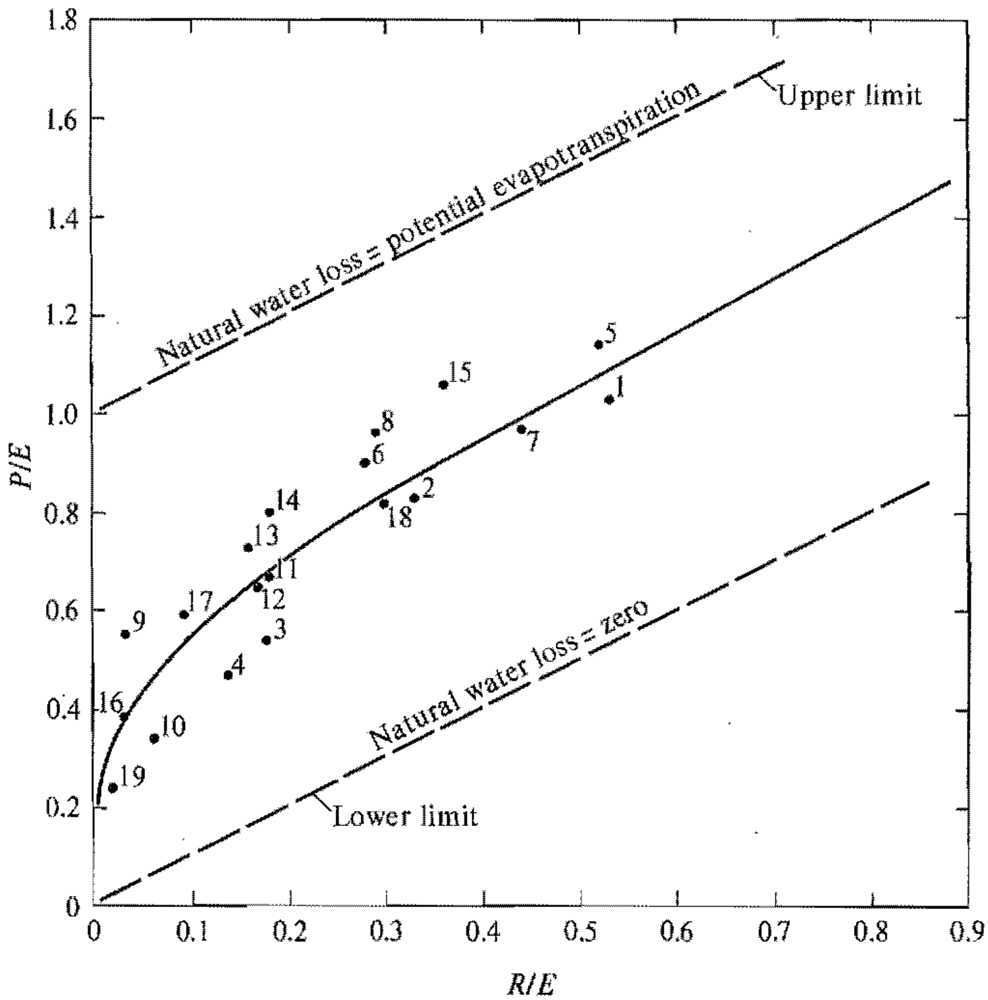


FIGURE 9.--Relation of $P/E$ to $R/E$. (After Crippen, 1965.)

Each of the plotted points in figure 9 represents a gaged basin. Crippen explained the scatter of points in figure 9 on the basis of differences in surficial rock type. He obtained a measure of the scatter by computing a coefficient, $K$, for each basin, which when multiplied by the curve value of $R/E$ gave the computed or plotted value of $R/E$. From a geologic map he determined the percentage of each basin predominantly underlain by each of six rock types. By cut-and-try he obtained a "retentivity number," ranging from 0 for very low permeability to 100 for very high permeability, to assign to each rock type. The sum of the products of percentage of basin times retentivity number gave a geologic index, $I$, for each basin. Values of $I$ were then plotted against $K$ in figure 10. Two curves were fitted to the plotted points--one for coastal basins and the other for the few desert basins for which Crippen had data.

To obtain the mean annual water yield for an ungaged basin in southern California by the Crippen method, an isohyetal map of mean annual precipitation is first used to obtain values of $P$ for each 1,000-foot altitude zone. Crippen's evaporation-altitude curve (not shown here) is then used to obtain values of $E$ for each elevation zone. The quotient, $P/E$, is applied in figure 9 to obtain the quotient, $R/E$, for each altitude zone. Because $E$ is known, the value of $R$, unadjusted for the effect of differences in basin geology, can be computed for each zone. These values of $R$ are then weighted by percentage of area in each altitude zone to give the total unadjusted average annual yield of the basin. The coefficient, $K$, needed to adjust the basin yield figure for the effect of geology is obtained by entering figure 10 with the geologic index, $I$, for the basin. The method of computing $I$ was explained in the previous paragraph.

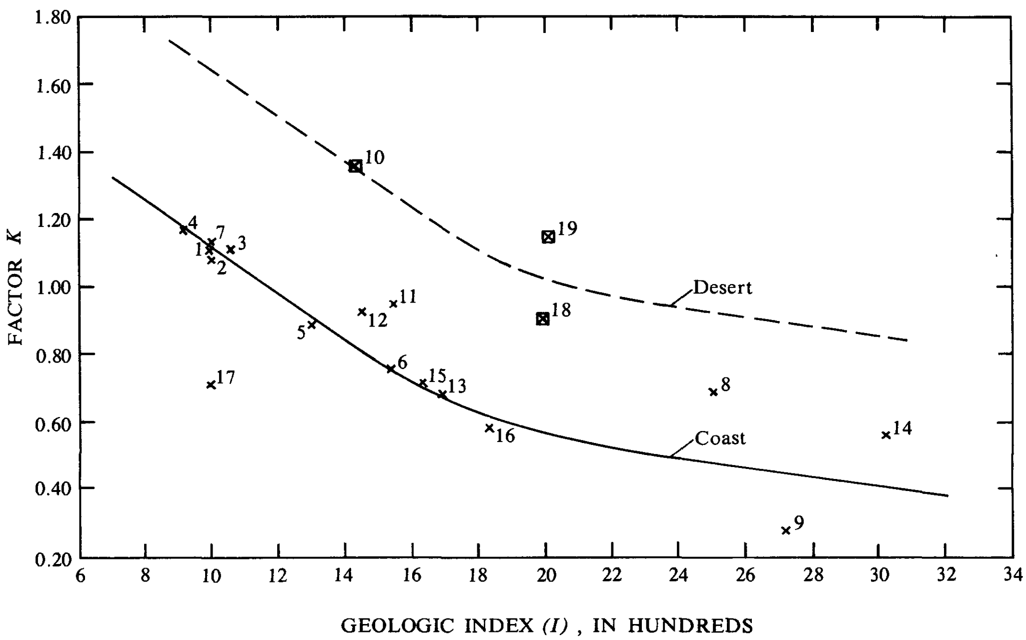


FIGURE 10.--Relation of adjustment coefficient ($K$) to geologic index ($I$). (After Crippen, 1965.)

## Langbein-Nace Method

Langbein and Nace (Nace and others, 1961, p. 29-47) used a similar technique in their investigation of the Raft River basin in Idaho and Utah. As in the Crippen method, the essential graph is one of $P/E$ versus $R/E$. There are two principal differences between the two methods. First, Langbein and Nace did not compute $E$ from evaporation pan data, but instead they averaged the values of annual evapotranspiration computed by four empirical formulas--Penman, Thornthwaite, Lowry-Johnson, and Blaney-Criddle. Second, they made no adjustment to the value of $R$ for the effect of differences in basin geology, because there was no indication that such an adjustment was needed.

## Langbein Method

In a later paper, Langbein (1962) modified the method used in the Raft River basin study. Instead of computing potential evapotranspiration, $E$, for use in a graph of $P/E$ versus $R/E$, he used a temperature factor, $e$, in a graph of $P/e$ versus $R/e$ (fig. 11). The temperature factor, actually an evaporation index, is computed from the formula

$$e = 10^{(0.015T)}$$

where $T$ is mean annual temperature, in degrees Fahrenheit. The temperature, $T$, is obtained from records for nearby Weather Bureau stations, adjusted to the altitude of the study basin by the temperature lapse rate in the standard atmosphere (-3.3° F per 1,000 feet of altitude). In place of a single curve, as in figure 8, Langbein used a family of curves (fig. 11), based on the seasonal distribution of precipitation among the months of the year. The seasonal precipitation factor is derived by dividing the mean annual precipitation by the products of the mean monthly precipitation times monthly coefficients that vary sinusoidly. These coefficients are listed in table 3. As in the study of the Raft River basin (Nace and others, 1961, p. 29-47), the mean annual yield of the basin was not adjusted for the effect of geologic factors. The method of determining the average annual yield of an ungaged basin by either of the two Langbein methods is similar to that described for the Crippen method.

TABLE 3.--*Coefficients for calculating seasonal precipitation factor*

(After Langbein, 1962)

| North latitude | Oct. | Nov. | Dec. | Jan. | Feb. | Mar. | Apr. | May | June | July | Aug. | Sept. | Sum |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 50° | 0.75 | 0.25 | 0.12 | 0.06 | 0.18 | 0.54 | 0.96 | 1.55 | 2.00 | 2.27 | 2.00 | 1.32 | 12.0 |
| 40° | .87 | .43 | .30 | .27 | .39 | .72 | 1.05 | 1.42 | 1.71 | 1.87 | 1.71 | 1.26 | 12.0 |
| 30° | .98 | .60 | .48 | .48 | .60 | .90 | 1.14 | 1.30 | 1.42 | 1.48 | 1.42 | 1.20 | 12.0 |

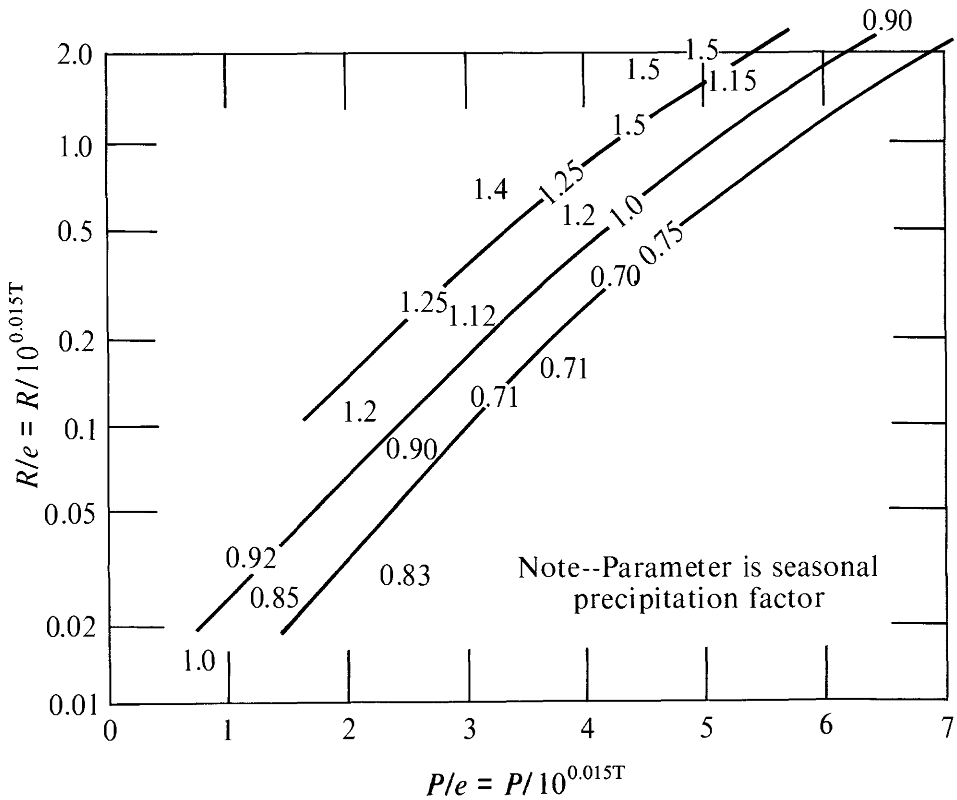


Figure 11.--Relation of *R/e* to *P/e*. (After Langbein, 1962.)

## Bagley Method

Bagley and others (1964) used a multiple-regression technique for determining the yield of ungaged basins in Utah. They derived regression equations for each of three regions of hydrologic homogeneity into which they had divided the State. The dependent variable used in each equation was long-term average annual runoff. The independent variables used included the following: Average annual basinwide precipitation (from an isohyetal map of Utah), average basin altitude, latitude of the centroid of the basin, longitude of the centroid of the basin, a basin-slope factor, average land-slope in the north-south direction, drainage density, aspect or azimuth angle of the drainage basin, a weighted geologic factor (similar to the geologic index used by Crippen), and a weighted vegetative-type factor. The use of other physical parameters, such as maximum altitude, streambed slope, and average land slope without regard for direction, were considered but were not found to be statistically significant. The report by Bagley and others presents regression equations for various combinations of the independent variables, along with such pertinent statistics as correlation coefficients, standard errors of estimate, and confidence limits.

The regression equations were used to prepare a map showing isopleths of average annual runoff for Utah. The Bagley report states:

> "In areas where records of natural runoff were available the isolines were distributed by use of appropriate equations previously developed. The runoff was then computed for comparison with the actual amount measured. Isolines were adjusted appropriately until measured and computed yields were in close balance.
>
> "For areas not having streamflow records, locations of isorunoff lines were largely determined by use of the better relations which included elevation and precipitation. The previously located lines in the areas having runoff data served as additional 'bench-marks' to guide the location of isorunoff lines in areas having no data with which to check results."

In other words, the authors of the report do not recommend any particular combination of independent variables, other than to recommend that average annual precipitation and altitude be included. They apparently tried several of the equations; for gaged areas they used those equations whose results best fit the known basinwide runoff, and for ungaged areas they used those whose results best matched the isopleths already drawn for adjacent gaged areas.

A multiple-regression method, similar to that used by Bagley, has been used successfully by the Geological Survey for several years in humid regions to relate runoff characteristics to physiographic and climatic parameters. It is now (1969) being tested in arid regions.

### Riggs-Moore Method

Riggs and Moore (1965) describe a method of estimating mean annual runoff at ungaged mountain sites in Nevada, using streamflow records and topographic maps. For regions where precipitation data are available, graphs are constructed relating mean annual precipitation to altitude, and these precipitation-altitude relations are used as guides to delineate regions of hydrologic homogeneity. Where precipitation-altitude relations are unavailable, the tentative boundaries enclosing a region adjudged to be homogeneous are first selected on the basis of topography. Mean annual runoff values are then derived from streamflow records, and parts of the region that appear to produce anomalous runoff values are excluded. In each homogeneous region the basins are divided into 1,000-foot altitude zones, and a runoff value is arbitrarily selected for each zone. Precipitation-altitude relations, where available, provide a guide for the selection of these zonal runoff values, and it is assumed that no runoff is contributed by areas on the valley floors--areas of low precipitation lying below the 5,000-foot altitude contour. In some selected gaged basin the zonal runoff values are weighted in accordance with the percentage of area in each altitude zone to give mean annual basinwide runoff. If the computed runoff for the basin does not agree with the runoff obtained from runoff records, the zonal runoff values are adjusted until agreement is achieved. This procedure is repeated for other gaged basins in the region and an average set of zonal runoff values is finally derived that provides values of mean annual runoff that are consistent for all basins in the region. Moore (written communication, 1969) states that he has now identified eight homogeneous regions that cover the entire state of Nevada, and has prepared runoff-altitude relations for each.

To apply the Riggs-Moore method to the computation of mean annual runoff for an ungaged basin in one of the eight homogeneous regions, it is necessary only to determine the area of the basin that lies within each 1,000-foot altitude zone and to apply the appropriate zonal runoff value to each of these increments of area.

### Oltman-Tracy Method

Oltman and Tracy (1949) used a method similar to that of Riggs and Moore in their study of the Wind River basin in Wyoming. They used records of precipitation and runoff to prepare graphs relating average annual precipitation to altitude, and average annual water loss to altitude. Annual water loss was computed as the difference between precipitation and runoff. With these two graphs and a hypsometric curve for an ungaged basin that related drainage area to altitude, they could readily compute the average annual runoff for the ungaged basin.

## Mundorff Method

Mundorff and others (1964) had an adequate isohyetal map of mean annual precipitation available for their study of the eastern Snake River basin in Idaho. They were therefore able to use a more direct approach than that used in either the Riggs-Moore or Oltman-Tracy methods. They correlated the mean annual precipitation of several gaged watersheds in the basin with the mean annual yield of those watersheds. They then applied the derived precipitation-yield relation to the mean annual precipitation for the ungaged areas being studied and thereby obtained the mean annual yield from those areas.

In a commonly used refinement of the Mundorff method each gaged basin is divided into about 10 precipitation zones, a yield value is assigned to each zone from a trial precipitation-yield relation for the region, and basinwide yields are then computed by weighting the yield values in accordance with the percentage of area in each zone. If the computed basinwide yields do not agree with the yields obtained from streamflow records, the trial precipitation-yield relation is adjusted and the procedure is repeated until agreement is reached. Then to obtain the yield of an ungaged basin, one first determines the percentage of basin area that lies within each of the precipitation zones into which the basin has been divided, and he then applies the regional precipitation-yield relation to the several increments of area. This refinement of the Mundorff method is recommended because it takes into consideration the fact that the precipitation-yield relation is curvilinear, particularly at low values of precipitation. If, in an arid or semiarid basin, the relation is applied to average basinwide values of precipitation--as Mundorff did--rather than to incremental values of precipitation, an additional source of error is introduced into the estimate of yield.

## Hely-Peck Method

Hely and Peck (1964), in their study of the lower Colorado River basin in Arizona and California, devised a method of estimating runoff from ungaged desert areas, using an isohyetal map of mean annual precipitation and daily precipitation records for seven widely spaced desert stations. Their analysis of the seven precipitation records, which are considered to be independent of each other, is summarized in table 4. The data in columns 2 and 5 of the table are assumed to represent the distribution of precipitation anywhere in the study region. For example, table 4 shows that 10 percent of the mean annual precipitation occurs in storms that have depths averaging 0.08 inch, and 24 percent occurs in storms that have depths averaging 0.25 inch.

The next step was to convert this precipitation distribution to equivalent average annual yield. This was done by using a modification of a method described in publications by the U.S. Soil Conservation Service (1957) and the U.S. Bureau of Reclamation (1960).

TABLE 4.--*Average distribution of daily precipitation at seven desert stations*

[After Hely and Peck, 1964]

| Class limits of daily precipitation (inches) | Representative storm rainfall (inches) | Percentages of mean annual precipitation | | |
|---|---|---|---|---|
| | | Maximum | Minimum | Rounded mean |
| 0.01-0.10 | 0.08 | 13.1 | 7.5 | 10 |
| .11- .30 | .25 | 25.8 | 21.3 | 24 |
| .31- .60 | .53 | 26.8 | 24.3 | 26 |
| .61-1.00 | .9 | 25.4 | 12.4 | 18 |
| 1.01-1.50 | 1.4 | 14.2 | 7.2 | 11 |
| 1.51-2.00 | 1.9 | 6.7 | 2.5 | 5 |
| >2.00 | 2.9 | 13.0 | 1.0 | 6 |

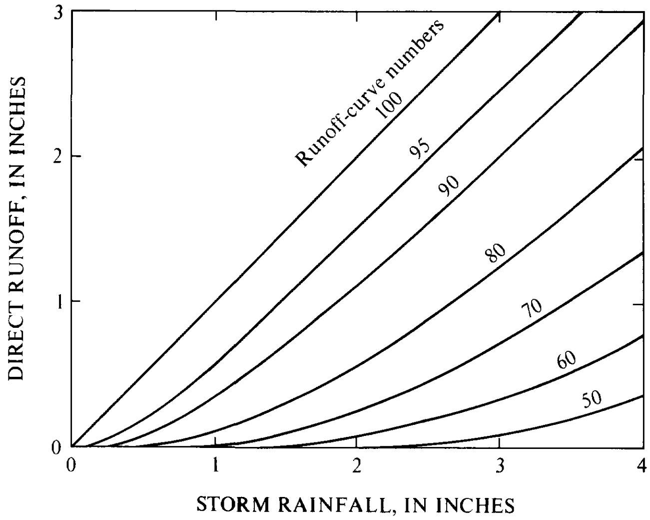


FIGURE 12.--Relation of direct runoff to storm rainfall and runoff-curve number. (After U.S. Soil Conservation Service, 1957.)

The principal element of the method is a family of curves, an abridged version of which is shown in figure 12. The ordinate scale in figure 12 represents direct, or storm, runoff, but because base flow is usually only a very small component of the yield of desert streams, direct runoff and yield were considered to be equivalent. The curves are numbered from 0 to 100, in order of increasing runoff. Curve 0 would apply to a sand or gravel so permeable that no direct runoff would occur for any rainfall. Curve 100 represents the unattainable condition of 100-percent runoff. The curve numbers between 0 and 100 are determined by a formula and are not percentages. The precipitation distribution in table 4 was applied to the curves of figure 12 to derive the graph in figure 13. Figure 13 shows the percentage of mean annual precipitation that will appear as mean annual runoff for any particular runoff-curve number, or soil complex, in the region. It remained for Hely and Peck to devise a method of relating the runoff-producing characteristics of subareas in the region to runoff-curve numbers.

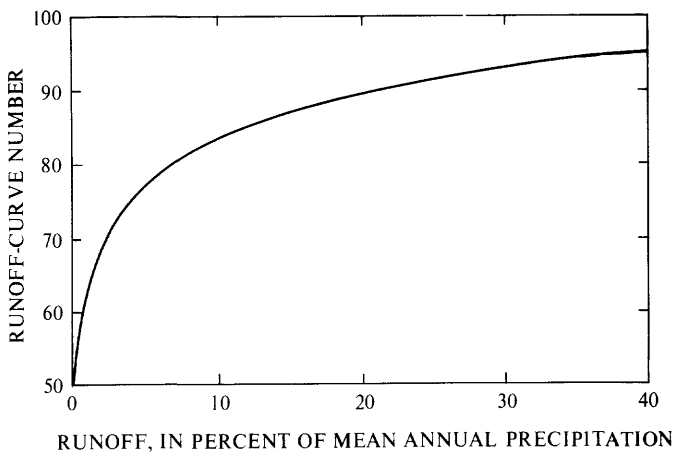


FIGURE 13.--Relation of runoff-curve number to mean annual runoff expressed as a percentage of mean annual precipitation. (After Hely and Peck, 1964.)

The method devised by Hely and Peck to accomplish this involved the determination of an infiltration index for the various subareas. Infiltrometer tests were run at nearly 100 test sites in the region by use of a portable infiltrometer of the rainfall-simulator type (McQueen, 1963), and these tests were supplemented by several hundred observations of the behavior of water poured into shallow depressions. The infiltration index selected for use--initial infiltration, in inches per half hour--was then related to runoff-curve number (fig. 14) on the basis of some subjective reasoning, coupled with a fairly intimate knowledge of the soils in the region

In applying the technique of Hely and Peck to determine average annual runoff from an ungaged area, the curves in figure 14 are first used for selecting the appropriate runoff-curve number for a specific subarea. This runoff-curve number is then applied to the graph in figure 13 to obtain the mean annual runoff, expressed in percentage of mean annual precipitation. This percentage, when multiplied by the mean annual precipitation, as determined from a regional isohyetal map, gives the mean annual yield, expressed in inches. The procedure is repeated for all subareas in the basin to obtain mean annual yield for the entire basin.

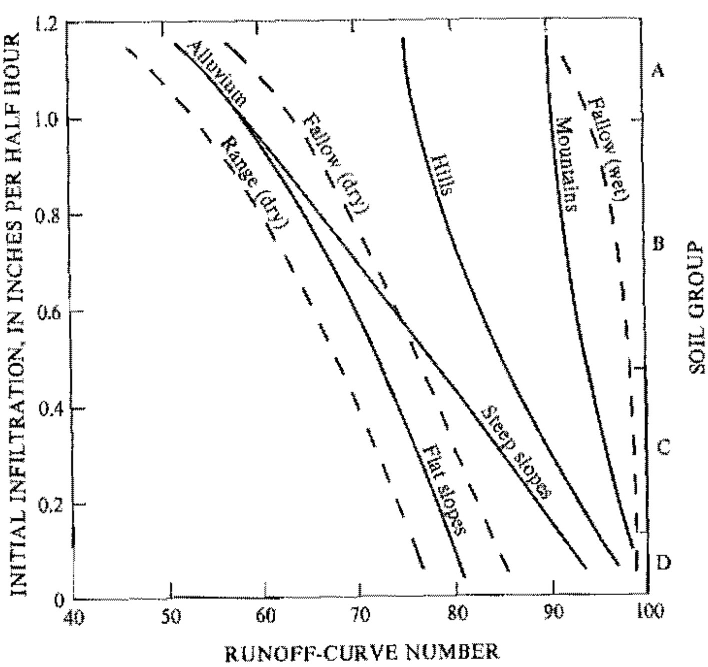


FIGURE 14.--Relations between initial infiltration and runoff-curve numbers. (After Hely and Peck, 1964.)

## Burnham Method

Burnham and others (1963), in an investigation of the water resources of San Nicolas Island, Calif., used a detailed hydrologic budget for determining average annual yield from an ungaged region. Their situation was somewhat favorable, in that they had 23 years of daily precipitation record for the island, which has an average annual precipitation of 8.8 inches. Aside from the precipitation data, all elements in the budget were inferred from prior studies of nearby coastal areas having similar climate and soils. A rainfall-runoff relation derived for storms on the mainland (fig. 15) was used for the island. From a comparison of mainland and island soils, a limiting value of soil-moisture deficiency, in inches, was determined for the four hydrologic units into which the island was divided. Evapotranspiration on the island was inferred from monthly pan evaporation on the mainland. The monthly pan figures were converted to daily average values for each day of the month, and these daily figures were in turn converted to equivalent lake evaporation by use of an appropriate pan coefficient. The resulting values were assumed to be the evapotranspiration that occurred during each non-rainy day of a particular month, as long as soil moisture was available in sufficient quantity to support the potential rate. As a final step before computing the hydrologic budget, the entire daily precipitation record for the 23 years was divided into periods of rainy and non-rainy days, which were tabulated chronologically.

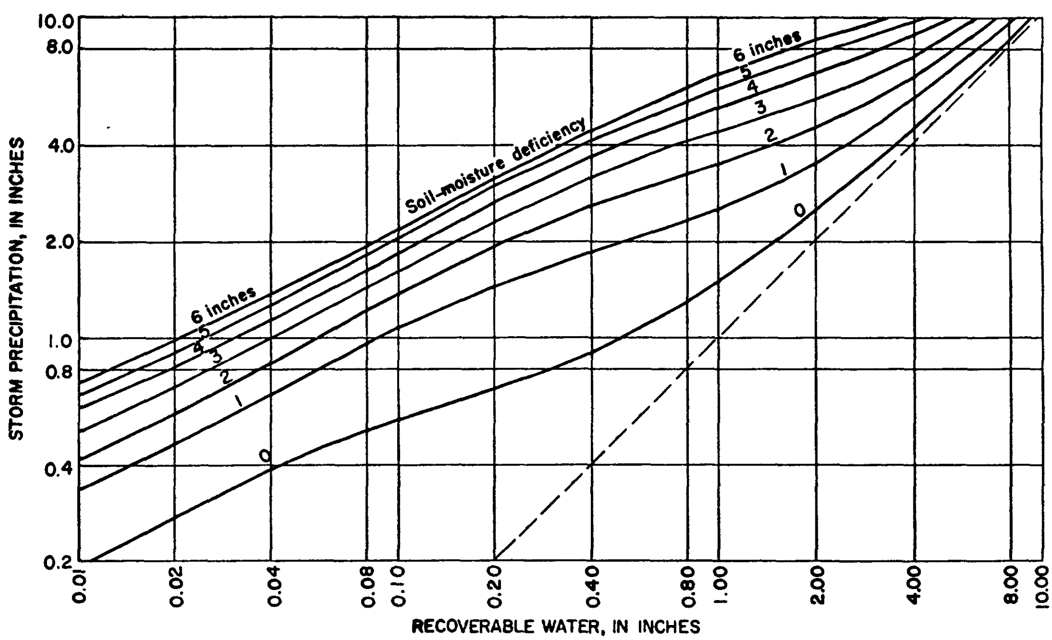


FIGURE 15.--Relation between storm precipitation, soil-moisture deficiency, and recoverable water on San Nicolas Island. (After Burnham and others, 1963.)

A continuous water inventory was prepared, using the data that had been assembled in graphs and tables. The soil-moisture deficiency was assumed to be at its maximum on the first day of the inventory. During rainy periods the precipitation was distributed to first provide for initial loss (interception) and soil-moisture replenishment, and the remaining precipitation represented recoverable water, or water yield. During rainless periods depletion of the soil moisture occurred in response to the evapotranspiration demand.

## Channel-Geometry Method

Langbein and Moore (Moore, 1968) are exploring the possibility of estimating mean annual runoff from the geometry of desert stream channels. The basic assumption of the method is that channel dimensions adjust themselves to the streamflow, and the width and average depth of the channel at point bars in meandering channels, or at island bars in braided channels, are indicators of the mean annual runoff. The datum to which depth is referred and along which the width is measured is the top of the bar that is used. W. B. Langbein (written commun., 1966) states:

> "The bed forms that are the basis of survey are usually also related to vegetal zones. Viewed in midsummer, three vegetal zones may be recognized: 1) the 'in-channel' is usually free of vegetation; 2) the zone between the level defined by the tops of the point and island bars and the flood plain is usually occupied by annuals (forbs and grasses); 3) the true flood plain is occupied by shrubs, some species of which may be phreatophytic.
>
> "In the arid country the point bars are at a level much below the bankfull stage, as the channel may be incised in the alluvium. In humid country the point-bar level may be equivalent to that of the true flood plain, and the intermediate zone may be absent. In straight channels the reference level is formed by island bars if the stream is braided. The tops of those island bars are usually armored and conformable with each other and with berms that form along the banks, especially in protected places. The crests of some island bars along perennial streams may be sprouting shrubs and becoming permanent or fixed in place. The reference level is then at the base of the shrubs."

Langbein goes on to state:

> "Because widths and mean depths will vary greatly among the different sections as the river exchanges width for depth and vice versa, or passes from pool to riffle, at least five sections should be measured. A straight arithmetic average of widths and mean depths at the several sections is reported for the reach. These sections should cover a reach or length of river channel equivalent to 100 river widths."

Moore (1968) empirically derived a family of curves relating mean annual discharge to width and depth and has successfully tested these curves at gaging stations in Nevada and east-central California.

## Evapotranspiration Method

The evapotranspiration method provides a means of making crude estimates of the average annual yield of undrained basins whose only discharge is through evapotranspiration on the desert floor. This evapotranspiration discharge is computed by use of figures 6 and 7. To use these figures, it is necessary that the depth to the water table be known, as well as the acreages occupied by phreatophytic vegetation and bare soil overlying a shallow water table. An additional water loss to be considered is the estimated evaporation loss from ephemeral channels that carry the water from canyon mouth to desert floor (Sorey and Matlock, 1969). Usually the evapotranspiration method is used only as a rough check on the yield as estimated by one or more of the methods described in this section of the report. If the yield as estimated from evapotranspiration discharge is significantly less than the yield indicated by another method, an investigation should be made of the possibility that storm runoff ponds on the desert floor and later evaporates. If such is the case, the additional evaporation from the open-water surface and from the soil lying between the ground surface and the "normal" water table would be added to the original estimate of evapotranspiration. An investigation should also be made of the possibility of error in the assumption of an undrained basin; there may actually be underground movement of water out of the basin. If the yield as estimated from evapotranspiration discharge is significantly greater than the yield indicated by another method, an investigation should be made of the possibility of ground water moving into the basin beneath topographic boundaries, either through secondary openings in consolidated rocks or as a result of difference between topographic and phreatic (ground-water) divides.

## Discussion of Methods

Twelve methods of making reconnaissance-type studies of mean annual yield have been described. None of these methods has a high degree of accuracy. The reasons for this lack of accuracy include the following:

1. The runoff from adjacent basins is usually poorly correlated because of the great variability of storm precipitation with respect to both time and area.
2. Yield is a small residual of the precipitation and its computation even from adequate hydrologic data is subject to large percentage error.
3. The necessary climatological data are scarce.
4. Only a few of the many pertinent physiographic and climatic factors are considered in most methods.
5. The results may be affected by significant movement of ground water into or out of the basin that is studied.

Because any one of the methods may give unreliable results, the use of at least two of the methods is recommended when a reconnaissance study is made. The last method described--evapotranspiration from the desert floor--is relatively simple to apply and will often be applicable as a rough check on whatever other method is used. When the use of more than one method gives results that are at variance with each other, the hydrologist will have to weight the results in accordance with his best judgment to arrive at a final figure of basin yield.

The user of the methods should be aware of their limitations. The streamflow-correlation method in which short-term records are correlated with concurrent long-term records for nearby streams is often unsatisfactory because, as previously stated, there is great variability in storm precipitation. Even in the case where both streams are fed by snowmelt and runoff does not fluctuate as widely as it does where storm runoff predominates, the correlation of runoff for periods shorter than a year is often poor. As for correlating, or otherwise relating, miscellaneous measurements of discharge with concurrent discharges for streams with long-term records, the same problem of poor correlation exists and is compounded by the fact that momentary or hourly discharges are even less likely to provide satisfactory correlation than discharges for longer periods, such as days or months. That is so because discharges of magnitude sufficiently large to define the mean annual discharge generally occur during periods of rapidly changing flow.

The next three methods discussed--the Crippen, the Langbein-Nace, and the Langbein methods--used a relation of the form, $P/E$ versus $R/E$. Benson (1965) warns that a spurious correlation may result when ratios with a common denominator are used, if the coefficient of variation of the common denominator is large with respect to that of each of the numerators. In other words, it is possible for the curve of $P/E$ versus $R/E$ to fit the plotted points closely even though $P$ and $R$ may be poorly related to each other. That may be the case if the coefficient of variation of $E$ is relatively large. Preferably, a family of curves of $P$ versus $R$, with $E$ as the parameter for the family, should be constructed. However, that requires the development of a relation among the three variables from basin and storm characteristics in the area, and usually there are too few gaged basins in the area for that purpose.

In these same three methods, various techniques are used for determining the potential evapotranspiration factor, $E$ (or $e$, as used in the Langbein method). The best measure of potential evapotranspiration is probably given by pan-evaporation data adjusted by regional coefficient to equivalent lake evaporation. In the absence of pan-evaporation data, such data can best be synthesized by the Weather Bureau (modified Penman) method, but rarely will all the climatological data required by the method be available for the study area (see p. 33). If an empirical formula for potential evapotranspiration must be used, the Blaney-Criddle formula is recommended, this recommendation being based on a study by Cruff and Thompson (1967). The Blaney-Criddle formula was discussed earlier in this report (p. 25). H. F. Blaney (written commun., 1966) recommends that the values of $K$ shown in table 5 be used for Southwestern United States.

TABLE 5.--*Values of* K *in the Blaney-Criddle formula for use in computing potential evapotranspiration in Southwestern United States*

| Climate | Values of *K* for period indicated | |
|---|---|---|
| | Water year | April through September |
| Hot, extreme aridity | 0.85 | 1.00 |
| Cooler, subhumid | .75 | .90 |

The multiple-regression techniques used in the Bagley method to relate water yield to precipitation and to many physiographic parameters has a weakness in that the interrelation of the many independent variables used may produce regression coefficients that have little physical significance. Any extrapolation of the relations must therefore be made with caution.

As for the Riggs-Moore, Oltman-Tracy, and Mundorff methods, their reliability depends to a large degree on the actual homogeneity of the regions.

The Hely-Peck method requires that base flow be an insignificant part of runoff because the relation in figure 12, which is used to obtain water yield, was derived for storm runoff. As for the procedure used to relate infiltrometer data to runoff-curve number, a more objective approach would have been to make infiltration tests in small gaged areas where the runoff-curve numbers were known and thereby relate infiltrometer data to curve number.

The Burnham method is probably the least practical of those described, in that it requires both a great deal of detailed hydrologic data from a nearby area of similar hydrologic characteristics and a great deal of subjective reasoning.

The channel-geometry method is attractive because of its simplicity. However, the channel properties measured are related to some discharge greater than the mean annual discharge, and therefore the reliability of the channel-geometry method is dependent on the consistency of the relation of this higher discharge to the mean annual discharge.

As for the evapotranspiration method, it was stated earlier that this method at best gives only crude estimates of yield, and it should be used only as a rough check on estimates made by one or more of the other methods described.

## SURFACE WATER

### Problems in Determining Discharge

There are numerous problems in determining discharge in arid regions that are rot commonly encountered in humid areas. The problems are associated primarily with the following conditions: (1) Unstable channels, (2) water loss in channels, (3) flash flows, (4) pulsating flow, and (5) mudflows. Before discussing these problems individually, it should be mentioned that some of them can be overcome by instrumenting an existing onstream reservoir or stock pond with a stage recorder and computing inflow from rated or gaged outflow, water loss, and change in reservoir contents.

Stream channels in alluvium, especially in materials of small particle size, are not stable and change configuration in response to changes in discharge or water temperature. The streambed changes may affect the operation of intakes of stream-gaging stations, but, more important, the stage-discharge relation changes with variation in streambed conformation. A conventional control structure placed in the stream to stabilize the stage-discharge relation will create a pool, but the pool usually fills completely with sediment in a short time, and if the structure is low it may be buried by the sediment. In either event, the conventional control structure becomes ineffective for stabilizing the stage-discharge relation. However, existing culverts often act as satisfactory controls for gaging small streams, and their use for that purpose should not be overlooked. Control structures are discussed in more detail on page 56.

Another problem is associated with the loss of water in stream channels of high permeability. A stream-gaging station at the mouth of a canyon may record perennial flow, including discharges of considerable magnitude, while no surface flow appears at the downstream end of the alluvial fan. Streamflow stations are usually installed at sites where the bulk of the surface flow produced in the basin can be gaged, because this is the simplest and most reliable way to determine the yield of the basin. However, such sites are not always available. Often a large part of the streamflow originates in numerous springs, much of whose flow seeps into the ground before reaching the main channel. It may not be economically feasible to gage many of the springs near their sources. Renard and Keppel (1966) and others have reported dramatic reductions in peak discharge as a flood wave progresses down an alluvial channel. This creates problems in flood-frequency studies, because only in the most detailed basin studies will it be feasible to relate the reduction of peak discharge to the area of channel wetted by the flow, downstream from that part of the basin that contributes runoff. Pratt (1961) has developed an empirical equation for predicting loss in volume of sustained floodflow between two gaging stations on the lower Gila River in Arizona. His equation is of the form

$$y = ax^b$$

where $y$ is cumulative percentage loss, $a$ is percentage loss on the first day, $x$ is day of flow at the downstream gage, and $b$ is an empirical exponent. An equation of this form may have applicability for other intermittent desert streams.

Storm runoff that occurs in stream channels as flash flow, meaning streamflow that swiftly rises and recedes, presents a problem to the hydrologist. The suddenness of the event makes it virtually impossible for the hydrologist to plan his arrival at the gaging station to coincide with the occurrence, and, should he by chance be on the scene when flash flow is in progress, the rapidly changing discharge makes it exceedingly difficult for him to obtain a reliable measurement of discharge. The seriousness of this is apparent when it is realized that most of the runoff of many small desert streams occurs in these flash flows. There are, however, indirect methods of computing peak discharge from debris lines along the streambanks that mark the crest elevation of the flow. These methods of discharge determination are not valid, however, for pulsating flow.

Pulsating flow is a manifestation of unstable flow conditions in which a series of translatory waves of perceptible magnitude develops and moves rapidly down the channel. These translatory waves are commonly called roll waves or slug flow. If the overriding wave carries an appreciable part of the total flow, conventional methods cannot be used to determine discharge. The occurrence of a single wave, or of perhaps a few waves, may indicate that the waves result from successive arrivals of flood peaks from different upstream tributaries. However, when the waves occur continuously, at somewhat regular intervals measured in minutes or fractions of a minute, their occurrence is associated with the hydraulics of the channel itself. Roll waves or slug flow, of the latter type, have been reported by many observers--for example, Holmes (1936), Jahns (1949), and Leopold and Miller (1956)--and the senior author has observed them in the Santa Anita Wash flood control channel in Arcadia, Calif., during a period of controlled releases of constant discharge from an upstream reservoir. In desert floods the wave front and succeeding flow may include all gradations of fluvial material, from nearly clear water to mudflows.

A mudflow is defined, for the purpose of this report, as a well-mixed mass of water and alluvium of high viscosity and low fluidity, which generally moves at much slower rates than clearer water would. The mechanics of mudflows are described in detail by Woolley (1946) and by Sharp and Nobles (1953). The proportion of water to solid material in mudflows may vary widely, and it is virtually impossible to determine the quantity of water involved in the flows. Where the percentage of water to solid material is relatively large, the water may move the material as bedload; where the percentage of water is low, the entire mass may become a viscous mixture whose behavior is similar to that of wet concrete. According to Woolley (1946) large boulders that appear to float in the mixture are actually transported by a rolling and sliding action whereby the weight of the boulder seen at the surface is transmitted to the streambed through a layer of well-lubricated silt, sand, gravel, and smaller boulders, which may be likened to a mass of ball bearings. Where the water-sediment mixture is not a viscous mass, but acts as a Newtonian fluid, the discharge of the mixture can be computed by the methods used for clear water. The percentage of water in the flow will not be known, however, unless samples of the flow can be taken for analysis. Where the mixture is a viscous mass and acts as a non-Newtonian fluid, there is no way of determining the rate of discharge of the mass, other than by simultaneously observing the cross-sectional area occupied by the flow and timing the movement of debris on the surface of the mass. The discharge value so obtained will, of necessity, be crude. Channel surveys made before and after the mudflow may give some indication of the total quantity of solid material transported.

## Determination of Discharge from Stage-Discharge Relation

The usual method of obtaining a continuous record of discharge is by periodically determining the discharge at observed stages to define a stage-discharge relation. This relation is then applied to a continuous record of stage, obtained by a recording gage, to determine the continuous record of discharge. The water-stage recorder may either be float-operated or pressure-operated, and it may record either on an analog strip chart or on a digital tape. Discharge measurements are made either by standard current meter, optical current meter (Smith and Bailey, 1962), or by the dye-dilution method (Cobb and Bailey, 1965). The latter two methods are used when measuring conditions are unfavorable for use of the standard current meter.

The discharge measurements, however obtained, usually will not define a stable stage-discharge relation for alluvial channels because of changes in elevation and configuration of the streambed. The elevation of the streambed does not remain constant; it usually changes continually in response to scour and fill. Commonly, the bed elevation fluctuates about some equilibrium datum, but after a major flood there may be progressive scour or fill until some new equilibrium datum is reached. These changes in elevation of the bed result in changes in the stage-discharge relation, but they may have little effect on the depth-discharge relation. A stable depth-discharge relation is helpful in defining the basic shape of the stage-discharge relation, but, because the actual elevation of the fluctuating streambed is known only at the time of measurements, there is usually some question about the magnitude of changes to the stage-discharge relation between measurements.

The problem of a shifting stage-discharge relation is compounded by changes in the configuration of the streambed in response to changes in discharge or water temperature (Dawdy, 1961, and Simons and Richardson, 1962). By configuration we mean the shape of the bed profile, classified with respect to the flow regime. An idealized diagram of the bed configurations in alluvial channels associated with various regimes of flow is shown in figure 16. Configurations A-C are classed as "lower regime," configuration D as "transitional regime," and configurations E-H as "upper regime." Because resistance to flow varies greatly as discharge and streambed configuration vary and because the rate of development of the various configurations depends on the rate at which the discharge changes, several different depth-discharge relations are possible at a single station. In view of the uncertainties concerning bed elevation and configuration, it is not customary to consider all possibilities in constructing the relation between depth and discharge. Usually one average curve is drawn for upper regime flows and another for lower regime flows, the two curves being separated by transitional regime flows. The general effect is illustrated in figure 17, where hydraulic radius has been plotted against velocity. The average streambed elevation is used with the depth-discharge relation to obtain an average stage-discharge relation. This relation is shifted with respect to stage on the basis of discharge-measurement data and the judgment of the hydrologist, to obtain the stage-discharge relation that is effective at any given time or date. The state of the art has not yet progressed to the point where subjective reasoning can be eliminated.

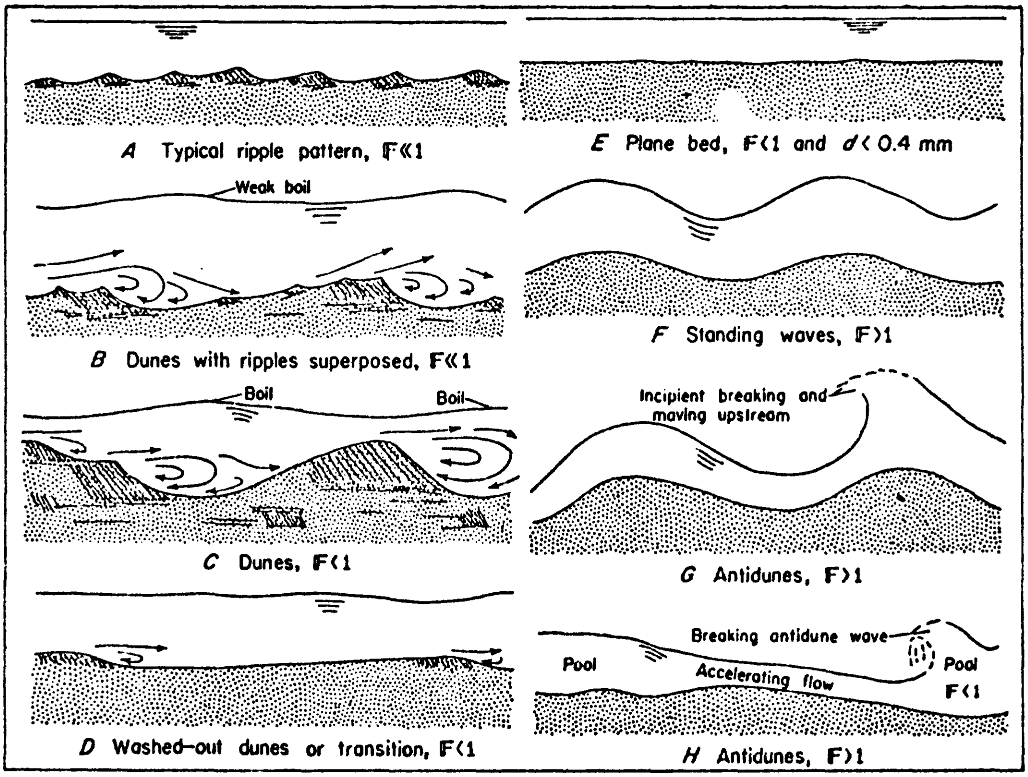


FIGURE 16.--Idealized diagram of bed and surface configurations in alluvial channels associated with various regimes of flow. (After Simons and Richardson, 1962.)

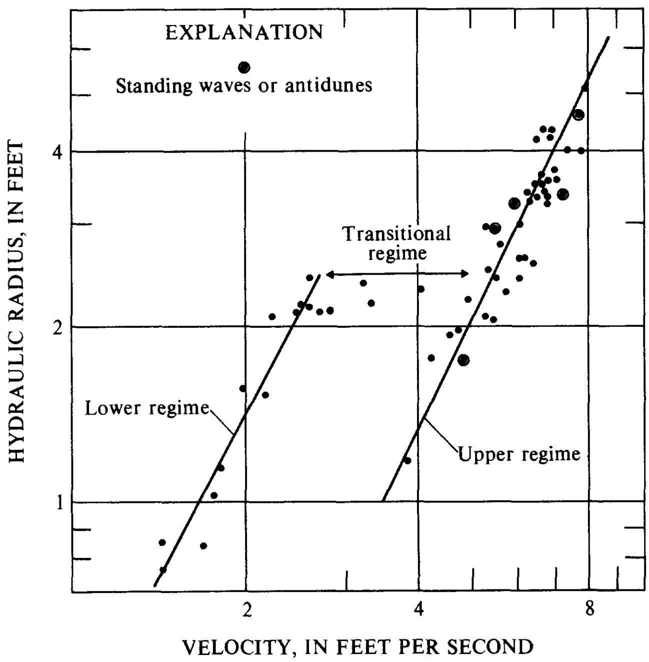


FIGURE 17.--Relation of velocity to hydraulic radius, Rio Grande near Bernalillo, N. Mex. (After Dawdy, 1961.)

In an effort to eliminate, or at least minimize, the effect of streambed changes on the stage-discharge relation, control structures are often placed in the stream. Ideally, these controls should establish a unique stage-discharge relation. For the most part the control structures tested have been those that have been used successfully in stable channels--various types of Venturi flumes and weirs of various cross-sectional shapes, some of which had upstream faces of mild slope to pass sediment and debris over the crest. Few of these have operated successfully in alluvial streams carrying heavy loads of sediment. Almost invariably the pool behind the control became filled with sediment and often the entire structure was buried in the sediment, as mentioned earlier. Costly monolithic control structures designed for use in experimental desert watersheds and for special project investigations have been described by Osborn and others (1963) and by Harris and Richardson (1964). However, it is not economically feasible to construct such controls for routine stream-gaging operations.

The Geological Survey is actively engaged in the design and testing of control structures that are both economical and effective. As yet (1969) there is no signal success to report, but F. A. Kilpatrick of the Geological Survey states (oral commun., 1969) that he is encouraged by the operation of a dual-weir control he has installed on a few streams. The control consists of two parallel rows of sheet piling, spaced about 10 feet apart, installed across the stream (fig. 18). The space between the rows is filled with crushed rock. The downstream weir acts as the control for the gage. The upstream weir acts as a baffle to produce overfall of the water, whose resulting turbulence tends to maintain sediment in suspension as the water moves over the downstream or lower weir. Optimum dimensions of the dual control have not been determined as yet; tentative dimensions are shown in figure 18. The backwater produced by the combination of weirs must be evaluated carefully with regard to channel conditions. The upstream weir crest must be high enough to produce the desired turbulence between weirs, without creating excessive backwater. The downstream weir crest must be low enough to become submerged at moderate flows, or excessive scour will occur at downstream bed and banks.

In an effort to bypass the difficulties connected with determining discharge by means of a stage-discharge relation, study is also being made of the feasibility of eliminating the stage-discharge relation and substituting an automated system of making either continuous or periodic dye-dilution determinations of discharge. Results of the measurements would be automatically recorded. However, no such system is operational at this time (1969).

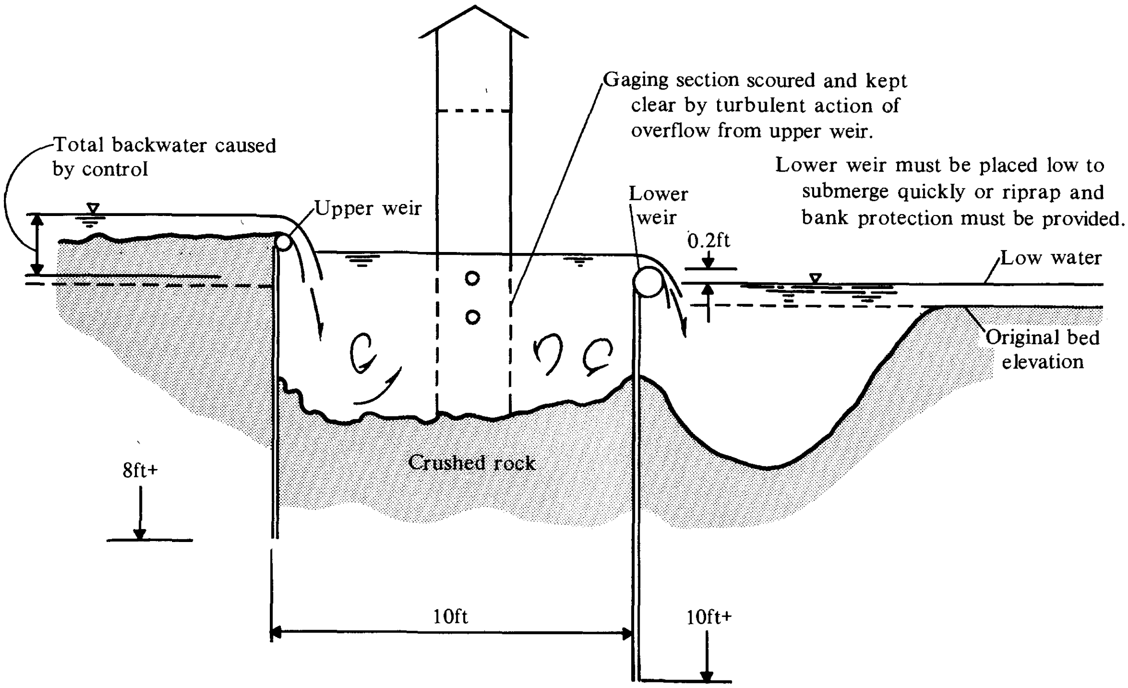


FIGURE 18.--Sectional view through centerline of dual-weir control.

## Determination of Discharge During Pulsating Flow

If the hydraulic characteristics of the channel are such that pulsating flow occurs, the hydrologist is in difficulty because, as stated on page 53, conventional methods cannot be used to determine discharge if the overriding wave carries an appreciable part of the total flow. Furthermore, without visual or photographic evidence the hydrologist has no way of knowing that pulsating flow has occurred. Conventional water-stage recorders of either the float or pressure-sensing type do not react swiftly enough to record the rapidly fluctuating stage. In addition, the state of our knowledge is such that we cannot yet predict whether or not pulsating flow will occur in a particular steep channel. If the hydrologist arrives at the scene after a runoff event, he may find a debris line along each bank marking the crest elevation of the flow, but he has no way of knowing whether the channel was completely filled to the debris line or whether only wave crests reached that height. If the hydrologist is on the scene during the runoff event, he cannot measure the discharge by conventional or optical current meter because depths and velocities change too rapidly.

Thompson (1968) reports that satisfactory results were obtained by the Geological Survey in a preliminary test to devise a method of computing discharge during pulsating flow. The test was made in Santa Anita Wash, Calif., during a period of controlled reservoir releases. Visual observations of staff gages were used to obtain depths and wave dimensions, and stopwatches were used to time the travel of wave crests through a 100-foot reach of channel. Observations of this type were made at three sites on the channel. Thompson made separate computations of the discharge in the overriding wave and in the shallow-depth (overrun) part of the flow and combined them to obtain total discharge. He describes three types of fairly elaborate instrumentation, any of which might be used unattended to obtain the necessary field data for computing discharge during pulsating flow. None have been tested as yet (1969).

## Determination of Peak Discharge by Indirect Methods

During floods it is usually impossible or impractical to measure peak discharges when they occur because of conditions beyond the control of the hydrologist. He may be unable to reach the measuring site before the arrival of the flood peak; the peak may be so sharp that a satisfactory discharge measurement cannot be made even if the hydrologist should be present when the peak arrives; or the structure from which discharge measurements are normally made may have been destroyed by the floodwater. Consequently, many peak discharges must be determined after the passage of the flood from debris lines left on the streambanks by the flood. This requires the use of such indirect procedures as the slope-area, contracted opening, flow-over-dam, and flow-through culvert methods. Descriptions of the techniques used in the various methods abound in the literature, and recently the Geological Survey prepared five manuals--Benson and Dalrymple (1967), Dalrymple and Benson (1967), Hulsing (1967), Bodhaine (1968), and Matthai (1967)--that describe the latest procedures. The reader is reminded that the methods are not applicable to pulsating flow.

Because the methods of determining peak discharge are, in general, standardized and familiar to the profession, they will not be described here, except for a discussion of some relatively new procedures that relate to the selection of the Manning roughness coefficient, $n$, for alluvial channels with unstable beds. The three regimes of flow--lower, transitional, and upper--were previously described. Figure 19 provides a means of determining the flow regime for various sizes of bed material and for various values of stream power. A discussion of stream power is beyond the scope of this report. Suffice it to say that stream power equals $62RSV$, where $R$ is hydraulic radius, $S$ is energy gradient, and $V$ is velocity. For flow in the lower regime, $n$ values will range from 0.021 to 0.035, in the transitional regime they will range from 0.020 to 0.028, and in the upper regime they will range from 0.012 to 0.026. In the lower regime, and to a lesser degree in the transitional regime, form roughness has an important influence on the value of $n$. In the upper regime the value of $n$ depends primarily on the size of bed material, as indicated in table 6. During major floods the flow will invariably be of the upper regime type.

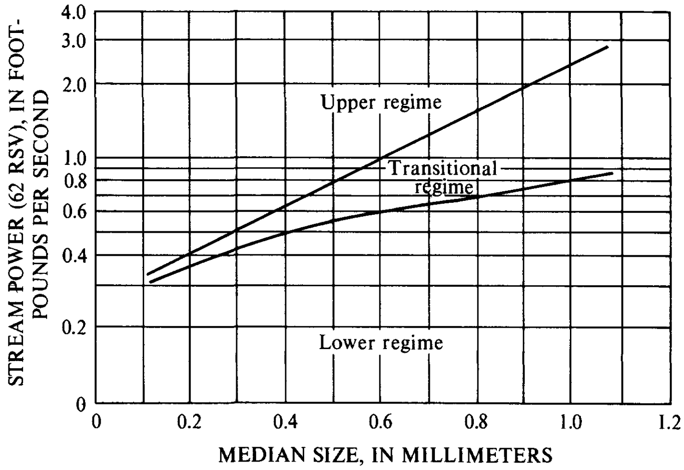


FIGURE 19.--Relation of stream power and median grain size to form of bed roughness.

TABLE 6.--*Relation of the Manning roughness coefficient* n *to median grain size in upper regime flow*

| Median grain size (mm) | Manning roughness coefficient *n* |
|---|---|
| 0.2 | 0.012 |
| .3 | .017 |
| .4 | .020 |
| .5 | .022 |
| .6 | .023 |
| .8 | .025 |
| 1.0 | .026 |

## Snowmelt Runoff

Snowmelt is a significant element in the hydrology of those basins whose headwater areas are sufficiently high and humid to maintain a snowpack throughout the winter months. The water equivalent of the snowpack April 1, as determined from snow surveys, is usually the principal factor in making both seasonal and short-term forecasts of snowmelt runoff in the months that follow.

Seasonal forecasts of the volume of snowmelt runoff are generally made for the period April 1-July 31, using correlation procedures. The principal independent variable in the correlation is the water equivalent of the snowpack, as observed in preceding years, and the dependent variable is the corresponding seasonal volume of snowmelt runoff. The correlation can usually be improved by the inclusion of other independent variables such as soil-moisture observations, base flow in March, and precipitation and runoff during the period October through March. Where snow surveys are also made on February 1 and March 1, the snowpack data are used to make preliminary runoff forecasts that will be updated in April. These early forecasts are helpful in preliminary planning for water use during the following summer months. The seasonal forecasts made April 1 are often updated in May on the basis of April precipitation, April runoff, and, where available, measurements of snowpack water equivalent on May 1. Seasonal snowmelt forecasting is described by Hannaford (1956) and by Kohler (1957). The report by Kohler contains an annotated bibliography on the subject. The procedures used in making seasonal forecasts may also be used for computing seasonal snowmelt runoff for years when runoff was not gaged.

Short-term forecasts of snowmelt discharge are made on the basis of the areal extent of the snowpack and the meteorological elements that influence snowmelt. The forecasts cannot be made for more than a few days in advance because it is impractical to predict meteorological elements any farther in advance. The meteorological elements include precipitation, incident solar radiation, cloud cover, air temperature, dew-point temperature, and windspeed. All these elements are observed or predicted for few basins, but synthetic values of the missing elements can often be derived. The snowmelt formulas are based on physical laws of heat exchange and in them are incorporated constants that reflect the effect of such environmental influences as forest cover and basin exposure. The formulas are used to compute the daily magnitude of the various components of snowmelt; namely, those associated with shortwave and longwave radiation, convection, condensation, and rain. These melt components are then totaled and routed to the gaging station under study. The procedures are described in two reports by the U.S. Army Corps of Engineers (1956, 1960) and in a paper by Rantz (1964). Best results are obtained when the ablation of the April snowpack is monitored by aerial photography and by remote radio-reporting isotope snow gages or snow-pressure pillows (Ord, 1968). However, in the absence of such instrumentation, the ablation can be estimated from a daily budget of computed snowmelt (Rantz, 1964). The procedures used in making daily forecasts may also be used for computing daily snowmelt runoff for years when runoff was not gaged.

## Flood Frequency

The study of flood frequency is an integral part of any program concerned with basic hydrologic data. Four comprehensive flood-frequency reports for Southwestern United States have been published by the Geological Survey. One prepared by Benson (1964) covered western Gulf of Mexico basins and included southwestern Louisiana, most of Texas and New Mexico, and a small area in southern Colorado. Benson separated his basins into two categories--those whose flood peaks result from rainstorms and those whose flood peaks result from snowmelt. In his analysis peak discharges of various recurrence intervals were correlated with physiographic and climatic parameters. Peak discharges within the rainstorm-flood area were found to be significantly related to seven factors: Drainage-area size, 24-hour rainfall for a 10-year recurrence interval, main-channel slope, basin length, surface area of lakes and ponds, the ratio of runoff to rainfall during the months of annual peak discharge, and the annual number of thunderstorm days. Peak discharges within the snowmelt-flood area were found to be significantly related to six factors: Drainage-area size, main-channel slope, surface area of lakes and ponds, altitude, mean annual precipitation, and the annual number of thunderstorm days.

Benson concludes that:

> "After use of the significant variables, most of the variability remaining is random and is believed attributable to the great variability of storm occurrence in this region. However, some of the residual variations in peak discharge appear to show some local patterns that indicate the influence of important factors not considered."

The other three regional reports by the Geological Survey were prepared by: Butler and others (1966) for the Great Basin, Patterson and Somers (1966) for the Colorado River basin, and Patterson (1965) for the western Gulf of Mexico basins (identical region covered by Benson, 1964). The index-flood method of analysis (Dalrymple, 1960) was used in all three of these studies. The index-flood method requires the preparation of two curves. One curve relates the mean annual flood to contributing drainage area. Where applicable, mean altitude of the basin is used as a third parameter in this relation. The second curve relates peak discharge, expressed as a ratio of the mean annual flood, to recurrence interval. A homogeneity test is used to delineate the boundaries of homogeneous subregions and individual curves are drawn for each subregion. By combining data from the two curves, a flood-frequency curve can be constructed for any basin whose contributing drainage area and mean altitude are known.

The published flood-frequency reports are generally fairly satisfactory, but they all have one shortcoming--flood-frequency relations had not been developed for subregions that are truly arid because of the lack of long-term flood records for the small streams that occupy most of the arid areas. However, since the time when those reports were prepared, numerous small-stream gaging stations have been in operation long enough in most of those arid areas to provide records of sufficient length to be helpful in an analysis of flood frequency. At present (1969) all district offices of the Water Resources Division of the Geological Survey are updating their regional flood-frequency studies, using these additional data, and by the time this paper is released, regional studies covering all areas in the entire Nation will probably have been completed. Results of the studies probably will either have been published or will be available for consultation in the appropriate district office of each State. All analyses will have been made using the multiple-regression method (Benson, 1964). At those recently established small-stream stations where both stream stage and precipitation were continuously recorded, relatively long synthetic peak-discharge records will usually have been generated using the Dawdy-Lichty flood-hydrograph model (Lichty and others, 1968).

Several empirical methods of deriving flood-frequency relations have also been used, the most popular of which is the SCS method (U.S. Soil Conservation Service, 1957). That method entails the use of generalized precipitation-frequency and rainfall-runoff relations. Generalized precipitation-frequency relations are available in the form of maps in Technical Paper 40, a publication of the U.S. Weather Bureau (1961). The relation of volume of direct runoff to storm rainfall is given in the Soil Conservation Service report (1957) for various soil groups, vegetative cover types, and antecedent conditions that relate to infiltration capacity. In a final step of the method, the volume of direct runoff corresponding to a storm rainfall of given frequency is distributed in time in accordance with a standard triangular unit-hydrograph.

The principal shortcoming of the various empirical approaches to the problem of defining flood-frequency relations lies in the fact that no unique relation exists between precipitation frequency and flood frequency. For example, it does not follow that the 50-year storm of critical duration will cause a 50-year flood, because the antecedent soil-moisture condition is a governing factor with regard to the amount of storm precipitation that will be available for direct runoff. To illustrate this, the storm of January 21-23, 1943, in southern California was the most intense of record in much of the region from a standpoint of flood-producing potential. However, flooding was not severe in 1943 because the rains fell on dry ground. Unfortunately, there is no way to associate antecedent conditions with the rainfall frequencies given in Weather Bureau Technical Paper 40. There are other shortcomings to the empirical methods, such as the use of a hydrograph of standard shape and the use of a standard time-distribution of the rainfall within the selected period of storm duration. However, it is not to be inferred that the use of rainfall-runoff relations in flood-frequency analysis is discouraged. On the contrary, it is recommended that in areas where streamflow records cover relatively few years but where long-term recording-precipitation records are available, rainfall-runoff relations be used to synthesize a long-term record of peak discharges. However, instead of using empirical methods, modeling techniques such as those described by Lichty and others (1968) or by Crawford and Linsley (1966) should be used.

## Reservoirs and Lakes

The subject of reservoirs and lakes (including stock ponds) has already been mentioned in the discussion of evaporation from open-water surfaces. This section of the report tells of the effect of the construction of a reservoir on the hydrologic budget of a basin. It is apparent, of course, that the addition of a reservoir to a stream system will increase the evaporation loss from the basin because of the increase in area of exposed water surface. However, the increased evaporation loss will be offset, in part, by the fact that all rainfall on the lake surface becomes a part of the water yield whereas previously an appreciable part of this rainfall was retained as soil moisture for subsequent evapotranspiration. There may be a further gain in yield, too, as a result of riparian and phreatophytic vegetation being inundated by the reservoir, thereby eliminating a source of evapotranspiration loss. On the other hand, the reservoir may create conditions conducive to an increased stand of phreatophytic growth and thereby create a source of increased evapotranspiration loss.

The direction and rate of ground-water movement in the vicinity of the reservoir will change. If the stream was originally a gaining one--that is, one that received effluent ground water--the rise in the water table in the vicinity of the lake will flatten the upstream water-table gradient, thereby decreasing seepage into the reservoir, at least until conditions approaching equilibrium are established. Thereafter the direction and rate of ground-water movements, as well as the quantity of bank storage, will probably vary in response to fluctuations of water level in the reservoir. In addition to these effects there will be seepage from the reservoir in the downstream direction. If the stream was originally a losing one, the only change brought about by the reservoir will be a probable increase in seepage loss in the reservoir area.

Kennon (1966), reporting on the hydrologic effects of small reservoirs on ephemeral streams in western Oklahoma, found that reservoir seepage appeared as surface flow in the channels downstream from the reservoirs, thereby changing ephemeral streams to perennial streams. The perennial flow accelerated the growth of riparian vegetation, which led to a change in channel geometry and to a reduction in channel capacity. Gilbert and Sauer (1969, p. 38) show that seepage losses in ephemeral channels downstream from reservoirs may be smaller under conditions of reservoir regulation than under natural conditions. Because flood discharges are smaller under regulated conditions, they occupy a smaller cross-sectional area of the channel, and consequently a smaller area of river bank is exposed to seepage.

No discussion of reservoir storage would be complete without further mention of the effect of bank storage. Bank storage can be an appreciable item in the water budget, and if so, failure to consider it will result in a budget that is out of balance in one direction when the reservoir stage is rising, and out of balance in the opposite direction when the stage is falling. Langbein (1960), in a discussion of bank storage at Lake Mead on the Colorado River, demonstrates appropriate computational procedures.

## Sediment Movement and Deposition

Knowledge concerning sediment movement and deposition is needed for studies pertaining to: Siltation of reservoirs, channel degradation and aggradation, scour and fill around manmade encroachments on a channel, design of channel diversions, design of stable channels, effect of land and water conservation or of urbanization on sediment yields and erosion, and the areal distribution of sediment sources. The discussion that follows has been abstracted almost verbatim from a report by the Hydrology Subcommittee of the Pacific Southwest Inter-Agency Committee (1966). Details of the methods used in the measurements and analysis of sediment loads will not be discussed. For such information the reader is referred to two reports, one by the Inter-Agency Committee on Water Resources (1963) and the other by Colby (1963).

*Suspended-sediment discharge.*--Determinations of suspended-sediment discharge are most meaningful when they cover the entire range of discharge of a stream. It is readily apparent that a few field determinations made at random times during the year may lead to erroneous conclusions concerning the annual transport of sediment. However, the same effect is possible even if numerous field determinations have been made. If most of the determinations have been made in a particular season, or if most of them have been made during a year of either drought or excessive runoff, they will yield erroneous results when applied to a long-term flow-duration curve. Flashy streams in remote areas are the most difficult to sample adequately because peak flows, which carry the bulk of the suspended sediment, occur with such suddenness that they often pass the sediment stations before the observers are aware of their occurrence. Automatic samplers, which have been developed in recent years, are only a partial answer to the problem of sampling flashy streams because the samplers are subject to damage and clogging in streams that carry large quantities of floating debris.

There are other difficulties, all connected with variations in the transport rate of suspended sediment. At a given discharge, the suspended-sediment concentration on a rising stage commonly differs significantly from that on a falling stage. At this same discharge, but at different times of the year, concentrations may differ because temperature affects the viscosity of the water, which in turn affects the transport rate. Other factors that affect the transport rate of suspended sediment are: (1) Changes in the source of sediment as a flood rises and wanes; (2) the time between stream rises, which may or may not be sufficiently long for bed and banks to become stabilized; and (3) changes in the flow regimen caused by diversions and storage of streamflow, changes in land use, or other cultural changes. It is apparent, therefore, that meaningful suspended-sediment records will be difficult to obtain for desert streams, many of which are not readily accessible.

Suspended-sediment transport is often estimated from past records of streamflow. When this is done, sediment-transport rates should be estimated from daily or momentary data by successive increments of time, with appropriate allowance for increasing and receding flows. Less preferable is the use of a flow-duration curve in which the identity of distinct sequences of flow is lost. Sediment discharges should not be estimated from monthly streamflow data, because the use of monthly records tends to suppress the effect of extreme high flows which transport a disproportionately large part of the total suspended sediment. Accordingly, sediment discharges estimated from monthly streamflow data are likely to be too small.

If annual sediment discharges at a site are to be computed for a period prior to the collection of records, the appropriate relation of daily sediment transport to discharge must be applied to a synthetic streamflow record for the ungaged period of years. Two methods are available. In the first method, daily discharges for the ungaged period are computed by use of a rainfall-runoff model, such as that of Crawford and Linsley (1966). The model is calibrated from records for the period of operation of the gage. In the second, and more commonly used method, we forego some of the refinements discussed in the preceding paragraphs and apply an average sediment-transport relation to discharges obtained from a synthetic flow-duration curve of daily discharge. This synthetic curve is derived by first correlating concurrent known discharges at the study site with those at a long-term gaging station, and then applying the resulting discharge relation to the flow-duration curve for the long-term station.

Particle-size analyses of the sediment--that is, percentage distribution among the several standard ranges of particle "diameter"--indicate the volume-weight ratio of the sediment as it will be deposited in reservoirs and, consequently, the quantity of sediment-storage space that must be reserved. If particle-size analyses are available only for infrequent samples, each analysis should be weighted in proportion to concurrent discharge to derive a reasonable particle-size distribution in yearly sediment discharge.

*Bedload.*--Measurements of bedload are seldom made because satisfactory bedload samplers have not yet been developed. However, on some sandy streams a channel constriction will accelerate the velocity and cause sufficient turbulence to put the moving bedload into suspension, so that the entire sediment load can be measured with a suspended sediment sampler. A similar turbulence effect can sometimes be achieved by the construction of a special turbulence weir or turbulence flume (Hubbel and Matejka, 1959).

In lieu of measurements, bedload commonly is estimated or computed as a percentage of total load or of suspended load. Error in such estimates may be substantial but it can be minimized by field observation of streambed and overbank deposits, which indicate the size range of materials available to the stream. It should be borne in mind that what constitutes bedload at one cross section may be suspended load at another section where stream velocity is greater.

Computations of bedload transport, and of potential depth of channel scour in noncohesive materials, depend on particle-size distribution. To be valid for such a computation, a particle-size analysis should involve a sample sufficiently large so that all channel-bed materials are represented. Thus, the mass of a sufficient sample might range from a few pounds to 100 pounds or more, depending on the range of particle sizes. Also, if texture of the streambed materials changes with depth, samples from successive depths should be analyzed.

Numerous methods and formulas for computing bedload have been proposed, but all are subject to considerable error. Even collectively they do not cover all field situations. Because these bedload formulas have been derived from studies in flumes, they apply best to stream reaches that are straight in alinement and uniform in cross section; nonuniform reaches may range considerably in capacity to transport bedload. For wide, shallow streams with sand beds the modified Einstein method (Colby and Hembree, 1955) is usually considered the most reliable procedure for computing bedload capacity. Application of the method requires data on stream velocity and suspended-load concentration. In the absence of information on suspended-load concentration, the transport of sand in a sand-bed stream may be computed by a less refined method. Colby (1964) has found that the transport of sand correlates fairly well with mean velocity. Two other formulas that are commonly used for computing bedload are the Schoklitsch (Shulits, 1935) and Meyer-Peter and Muller formulas (U.S. Bureau of Reclamation, 1960). A comprehensive discussion of the various bedload formulas is presented in a recent report by Shulits and Hill (1968).

The above discussion of bedload is not applicable to viscous mudflows that act as non-Newtonian fluids. Mudflows were discussed on page 53.

*Deposits in reservoirs.*--Volumes of sediments in reservoirs are determined by one of two methods--direct measurement of sediment thickness by comparison of reservoir-bottom profiles along permanent ranges, or recontouring of the reservoir basin and comparison with the original or other antecedent topography. Because the deposits may vary widely in density from one reservoir to another owing to differences in compaction, representative undisturbed samples are required to obtain in-place density.

Distinguishing sediment deposits in reservoirs from the underlying soil may be difficult in Western valleys where, under an arid or semiarid climate, typical soil profiles are indistinct or nonexistent. In those circumstances, examination of the alluvial soils upstream and downstream from the reservoir may be helpful. Commonly the soil particles naturally range rather widely in size and are arranged heterogeneously, whereas reservoir deposits, except those of deltas, are characteristically of uniform texture.

For determination of in-place density (volume weight or specific weight) samples of undisturbed reservoir sediment may usually be obtained without great difficulty. For accurate results it is necessary that a representative array of sampling points be selected. When samples cannot be obtained, and the in-place density must be estimated, experience and good judgment become critical. For a sediment composed of sand-size particles, most authorities agree on a narrow range of values whose mean is about 85 pounds per cubic foot. However, if the particles are principally of clay sizes, the density may range widely.

If data on volumes of reservoir sediment are to be compared or combined, appropriate adjustments must be made for differences in compaction and differences between dry density and wet density. In this connection, volume per unit area of watershed is much greater for reservoirs that hold water perennially than for reservoirs that are intermittently drained. Proper sampling will eliminate much of the complexity in analyzing the data.

*Deposits on fans and flood plains.*--In the Southwest most measurable deposition of sediment, aside from that in reservoirs and deltas, occurs on fans rather than on flood plains. In other words, along a river valley deposition is local and discontinuous rather than general. Recent sediment sometimes can be discerned and measured with reasonable accuracy. Ordinarily, however, it is difficult to distinguish between recent and old deposits because, as has been noted, typical soil profiles are indistinct or nonexistent.

*General aspects of sediment yield.*--Sediment yield is conventionally expressed as a yearly average, in terms of acre-feet, or in terms of tons per square mile or per acre of drainage area. In the West, however, much of the sediment is derived from valley trenches, and it is exceedingly difficult to determine the actual source areas of sediment within a gaged basin and their relative rates of yield. Also, the transposition of data from one basin to another may be invalid unless the data are adjusted in accordance with the relative extent of each particular type of erosion. Obviously, a computation of sediment yield per square mile or per acre requires that size of the contributing area be known with reasonable accuracy, especially if the area is small.

*Sediment yield as related to deposits in reservoirs.*--Volume and weight of sediment in reservoirs may not represent total or long-term average yield from the drainage area. The reasons for this include the following: (1) The period spanned by data may not be a period of average climatic conditions; that uncertainty applies particularly in the Southwest, where rainfall varies greatly from place to place, from year to year, and from decade to decade. (2) Not uncommonly, data on sediment yield are from reservoirs of low trap efficiency; there is no reliable substitute for actual measurements of suspended sediment for computing the quantity of sediment passing over spillways. (3) Some data are for reservoirs with ungated outlets; as in the case of sediment passing over spillways, sediment passing through such outlets cannot be calculated satisfactorily. (4) Available data commonly fail to cover deposits above spillway level--that is, upstream from the reservoir; in many Western streams coarse sediment is a substantial part of the total yield and much of it may be deposited outside of reservoirs.

*Conclusions.*--The present status of sediment measurements and analysis suggests caution in interpreting small differences between observation periods or places, in interpreting differences between components of sediment discharge, and in inferring sedimentary processes from sediment-discharge measurements.

## Chemical, Thermal, and Biological Quality

Water samples are collected and analyzed to determine the characteristics of a body or mass of water. The sample is usually only an infinitesimal part of the total volume and is therefore representative of the total mass only to the degree that uniformity of chemical composition exists within the total mass. In its natural state, surface water is subjected to forces that promote mixing and homogeneity. The fact that such tendencies exist, however, is not sufficient cause for assuming that a body of water is so well mixed that no attention to sampling technique is required. For example, a station to sample the chemical quality of a stream usually should not be established immediately downstream from a tributary or an outfall for waste because of the possibility of streamline segregation of unlike waters. Most lakes and reservoirs provide excellent examples of the nonuniformity of composition that may be found in a mass of surface water.

The chemical composition of water is the resultant of the geologic, hydrologic, biologic, and cultural environment and varies from time to time, as well as from place to place. The variation of surface-water quality with time can be quite pronounced, as when base flow is augmented by large quantities of surface runoff, or when a major part of the streamflow becomes regulated by storage reservoirs. Such changes in the cultural development as urbanization or increased irrigation activity likewise produce pronounced changes in water quality.

The methods and frequency of sampling and the precision of the analysis of water to determine its quality characteristics depend primarily on the use--present or future--to which the water is to be put. For example, a few infrequent samples may suffice where large-volume agricultural uses are involved, whereas continuous or composite sampling may be necessary where waste discharges are involved. Partial or approximate analysis of water samples may be satisfactory in some cases; in others, a complete analysis is needed, and because some of the constituents dissolved in the water may change after a sample is collected, fixing or field determination of these constituents may be necessary. The choice of field or laboratory procedures to be used is also governed by the economic value of the study. For a detailed discussion of methods for the collection and analysis of water samples, the reader is referred to a report by Rainwater and Thatcher (1960).

From the foregoing discussion of the variability of the chemical composition of water it is apparent that each surface-water sample should be identified not only with regard to location, date, and possibly the hour, of sampling, but also with regard to all relevant aspects of the hydrologic situation. Networks of sampling stations should be designed from a consideration of either the geographic variations in water quality or the particular quality problem under study. For example, some deleterious constituent in the water, such as an organic pollutant, may form or be introduced into the stream and then, within a relatively short reach of channel, be dissipated by reaction with another dissolved or suspended substance or with a streambed material. A situation of this kind may escape detection if sampling stations are widely spaced. Where continuous monitoring of specific constituents or properties is done, the monitoring should be supplemented, for a limited period, by frequent sampling for laboratory analysis. The results of the analyses should be used in correlations to establish relations between monitored and nonmonitored constituents or properties.

All the requirements for adequate chemical sampling may not be obvious in the early stages of a water-quality study. Conversely, water-quality records without comprehensive information as to location, time, and circumstances of the sampling may not be explicitly interpretable. However, even with a clear understanding of the environmental background of the data that have been collected, the hydrologist must use caution in transposing or extrapolating his information. Guidelines for the interpretation of chemical records and data have been established by Hem (1959).

Thermal quality, or water temperature, is a highly significant property of surface water. Because of its effect on the viscosity of water, temperature affects the rate at which streamflow may seep into the streambed and is an important factor in the transport of sediment. Water temperature also has an effect on the bed configuration of sand-bed channels. Water temperatures that bear little relation to the air temperature may also be an indicator of the source of the water in a stream. In developed areas temperature is important in determining the value of water for cooling purposes and for the irrigation of specialized crops, particularly those whose growth may be impaired by cold water released from reservoirs that store snowmelt runoff. The collection of water-temperature data is relatively simple and many recording temperature instruments, that are both dependable and inexpensive, are commercially available.

The study of biological quality requires sampling that differs in several respects from chemical sampling. For example, because there is no single universal indicator species, a rather long series of biological samples is required. Furthermore, the analyst must detect and interpret either extreme or widely variable conditions because water quality must be keyed to critical or threshold levels where aquatic life is concerned. The determination of biological quality is usually considered as being beyond the scope of a basic-data program and will receive no further treatment in this paper.

## Data for Studies of Fluvial Morphology

Studies of fluvial morphology are beyond the scope of a basic-data program. However, because data that are useful for such studies can be collected with relatively little additional effort, it is recommended that a few appropriate measurements be made in connection with the routine operation of a stream-gaging station. The authors suggest that a reach of channel, about 1,000 feet long, be selected for observation. This reach should include the gage site and the site where discharge measurements are made. The reach should be mapped to show the thalweg, the in-channel area, bars, berms, and the flood plain. Four or five cross sections should be surveyed--preferably, at the two ends of the reach, at the measuring section, at the gage pool, and at the first riffle, if any, downstream from the gage. The ends of the cross sections should be permanently referenced with iron rods or pipes driven into the ground. Cross sections should be resurveyed after each major runoff event. Roughness coefficients (Manning's $n$) should be selected for the reach, and stereophotographs that show sufficient detail of bed and banks for evaluation of these coefficients should be obtained. Where practicable, aerial photographs of the reach, in color, should also be obtained.

The bed material should be sampled at each cross section and, in addition, the magnitude of scour and fill at each cross section should be determined by the use of scour chains (Leopold, Emmett, and Myrick, 1966). Scour chains are chains buried vertically in the streambed with the top link at or slightly above the bed surface. Three chains should be used in each cross section--one at each quarter point of the channel width. When resurveying a cross section after a runoff event, the bed is dug until the chain is exposed. If scour has occurred, a part of the chain will be lying horizontally at some depth below the channel bed. The difference between the previous streambed elevation and the elevation of the horizontal chain is the depth of scour. If no scour has occurred the depth of fill is the increase in bed elevation. The chains provide a measure of maximum scour and fill at a point without respect to time; all points in a cross section do not necessarily attain maximum scour or fill simultaneously, and one side of the stream may actually be filling while the other side is scouring. Nevertheless, the measurements are a useful guide to the limiting values of scour and fill in the stream under the flow conditions experienced.

## GROUND WATER

### Description of Ground-Water Basins

The physical setting of arid-land basins in Southwestern United States has already been described briefly, but that description will be elaborated upon at this point. Meinzer (1923, p. 314) in discussing the region stated,

> "The principal source of water supply in this arid province is the alluvial sand and gravel of the valley fill underlying the numerous intermontane valleys that characterize the region.... In the elevated marginal parts of the valleys the water table may be very far below the surface or ground water may be absent; in the lowest parts, underlain by clayey and alkaline beds, ground water may be meager in quantity and poor in quality; at intermediate levels, however, large supplies of good water are generally found."

To expand upon Meinzer's description, the mountains enclosing or adjacent to the valley basins are usually the only areas that receive sufficient precipitation to produce either direct recharge or runoff that eventually recharges the ground-water body. The recharge water moves downward and laterally toward the basins. Some of the consolidated rocks store and transmit very minor amounts of water, but secondary openings in volcanic, carbonate, and to some extent in other consolidated rocks, may locally be quite productive and store much ground water. Usually, however, it is only the loose, coarse alluvial deposits underlying the basins that have the high permeability needed to yield large quantities of water to wells. Such deposits are generally found locally; much of the valley fill is of low permeability, being composed of either fine-grained or poorly sorted coarse deposits.

The yields to wells, therefore, may vary widely from place to place, but on the whole the valley-fill reservoir usually stores much water. The valley fill may be very thin in some basins, but in others it may be hundreds, or even thousands, of feet thick below the basin floor. The valley fill is commonly saturated and graded to the floor of the basin, but in some basins the fill may be dry for several hundred feet below the basin floor.

The water stored in ground-water reservoirs is usually many times, commonly many hundreds of times, greater than the average annual natural inflow to those reservoirs, and except in areas where the water table is close to land surface, there is little loss by evapotranspiration. However, this supply, despite its magnitude, is not limitless, as has been demonstrated in many areas of heavy pumpage. The continued demand for more water from developed basins, as well as new development in other basins, has emphasized the need for increasing the refinement of our estimates of the dimensions and capacities of ground-water reservoirs. It is also necessary that we define more precisely the entire hydrologic system of which the reservoir is part.

## Basic Elements of a Ground-Water Study

Methods used in the quantitative determination of elements of a ground-water system are, of necessity, less direct than those used in surface-water investigations. However, several of the ground-water elements vary slowly with respect to time, and quantitative determinations of those elements represent average values that are relatively stable over long periods of time. Examples of such elements are the quantity of ground water in storage, the rate of movement within the ground-water reservoir, and the variation in chemical quality of water within the ground-water system. More variable are such elements as total natural recharge and total natural discharge for periods of a year or less, and consequently those two elements are more difficult to determine reliably.

The following sections describe some of the principal methods used in evaluating the basic elements of a ground-water system. Those elements are treated under the broad headings of storage, transmissivity, recharge, discharge, and chemical quality.

### Storage

The amount of available water stored in an alluvial basin can be estimated by (1) delineating the horizontal and vertical limits of the reservoir, (2) computing the gross volume of lithologic units within those limits, and (3) determining the drainable pore space, or specific yield, of each lithologic unit. Because the area of the ground-water basin is commonly many tens or hundreds of square miles and the thickness of sediments is many hundreds of feet, an absolute determination of storage is rarely possible, and in any event, would be prohibitive in cost. Invariably, then, an approximation of the amount of storage is made. The degree of reliability of the approximation is generally assumed to be related to the time and effort used in collecting, analyzing, and interpreting the data. The validity of that assumption cannot always be demonstrated, but the thorough analysis of substantial quantities of data tends to increase the confidence of others in the estimate. As would be expected, the greatest effort in determining ground-water storage is put forth in areas where water has high value.

## Boundaries of the ground-water reservoir

The valley-fill reservoir is usually enclosed by relatively impermeable bedrock. The surface trace of the bedrock and valley-fill boundary, commonly called "bedrock-alluvial" contact, ordinarily can be identified rather readily. However, the subsurface interface between those two geologic units usually can be defined only generally even with substantial investigation. Projection of this interface to define the "bottom" of the reservoir requires a thorough understanding of the geologic framework, structure, and history. If the delineation is based entirely on surface mapping, subsurface projections may be appreciably in error. A few direct point controls from well data can aid significantly, but extrapolation from a few points still can result in large errors. If the bedrock and valley fill have sufficiently strong physical differences, one or more of the geophysical methods may provide a reliable definition of the subsurface form of the bedrock-valley fill interface. In developed areas, drillers' logs may provide helpful information. In any event, any firm delineation of the subsurface interface is almost always an item of high cost.

Most of the interest in ground-water storage is directly concerned with perhaps as much as the upper 1,000 feet of saturated deposits, but the economics of ground water use usually restricts the determination of storage to no more than the upper 500 feet. Furthermore, the effective depth of the ground-water reservoir may be limited by the presence of tightly packed sediments of low permeability or the presence of underlying saline water. The lateral limits of the ground-water basins are generally delineated by convenient controls such as (1) the surface contact of the bedrock and valley fill, (2) faults, or (3) hydraulic boundaries such as ground-water divides or recharging streams; or the limits selected may be such arbitrary, but practical, ones as political boundaries or land lines.

The use of arbitrary lateral boundaries seems absurd if one mistakenly thinks of the hydraulic response of a ground-water basin as being comparable to the rapid response characteristic of surface-water reservoirs. Thomasson and others (1960, p. 146) explained the difference in the characteristics of the two types of reservoirs in the following statement:

> "Ground-water basins or areas are frequently referred to as ground-water reservoirs. So far as capacity to store and give up water are concerned, the analogy with a surface reservoir is excellent with respect to storage volumes alone, but with respect to time required for the adjustments associated with changes in storage, the two reservoir systems are not at all similar. The surface reservoir adjusts almost instantaneously to changes in storage caused by excess inflow or excess outflow, whereas the ground-water reservoir reacts more like a body of highly viscous liquid such as cold molasses, in which considerable time is required for adjustment of fluid level. Even in the most permeable materials, such as uniform sand or gravel, the rate of movement of water is extremely slow compared to the rate of movement in open channel flow. In heterogenous materials, such as those underlying the Putah area (Solano County, Calif.), the rate of adjustment is even slower. Months or years may be required for the complete readjustment of an extensive ground-water basin to a major change in regimen, such as would be caused by a large increase or decrease in pumping draft from some part of the basin or by artificial replenishment to some part."

It is therefore the slowness of water movement from the more remote parts of the ground-water reservoir that permits the use of arbitrary storage boundaries. However, the fact that hydraulic connection still exists between the arbitrarily bounded storage unit and outlying parts of the ground-water reservoir should not be ignored.

## Gross volume of lithologic units within the ground-water reservoir

After the boundaries of the ground-water reservoir have been delineated, the total or gross volume of the rock material within those boundaries may be computed or estimated from the geometry of the reservoir. With the gross volume of the reservoir known, it is only necessary to determine the specific yield, or percentage of drainable pore space in the rock material, to obtain the effective storage capacity of the reservoir. However, the rock material in a reservoir is seldom so homogeneous that a single value of specific yield is everywhere applicable within the reservoir. Except for preliminary or reconnaissance studies it is therefore necessary to determine the boundaries of each lithologic unit present, so that the gross volume of each unit can be computed or estimated. Appropriate values of specific yield can then be applied to each unit.

To determine the boundaries of specific lithologic units, subsurface geologic and hydrologic conditions are examined by extension of surficial geologic data and the sequence of geologic events. The interpretation is strengthened by using well log information, such as drillers' logs, various bore holes, subsurface geophysical logs, sample logs, and cores. Correlations between these data points may be by eye, by common characteristics including physical or chemical properties, or by geophysical surveys. In the usual cost-limited ground-water investigations subsurface correlations are developed by constructing geologic sections or fence diagrams to aid in developing understanding of the subsurface conditions. The geology of the area may be further evaluated in relation to similar geologic and hydrologic environments where data are available, and the interpretation of drillers logs may be facilitated if there is similarity to log information and related data for other areas where intensive studies have been made.

The substantial lateral and vertical lithologic variations encountered in valley-fill or alluvial-basin deposits together with the varied degrees of reliable identification in drillers' logs combine to make section correlations rather difficult. However, peg models, a three-dimensional equivalent, have been used in a number of ground-water investigations in California (Worts, 1951; Poland and others, 1956; Davis and others, 1959; Thomasson and others, 1960). In the usual peg model a base map is mounted on a table top and dowels (pegs), representing wells that have been logged, are driven into holes in the table top at their proper geographic location. The table top forms an arbitrary datum plane--for example, 1000 feet below sea level--for vertical control. The different beds penetrated by a well are shown by color bands on the dowel, the width of the bands conforming to some selected vertical scale. Land surface is also marked on the dowel, whose length is cut so that about 2 inches of dowel remains above the land-surface mark. The well number is marked on that additional length.

The peg model provides an excellent means for examining the three-dimensional subsurface distribution of valley fill and is valuable to present the picture to others. Because the features shown are fixed, the peg model primarily is used for identifying subsurface geologic relations. The distribution of lithologic units then can be used to estimate volumes of the lithologic units. The volume of ground water in storage can be estimated by applying the appropriate specific-yield factors to the lithologic units and summing them by specified depth ranges and storage areas.

## Specific yield

Specific yield, which is defined as the percentage of drainable pore space in the rock material, may be determined by several methods, the most common of which are described below.

*Analysis of samples of rock material.*--The specific yield of various types of rock material has been determined by numerous investigators, who saturated samples obtained during well drilling and allowed them to drain. The volume of water that drained, expressed as a percentage of the gross volume of the sample, is the specific yield. Promeninent among such investigations were those by Eckis and Gross (1934), Piper and others (1939), Kues and Twogood (1954), and Cohen (1961). The results of those studies together with some less detailed studies, were used with some modification by Davis and others (1959) for assigning specific-yield values to six groups of material classified in the well logs as follows:

| *Material* | *Specific yield (percent)* |
|---|---|
| Gravel, sand and gravel, and related coarse gravelly deposits | 25 |
| Sand, medium- to coarse-grained, loose, well-sorted | 25 |
| Fine sand, tight sand, tight gravel, and related deposits | 10 |
| Silt, gravelly clay, sandy clay, sandstone, conglomerate, and related deposits | 5 |
| Clay and related very fine-grained deposits | 3 |
| Crystalline bedrock (fresh) | 0 |

In the absence of more specific information, average values of specific yield, such as those shown above, may be used in investigations of the reconnaissance type.

*Analysis of aquifer tests.*--The storage coefficient and transmissivity can be determined by controlled withdrawals from, or additions to, ground-water reservoirs and observation of the water-level response. (The storage coefficient for an unconfined aquifer is essentially equal to specific yield; the term "storage coefficient" is used here because this discussion deals with change in storage in both confined and unconfined aquifers.) Among the many reports of aquifer tests are those by Ferris and others (1962) and by Bentall (1963b). The methods require idealization or simplification to permit mathematical treatment. For some aquifers, conditions may be such that they respond more or less closely to the "ideal" aquifer. However, the "real" systems, particularly those in valley-fill deposits, are much more complex. Short-duration pumping tests commonly produce lower values for the storage coefficient than would result from long-term stress of valley-fill deposits. Many short-duration pumping tests of valley-fill deposits yield values of the storage coefficient in the artesian or semiartesian range of 0.001 or less, but under long-term pumping and consequent dewatering the tests generally yield values in the unconfined-aquifer range of 0.1 to 0.2.

Aquifer tests by pumping involve sampling a larger volume of aquifer than is possible by laboratory tests on core samples or in experiments with large size "tanks". However, as the volume increases the time needed to drain the aquifer sample increases and may require pumping tests of 2 to 5 years or more for the indicated storage coefficient to approach the average specific yield of valley-fill deposits.

*Analysis of long-term well-field pumpage and water-level response.*--Analysis of cumulative annual pumpage and the water-level change in an area provides a measure of response to pumping of an even larger sample of the ground-water reservoir than the single pumped-well aquifer test. Water-level changes between specified periods can be used to define the volume of drained valley fill. With pumpage or artificial withdrawal known or estimated, the average specific yield or storage coefficient can be computed. This method, used in an investigation in Avra Valley, Ariz., was described by White, Matlock, and Schwalen (1966, p. 31-36).

Similar studies involving the determination of storage coefficients were made by White and Childers (1967) in the Douglas Basin, Ariz., and by White and Hardt (1965) in the San Simon Basin, Ariz. In Avra Valley and the Douglas Basin, where the aquifers are unconfined, the storage coefficients determined had reasonable values--0.15 and 0.20, respectively. In the San Simon Basin a storage coefficient of 0.1 was indicated for the lower artesian aquifer. That value was relatively consistent for several increments of time although a satisfactory explanation for the high storage coefficient in an artesian aquifer was not possible with the data available.

## Determination of storage changes from long-term water-level fluctuations

When specific-yield data are available or can be estimated, a record of long-term water-level fluctuations may be used to compute changes in storage. Thomasson and others (1960) in their investigation of the Putah area, California used this method to compute changes in storage for two periods--one of rising water levels (1932-41) and a following period of declining water levels (1941-50). Contour lines of water-level changes in each of the two periods and their distribution within the areas of storage units provided the basis of computing average water-level change for the storage unit. The change in water level multiplied by the area of the storage unit gave the gross volume of material flooded or dewatered. The gross volume of material was then multiplied by the appropriated specific-yield value to obtain the change in storage of that unit.

A similar technique for shorter time periods was used by Clark (1915) in his investigation of the Niles cone area, California, and by Piper and others (1939) in their investigation of the Mokelumne area, Calif.

### Conclusions

The methods used for estimating storage commonly are dictated largely by the availability of data, manpower, time, and funds. All of the methods may produce reasonable results if they are used with care and consideration, but occasionally the results are difficult to reconcile with our knowledge, though limited, of the system. Ordinarily, however, if properly controlled, the analysis by well-field or total pumpage and water-level response should provide the more reliable results because it effectively samples a very large area--sometimes most of the upper few hundred feet of the total ground-water reservoir. Obviously, the method requires a long-existing development so that good records are available for 5 years or more. If control is to be significantly strengthened in undeveloped areas, the subsurface conditions must be defined by an adequate amount of test drilling and related subsurface logging and sampling, and hydraulic testing. From these data extrapolation can be made on the basis of surface geology and geophysical surveys.

## Transmissivity

Transmissivity, called transmissibility in earlier reports, is commonly determined by analysis of data obtained from aquifer tests. The subject is discussed in many publications--for example, Theis (1935), Wenzel (1942), Ferris (1949), Jacob (1950), Todd (1959), and De Wiest (1965). Aquifer tests may involve well methods (point sink or point source), channel methods (line sink or line source), areal methods, and analysis of hydrologic boundaries (Ferris and others, 1962, and Bentall, 1963a). Some shortcuts and special problems in aquifer tests are discussed in other reports, such as that by Bentall (1963b).

### Estimating transmissivity in developed valleys

Variations in the transmissivity commonly are appreciable in valley-fill deposits of the alluvial basins in the Southwest. Thus, aquifer tests using pumpage at a specific site may not produce values applicable to large areas. Usually, it is not practicable to run aquifer tests on numerous wells in a given study area. In many areas, pump-efficiency tests have been run by the power companies on most of the large capacity wells. Those tests provide information from which to compute specific capacity (yield per foot of drawdown). Thus, if a relation between specific capacity and transmissivity can be demonstrated then transmissivities for large areas can be identified where substantial development has taken place. Thomasson and others (1960, p. 220-223) discussed this as follows.

"Gunther Thiem (1906) published a formula for determining permeability based on the flow of water into a discharging well. Wenzel (1942, p. 81) modified this formula into terms commonly in use in the United States, and, by transposition of terms, Wenzel's formula may be written

$$T = \frac{(P_f)m(528\ Q)\log\left[\frac{r_2}{r_1}\right]}{(s_1 - s_2)}$$

in which $T$ is transmissibility (gallons per day per foot),
$P_f$ is field permeability (gallons per day per square foot),
$m$ is the thickness of water-bearing material (feet),
$Q$ is the pump discharge (gallons per minute),
$r_1$ and $s_1$ are respectively the distance (feet) from a pumped well to a nearby observation well and drawdown (feet) in the observation well,

and

$r_2$ and $s_2$ are corresponding quantities for an observation well farther away.

"If the nearest observation well were placed against the outside of the casing of the pumped well, the $r_1$ would become the radius of the well ($r_w$) and the drawdown ($s_1$) would be ($s_w$), the pumping drawdown inside the well less the well losses. Likewise, if the farthest observation well were placed at the outer edge of the cone of depression around the well, the term $r_2$ would be the radius of this circle at which the drawdown $s_2$ would be zero. Substituting these values in the equation above:

$$T = \frac{528\ Q \log\frac{r_{cone}}{r_{well}}}{s_w}$$

in which $\frac{Q}{s_w}$ would approach the specific capacity as the entrance losses become a small part of the pumping drawdown measured inside the pumped well.

"To demonstrate the order of magnitude of the factor that can be multiplied by the specific capacity to approximate the transmissibility the following two examples are presented. Assume a pumped well 12 inches in diameter with a radius of influence that in half an hour of pumping extends outward for 3,000 feet, as might occur under confined conditions:

Then, $$T = \frac{528\ Q \log\frac{3{,}000}{0.5}}{s_w} = 528\frac{Q}{s_w}3.77 = 1{,}990\frac{Q}{s_w}$$

"Assume for the same pumped well a radius of influence in half an hour that extends outward for only 300 feet, as might be the case for water-table conditions:

Then, $$T = \frac{528\ Q\ \log\frac{300}{0.5}}{s_w} = 528\frac{Q}{s_w}2.77 = 1{,}460\frac{Q}{s_w}$$

Thus, a variation of 10:1 in the assumed radius of influence would change the factor by only 25 percent, whereas any change in the value of $s_w$ would be reflected directly in the answer. It would appear that approximate values of transmissibility to demonstrate the order of magnitude and also general variations from place to place within an area can be estimated from average specific-capacity figures for wells distributed over the area, so long as the entrances losses form a minor and consistent part of the pumping drawdown in the wells. Similarity in type of well construction would tend to stabilize the results obtained.

"The coefficient in the above two examples derived to estimate transmissibility from specific capacity ranged between 1,500 and 2,000 for the limits assumed. For the three pumping tests made in Solano County this coefficient ranged as follows: Test at well 7/1E-6C1, it was on the order of 1,600; test at well 8/1E-20G1, between 1,540 and 2,050; and test at 7/1E-32H4, between 1,300 and 2,200.

"Similar studies of other tests on valley-fill sediments in California indicates that a surprisingly consistent empirical factor is obtained for the type of well construction generally employed here. Furthermore, an average factor of 1,700 falls within about 10-15 percent of most of the individual test results. This empirical relationship between specific capacity and transmissibility was applied to the results of the pump-efficiency tests to demonstrate approximate transmissibility and to estimate quantities of ground-water flow through various parts of the area."

Theis, Brown, and Meyer (*in* Bentall, 1963a, p. 331-340) also discuss aspects of estimating transmissivity from specific capacity of wells.

Estimating transmissivity in undeveloped valleys

The transmissivity of the ground-water system may be estimated for natural conditions if the necessary dimensions, gradients, and quantities can be estimated or determined. Eakin (1966, p. 266) used this approach to provide a general concept of the regional movement of ground water in the Paleozoic carbonate rocks of the White River area in Nevada. The quantities used for selected underflow sections, were derived from the relative distribution of areas for which recharge and discharge estimates were made. The natural discharge was largely spring discharge that was relatively constant. The widths of the underflow sections were selected on the basis of the estimated dimensions of the regional ground-water system, and the gradients were those considered to be the regional average minimum. The formula used for the computation was

$$T = \frac{Q}{0.00112\ IW}$$

where $T$ is the transmissivity, in gallons per day per foot; $Q$ is the underflow, in acre-feet per year; $I$ is the gradient, in feet per mile; $W$ is the effective width of the aquifer, in miles; and the constant 0.00112 is the factor to convert gallons per day to acre-feet per year.

As noted by Eakin (1966, p. 266),

> "The estimated transmissibilities for the three sections were computed by using the [above] equation. . . .these values suggest that a first approximation of the regional transmissibility of the Paleozoic carbonate rocks is on the order of 200,000 gallons per day per foot. The value is not large considering the substantial thickness of the carbonate rocks. However, as the actual transmission in the carbonate rocks is localized largely in fracture or solution zones, local transmissibility values undoubtedly are much higher, perhaps 10 times or more, than the indicated regional average. On the other hand, large areas of carbonate rocks that have little or no fracturing or solution openings transmit very small amounts of water."

A similar approach may be used for ground-water systems in closed valleys. If all the ground water discharged naturally in the lowland is supplied by recharge in the upper part of the alluvial apron or in the mountains, then there will be a peripheral segment of the alluvial apron below which ground-water movement is essentially lateral. A line drawn peripherally around the natural discharge area in this segment will define the width of "aquifer" through which all ground-water flow must pass. Based upon (1) estimates of the annual amount of water naturally discharged, (2) the length of the peripheral line, and (3) the average water-level gradient across that line, the transmissivity can be estimated in a rough way. It is emphasized that the results are useful only in forming a general concept of the quantities of water moving within the ground-water system. The results may have little value for estimating transmissivity at individual well sites because they represent broad averages and obviously local variations of an order of magnitude or more can be expected.

In lesser degree some idea of transmissivity may be obtained by analogy of the geologic environment to similar developed areas where transmissivity values have been adequately determined.

## Recharge

The valley-fill reservoirs are recharged naturally by (1) direct infiltration of part of the precipitation on the valley-fill deposits, (2) seepage of part of the mountain runoff into the bed and banks of stream channels on the valley-fill deposits, and (3) subsurface flow from the consolidated rocks.

In the generally idealized system most of the recharge to the ground-water reservoir is derived from streamflow seepage and a much lesser proportion occurs as direct infiltration of precipitation; the subsurface flow from the mountains that may occur is usually considered negligible. However, within the region of alluvial basins, the proportion between the three types of natural recharge may differ significantly from the idealized example. In developed areas return flow from irrigated fields, and seepage or outflow from canals, ditches, storm drains, and sewers may also contribute significantly to the recharge and therefore require evaluation. Rejected and artificial recharge may also be important in some areas.

The evaluation of recharge may involve several methods of investigation depending upon the data available. Ordinarily, data may be available or can be obtained on one or more of the following: Streamflow, precipitation, water lost by seepage, infiltration rates, vertical permeability, specific yield of sediments within a zone of water-level fluctuations, and lateral permeability or transmissivity.

## Seepage from streams

Determination of the recharge attributable to seepage from streams may be based either on measurement of streamflow losses or on changes in storage in the ground-water reservoir.

*Determination of recharge based on measured streamflow losses.*--A relatively direct method of estimating recharge is the measurement of losses from streams between two or more points of control. Measured losses from streams include evapotranspiration, increase in soil-moisture storage, and seepage to the ground-water reservoir. Artificial diversions, if any, between points of measurements also must be considered. Seepage to the ground-water reservoir is determined by selecting periods when evapotranspiration losses are at a minimum and when soil moisture storage requirements have been satisfied. Loss rates during those periods are then extrapolated to obtain annual values. Due regard is given to the relation of rate of loss to rate of flow. This method of estimating recharge works best for perennial streams with a moderately stable flow regime because soil moisture storage is then relatively constant. The problem of accounting for soil moisture deficiencies below field capacity increases as the duration of streamflow decreases and becomes most difficult where the channel carries water only after occasional storms. Intensive studies may include neutron-probe measurements in test holes (p. 18) to provide valuable data on relative moisture changes in the soil adjacent to the access tubes.

The measurements of seepage losses may not be reliable if the losses are very small in comparison with the measured flow. For example, during periods of high flow, the magnitude of the seepage loss in some channels may be less than the probable error of measurement. Nevertheless, the seepage loss from streams is probably the most directly measurable element of ground-water recharge and, with care, should be the most reliable component of the recharge estimate.

Seepage losses, being related to runoff, may be highly variable from year to year. Worts (1951, p. 79) in his ground-water study of the Santa Maria Valley, Calif. estimated the annual recharge by seepage from streams to range from 4,800 acre-feet in 1931 to 150,000 acre-feet in 1941.

Briggs and Werho (1966) described the infiltration and recharge resulting from an unusually large controlled release of water flowing in the normally dry Salt River channel near Phoenix, Ariz. The 20,000 acre-feet of water released to the Salt River channel below Granite Reef Dam, was reduced by seepage to about 100 acre-feet at 7th Avenue, a channel distance of about 25 miles. Calculated seepage rates were 2.5 feet per day in the reach between Granite Reef Dam and 48th Street, and 1.4 feet per day from 48th Street to 16th Street. For gravel pits in the channel upstream from 7th Avenue initial seepage rates were calculated to be 1.5 feet per day and averaged 1.1 feet per day for the 10-day period following the end of flow into the pits. Streamflow measured at six selected sites were used to calculate peak flows, volume of flow, and downstream alteration of flow. Measurements of water level in observation wells showed rises in response to recharge. The maximum rise, 25 feet, occurred in a well about 500 feet from the river.

Extension of the results obtained for particular streams to the valley as a whole requires data or estimates of runoff from the contributing areas. These may be obtained by several of the methods discussed in the surface-water section of this report. The large seasonal and year-to-year variation in runoff commonly presents difficulties in determining average runoff from which to estimate average recharge. Determining average runoff requires long periods of record as a basis of most methods. This problem appears to be significantly reduced by the channel-geometry method developed and used in Nevada by Moore (1968). The method is undergoing further development. That method, through the measurement of channel dimensions as defined by the occurrence of berm and bar forms, permits rapid determination of average runoff values in the section of maximum flow at the mountain fronts. Sequential measurements along the stream channel also permit delineation of the proportional distribution of runoff within the contributing area of the stream. Current work of D. O. Moore and J. R. Harrill (oral commun., 1968) in Paradise Valley, Nev., indicates the possibility of obtaining reasonable estimates of average stream loss between selected sites by the channel-geometry method (p. 48), in reaches where seepage loss is significantly large with respect to streamflow.

*Determination of recharge based on storage changes.*--Where the direct measurement of seepage from streams is impractical, that type of recharge may be estimated from changes in ground-water storage. The method involves (1) delineating the area in which water-level fluctuations occur due to seepage from a stream or streams, (2) determining the volume and effective storage capacity of the rock materials in which the fluctuations occur, and (3) relating seasonal or other period fluctuations to streamflow and seepage losses. The mechanics of computing storage changes were discussed earlier.

Numerous investigators, including Clark (1915) and Piper and others (1939) have used the method. It may also be used as a check on the results obtained from incomplete measurement of all streamflow losses, as when the measured losses from a particular stream are extrapolated to the valley as a whole. In irrigated areas the method can be used to determine the recharge attributable to the seepage of applied water.

## Direct infiltration of precipitation

Recharge also occurs as a result of direct infiltration of precipitation in amounts sufficiently large to exceed the soil-moisture deficiency. The amount of recharge varies significantly with precipitation, soil conditions, and infiltration characteristics. Direct practical measurement of infiltration is limited to point determinations, as by lysimeters or tanks. The data then are extrapolated to larger areas on the basis of equivalent or proportionally different soil, slope, and precipitation. Blaney, Taylor, and Young (1930) used this method in their study of rainfall penetration in the Santa Ana River Valley and coastal plain. Rainfall and runoff at test sites were recorded and the penetration of simulated rain was observed in tunnels 12 and 20 feet below the surface of an alluvial fan. Sampling with soil tubes to determine water content at various depths permitted determination of soil-moisture deficiency, and changes in soil moisture identified rainfall penetration. The measurements were made under conditions representative of the general range of physical conditions in the area. The observed data were used together with areal distribution of precipitation, temperature, and soil type to compute rainfall penetration for the area.

Blaney (1933, p. 82-90) used similar rainfall-penetration test plots, observed for one season, in applying the method in an investigation in Ventura County, Calif. Worts (1951, p. 80-82), in studying the Santa Maria Valley area, did not use test plots to determine recharge from precipitation, but adapted the results of Blaney's investigation in Ventura County.

## Subsurface inflow from consolidated rocks

Although subsurface inflow is commonly considered to be a minor component of recharge, Eakin found it to be dominant in some valleys in Nevada. Precipitation infiltrated to bedrock in the surrounding mountains and reached the valley-fill reservoir as subsurface inflow. Eakin (1966) estimated the recharge from a precipitation-recharge relation. His method will be discussed in the following section of this report rather than at this point, because the method does not pertain to the subsurface-inflow component of recharge as such; it is a method of computing total recharge.

Several investigators have estimated the subsurface-inflow component by first computing total water yield by one of the methods described earlier in this report (p. 37-49) and then subtracting the streamflow. For example, Feth and others (1966 p. 41, 42) in their study of the Weber Delta district, Utah, found evidence of subsurface flow along the mountain front of the Wasatch Range. Evidence included configuration of the potentiometric surface, chemical-quality data, radon data, and discharge measurements and characteristics of the flow from the Gateway tunnel. By using the Oltman-Tracy (p. 43) method they computed water yield, and from that value subtracted streamflow to obtain the quantity of subsurface flow that subsequently recharged the ground-water reservoir.

Mundorff and others (1964) in their study of the Snake River plain, likewise computed subsurface flow as a residual, having first obtained a value of yield by use of the Mundorff method (p. 44). In that particular study the subsurface flow was outflow from the study area that subsequently recharged a down-gradient ground-water basin.

Determination of average annual total recharge from precipitation-recharge relation

Eakin (1966, p. 260) described a method of estimating total recharge from an empirically derived precipitation-recharge relation. From the results of regional reconnaissance investigations, the following table was empirically prepared based on 1,000-foot altitude zones in that region. The values in the table were then applied to altitude zones in the study basin to obtain the total basin recharge.

| Altitude zone (feet) | Precipitation zone (inches) | Assumed average annual precipitation (feet) | Assumed average annual recharge to ground water (percentage of average precipitation) |
|---|---|---|---|
| Below 6,000 | Less than 8 | Variable | Negligible |
| 6,000 to 7,000 | 8 to 12 | 0.83 | 3 |
| 7,000 to 8,000 | 12 to 15 | 1.12 | 7 |
| 8,000 to 9,000 | 15 to 20 | 1.46 | 15 |
| More than 9,000 | More than 20 | 1.75 | 25 |

The above table is applicable only to the region for which it was derived. If Eakin's approach is used for another region, the specific values in the table will probably have to be adjusted on the basis of a careful consideration of the geologic, topographic, and hydrologic parameters of that region. Furthermore, although such a generalized table can give reasonable estimates of recharge for large areas in a region, it may give poor estimates when applied to an individual valley in the region.

## Computation of average annual total recharge as a residual in the hydrologic budget

In the earlier discussion of subsurface inflow, it was shown that this component of recharge may be computed as the residual when streamflow is subtracted from water yield. Average annual total recharge may also be computed as the residual in the average annual hydrologic budget when all other elements in the budget have been computed or estimated. An example of this is given in a report by Loeltz and others (1949) describing an investigation in Paradise Valley, Nev. By examining long-term runoff and precipitation records for the region they found that approximately all precipitation in excess of 9 inches became runoff. From that relation and a knowledge of the mean annual precipitation by elevation zones, mean annual runoff for the valley was computed. From a knowledge of the irrigated acreage and the consumptive use of the irrigated grasses and riparian vegetation along streams and ditches, Loeltz and his colleagues estimated that about two-thirds of the mean annual runoff was lost to evapotranspiration, leaving about one-third available for recharge to the ground-water reservoir in the valley.

## Conclusions

Because of the difficulty in obtaining adequate data, it is difficult to estimate recharge reliably. Stream-channel seepage can often be measured directly and where the great bulk of the recharge is supplied by such seepage, satisfactory estimates of recharge are often obtained. Generally, however, it is necessary to rely on a hydrologic budget where all elements other than recharge are estimated and recharge is computed as a residual. The figure of recharge so obtained is not only dependent on the validity of the estimates used for the other elements of the budget, but there is no way of checking the reliability of the computed recharge. In undeveloped basins the most desirable course of action, where feasible, is to estimate average annual natural discharge from the ground-water reservoir and proceed on the premise that over the long term, average annual natural recharge equals average annual natural discharge in a given ground-water system. Methods of estimating ground-water discharge are discussed in the section of this report that follows.

## Discharge

Ground water is discharged naturally by evaporation and transpiration and by surface or subsurface outflow. In the idealized example of the closed basin, all the ground water is discharged naturally by evapotranspiration. Under some conditions the discharge may be by subsurface outflow (Eakin, 1966). Locally, a considerable proportion may discharge from springs and subsequently be removed by evapotranspiration. Where a valley is tributary to another, or to the sea, significant outflow may occur either as subsurface flow or as streamflow. In developed valleys a significant amount of water is discharged artificially through wells.

### Evapotranspiration losses from ground water

Several methods of determining rates of evapotranspiration and evaporation were discussed in an earlier chapter and are not repeated here. Relating experimentally determined evapotranspiration rates to large areas of phreatophytic vegetation is an uncertain procedure and continues to be a vexing problem in ground-water investigations. Field rates of evapotranspiration must be extrapolated from in-area experimental plots or by analogy from other areas. Commonly, limitations of time preclude local experimental determinations, and rates from other areas are used. In either case the experimentally determined rates rarely span the full range of field conditions for most of the phreatophyte species. Thus significant error can result in extrapolating rates to field conditions. Surveying the field distribution of the phreatophytes by the standard method (Horton and others, 1964) requires a considerable amount of time in proportion to that available in most investigations. This results in the use of a variety of short cuts in delineating the distribution of phreatophytes and often leads to a mere delineation of the gross area in which phreatophytes occur. This is useful for reconnaissance investigations but is inadequate for more intensive studies.

Experimental work is being done to develop energy-budget and mass-transfer methods of computing evapotranspiration, but no fully operational procedure is yet available. It will be recalled that these two methods are widely used for computing evaporation from open-water surfaces.

### Surface outflow

The surface outflow from a ground-water body may occur as the discharge of springs and seeps or as the base flow of perennial or intermittent streams. Surface outflow can usually be gaged satisfactorily by the use of volumetric or current-meter measurements to define a stage-discharge relation.

*Discharge of springs and seeps*.--In some areas the discharge of springs may represent a moderate to large proportion of the total natural discharge of ground water. The discharge may be localized in well-defined channels and thus be susceptible to accurate measurement. In many places, however, only a small fraction of the spring discharge is channelized with the remainder issuing as seeps over an extensive area. In the latter case the channel flow is a poor indicator of total discharge, and part or all of the discharge must be estimated indirectly by estimating the evapotranspiration from the spring area. Under some conditions the importance of evaporation can be minimized to a degree by measurement of channel flow during the winter period of minimum evapotranspiration.

Eakin (1966) utilized the discharge of springs as a key element in describing a regional ground-water system in southeastern Nevada. A sequence of measurements on Muddy River springs showed relatively uniform discharge of about 50 cfs (cubic feet per second). Eakin and Moore (1964) related the series of spring measurements to a nearby long-term gaging station on the Muddy River by adjusting for the effect of locally developed overland runoff and for evapotranspiration between the gage and the spring areas. With those adjustments the Muddy River gaging record effectively represented long-term spring discharge.

In a study of Clear Lake Springs in Millard County, Utah, Mower (1967, p. 28) found that fluctuations responded to both seasonal variations of precipitation and to pumpage in upgradient areas, particularly the Flowell District.

Mundorff and others (1964, p. 171-181) analyzed the spring discharge to the Snake River between Milner and King Hill. By subtracting measured or estimated inflow from other sources, they attributed about 6,500 cfs of the total gain of 8,150 cfs in 1955 to ground-water discharge from the Snake Plain aquifer.

It is evident, then, that determination of the quantity and variations in spring discharge may be significant and provide valuable information for defining the ground-water system in some areas.

*Base flow of perennial or intermittent streams*.--Ground-water discharge sustains the base flow of perennial or intermittent streams. All of the "fair-weather" discharge of a stream that is not fed by spring (season) or summer snowmelt is base flow. During a fair-weather period base flow declines with time, the rate of decline being governed by the relation of outflow to volume of ground-water storage available for base flow. In other words, if the volume of base flow in a day reduces the head on the ground-water reservoir only slightly, the decline in base flow will be slight; if the head is reduced significantly, the decline in base flow will be more rapid. During periods of surface runoff or snowmelt, base flow increases as a result of recharge to the ground-water reservoir which in turn increases the head on the reservoir. Methods of analyzing the discharge hydrograph to separate the base-flow component of runoff from surface runoff and interflow are given in most standard hydrology texts (for example, Linsley and others, 1949, p. 397-404).

The amount of ground water discharged to a stream usually varies along the stream. Changes in discharge may be studied in segments between a series of gaging stations, or by means of seepage runs involving a series of discharge measurements repeated at intervals at selected sites along the stream. In a simple situation where flow is steady at the upstream and downstream sites on a reach of stream, the measured gain in discharge between sites represents ground-water outflow. In a more complex situation such elements as surface inflow within the reach, pumpage, evapotranspiration, and change in storage must be considered in computing ground-water outflow. Gatewood and others (1950, p. 163) in a study of Safford Valley, Ariz., considered 17 elements in the equation of inflow, outflow, and change in storage.

## Subsurface outflow

Parts of the ground water in some valleys is discharged by underflow. In most basins, underflow is considered to be confined to the alluvial deposits underlying the flood plain or contained in a bedrock narrows. For those conditions, underflow can be computed from determinations of the cross-sectional area, unit permeability, and gradients. Worts (1951, p. 76,77) used that approach in his study of Santa Maria Valley, his value of permeability being estimated from a test on a well 2 miles downstream from the section used for computation. Estimates of underflow can be developed based on extensive test drilling and other tests in the underflow section to determine gradient, cross section, and permeability. That degree of control provides a firm estimate of underflow. More commonly, however, only approximate values of the three parameters are obtained and the estimate of underflow is therefore also approximate. In the absence of subsurface control, the unit permeability or unit transmissivity and the cross section are sometimes assumed from general geologic conditions to obtain a rough estimate of the quantity of underflow (Eakin, 1961, p. 24).

If underflow is more extensively distributed so that the cross-sectional area cannot be specifically defined, underflow may be identified on the basis of the quantity of water that moves between the area of recharge and the area of discharge, where estimates or determinations of recharge and discharge have been made in the manner used by Eakin (1966, p. 266) and discussed in the previous section on transmissivity.

## Withdrawal by pumping

In areas of ground-water development, withdrawals from wells commonly are a significant part of the total ground-water discharge. Even in areas having only a few wells, a knowledge of the withdrawals is of value in better understanding the ground-water system. As such, attention must be given to well inventories, pumpage yield, and overall performance. However, because of the size of many ground-water systems substantial pumpage is required before the effect on the system is significant.

The methods of determining the annual discharge of a well are partly controlled by the well outlet arrangements, the pattern of pump operation, the variation in pumping level through the pumping interval, the type of power driving the pump, and the design and efficiency of the pumping equipment.

*Volumetric determination.*--The determination of the volume of water discharged by the well or well field by continuous measurement is the most accurate method of determining annual volume of ground-water withdrawal. This may be done by totalizing flow meter, or by continuous recording of stage or head of water flowing through a weir, flume, or other flow-rated conduit.

*Determination based on period of pump operation.*--If the time that pumps are operated can be determined, such as with hour-totalizing meters, the discharge rate can be determined and total annual withdrawal estimated.

*Determination from power use by electric-powered pumps.*--Electrically driven pumps commonly have a consistent relation of rate of power use to rate of discharge. Determination of that relation together with annual power consumption permits determination of annual ground-water withdrawal. If the pumping lift significantly varies during the year, the wire-to-water efficiency per unit of water per foot of lift should be determined to compute annual discharge. Wire-to-water efficiencies should be determined if discharge determinations based on a few wells are to be applied to many wells by means of total power consumption and average pumping lifts.

*Determination from fuel use by fuel-powered pumps.*--Pumps driven by oil, gasoline, or natural or manufactured gas commonly have a somewhat variable relation between rate of discharge and rate of fuel use, due to variations in pumping operation. This is somewhat analogous to automobiles where different driving patterns result in differing rates of fuel consumed per mile of driving. The resulting estimates of ground-water withdrawal by fuel-operated pumps obviously are less reliable than those similarly estimated for electrically driven pumps, other conditons being equal.

*Determination from estimates of water use.*--Estimates of withdrawals by other methods may be checked by estimating the amount of water used for agriculture, industry, and public or rural supplies. The acreage under irrigation commonly can be determined readily. That acreage can be divided by crop groups using different amounts of water. Evaluation of irrigation efficiency, including conveyance losses and leaching requirements, permits estimates of the quantity of water pumped to meet those needs. The several elements of the combined estimate are subject to errors which may be either compensating or additive. Thus, the estimates of the individual elements should be made with careful consideration of the irrigation practices of the area under study.

Commonly, pumpage records are kept for public and industrial supply uses. If these data are not available, one or more of the previously discussed techniques may be used. On occasion it may be necessary to estimate the use by analogy to similar operations elsewhere where water use demands are comparable. Thus, a public-supply use may be estimated if the approximate population is known and if the use has been determined for another public-supply community having similar water-use conditions. Per capita use of water in the community where water use has been determined can be applied to the community under consideration.

Poland (1959, p. 14-28) used most of the above-described methods in estimating withdrawals in the heavily developed Long Beach-Santa Ana area in southern California.

## Chemical Quality

Unlike surface-water quality, which is subject to rapid change and a wide range due to the vagaries of nature or the influence of man's activities or both, ground-water quality remains relatively uniform in time and exhibits a narrow range. Although the principal single factor that controls the chemical characteristics of ground water is the geologic environment to which the water is exposed, other factors also exert an influence. Those factors include the chemical quality of both artificial recharge water and irrigation return flow, excessive withdrawals that cause a change in the hydraulic gradient and eventual intrusion of water of differing quality, and the disposal of wastes in such manner that they reach a ground-water body.

Most ground-water investigations include some sampling and analysis of the chemical quality of the water, usually for the purpose of determining the suitability of the water for use. Aspects of the collection and handling of water samples and analytic techniques for the determination of the water properties are well discussed by Rainwater and Thatcher (1960). Because ground-water quality is less variable in time than surface-water quality, ground water is usually sampled less frequently; continuous monitoring of ground-water quality is generally unnecessary because gradual long-term changes or trends can be defined by periodic sampling. Each major aquifer system and subsystem should be sampled.

Ground-water samples are usually obtained at existing wells, or their effluent discharges are sampled at springs or in stream segments in which the surface-runoff component of flow is relatively insignificant. The movement of water within the ground-water body itself is usually slow and nonturbulent so that water of different qualities may be stratified within a single thick aquifer. Commonly, a single well will tap two or more aquifers that contain water of distinctly different qualities, and each aquifer may yield water at different or variable rates. Although the chemical quality of the water in each aquifer may remain constant, water of variable quality may be obtained at different phases of the punping cycle, because pumping may vary the yield of one or more of the aquifers. The variation in quality of the composite water that is withdrawn may be diurnal, seasonal, yearly, or may involve an even longer period.

In view of such complexities, a ground-water body ideally should be sampled from numerous wells and springs. The hydrologic regimen of the ground-water body should be determined in considerable detail and the place of each sample in that regimen should be established in both space and time. In areas where wells of large yield and various depths are numerous, it may also be desirable to obtain extensive information concerning: Depth and type of well casings and screens; type, capacity, depth of setting, and operating schedule of pumps; and fluctuations of the water level in wells, including those that are not sampled. With information any less comprehensive it may be impossible to interpret the analytical data fully. The ideal network of sampling points discussed above is seldom found; usually the distribution of available sampling points can provide only a rather inadequate representation of the overall chemical character of the ground-water system. That is especially true in undeveloped desert valleys where data from even fewer sampling points are available.

As for sampling methods, if a well is being pumped steadily a grab sample from the discharge is usually adequate, if the sample can be related to the drawdown and to the elapsed pumping time. If there is no active pump on the well, it is desirable to install a temporary pump and operate it long enough to be sure that a sample of the discharge represents water coming directly from the aquifer. Field tests for temperature, pH, and conductivity at the time of sampling provide useful clues to the ground-water hydrology. Because the ground water of an area is often related to the surface water, sampling schedules for each should be correlated. The chemical characteristics of water as it exists in a deep aquifer may change drastically when the water contacts the atmosphere. Special procedures in sampling are therefore required if the hydrologist is concerned with the chemical nature of the water as it exists in the aquifer. Appropriate techniques are described by Barnes (1964) and Back and Barnes (1965).

To return to a discussion of the geochemistry, we refer to Hem (1959, p. 4) who comments:

> "Natural waters acquire some of their chemical characteristics through direct solution of some of the solids, liquids, and gases with which they may come in contact in the various parts of the hydrologic cycle where water is in the liquid state. In addition, in the presence of water, chemical reactions may occur in which soluble products are formed.
>
> "The final composition of a water is the result of a number of solutional and decompositional processes. The several chemical systems that are usually involved may present a very complicated combination, but one which may often be evaluated at least semi-quantitatively by the use of principles of theoretical chemistry. Certain reactions--for example, the solution of carbonates and the exchange of cations between solutions and clay minerals--are reversible and may lend themselves fairly well to theoretical treatment. The chemical processes are affected by certain variables in the environments of natural waters. Some of these factors include the type of mineralogic and geologic environment, amount of water available, its rate of circulation, the activity of micro- and other organisms, and temperature and pressure. If the chemical reactions that are involved can be fairly well described and understood, the results as indicated in water analyses may give some basis for evaluating the environmental factors.
>
> "Much research in natural water chemistry remains to be done before a full realization of the value of water analyses as an interpretative hydrologic tool can be realized."

Where considerable data are available to show the distribution of chemical quality in a ground-water system, water commonly can be grouped by types. The distribution of water types may lead to significant interpretations such as that achieved by Piper, Garrett, and others (1953) in the study of native and contaminated ground waters in the Long Beach-Santa Ana area of southern California. In areas of lesser development, rough areal or vertical grouping of water type may be possible with some degree of confidence. In valleys of essentially no development, sampling points of natural discharge may provide the only clues to the chemical quality of the ground-water system. Under these and commonly under other conditions the chemical-quality information may provide supporting evidence for flow-system interpretations based on other kinds of data.

White, Hem, and Waring (1963) list a number of analyses of ground water associated with different rock types. They show that a wide range of concentrations may occur in waters associated with a given rock type. However, some generalizations are indicated by the authors. For example, ground water from silicic igneous rocks tends to be of low mineral content with sodium and bicarbonate as dominant ions; silica is relatively high for cold dilute waters and fluoride is relatively high. Ground water from gabbro, basalt, and ultramafic rocks generally has high ratios of calcium to sodium and magnesium to calcium; the magnesium content of waters from gabbro and basalt is nearly always higher than that of water from silicic rocks; silica commonly is high and fluoride usually is low. Water from carbonate rock generally is characterized by high calcium, magnesium, and bicarbonate; the proportions of calcium and magnesium varying in response to variation in proportions of limestone and dolomite. In the Paleozoic carbonate rocks of eastern Nevada the calcium to magnesium equivalents ratio commonly ranges from about 1/0.50 to 1/0.85, and concentrations are in the range of 200 to 300 milligrams per liter. The concentrations are smaller in areas close to where recharge occurs and of course may be greater where water from other rock types has been added (Eakin, 1966, p. 266-268).

Constituents may function as "tracers" under favorable conditions. Thus radiocarbon, oxygen isotope, and tritium dating may be useful in approximating the rate of travel if the relative positions of the samples in the flow system are reasonably well defined. Similarly, injected tracers can be used, but they usually can be applied for checking short-distance movement only.

## AREAL INVESTIGATIONS

Areal or basinwide investigations are aimed at providing information that will be used in planning the optimum development of the water resource within the existing framework of water law or custom. However, there is usually no stated philosophy, as such, behind this planning in arid regions. Generally, cultural development begins in a basin and is supported by the local water supply. If the cultural development proceeds at a rate that exceeds the capability of the local supply, water is imported from other basins to meet the increasing water demand. Another alternative is to transfer the development from one basin to another as the local supply is exceeded. Because there is usually no master plan for the orderly development of large arid regions, and because of the high cost in time and money of making detailed hydrologic studies, such studies are generally made only in those areas that now have major economic significance or in those areas that are established as experimental basins for research in arid-land hydrology. Hydrologic studies made in other undeveloped arid basins are generally of the reconnaissance type to appraise the potential of the local water supply. Until such time as master plans of development evolve, that course of action, with respect to reconnaissance-type and the more intensive detailed hydrologic studies, will be the most practical one to follow. After reconnaissance-type investigations have been made for the entire area for which a water-resources agency is responsible, the number of detailed basin studies in the area may be expanded.

. The number of years required to complete a detailed investigation is a principal factor in limiting the use of this type of investigation. For example, in many basins there are not only few precipitation and runoff events in a series of years, but there is also such great variability in the geographic distribution of the events, that intensive investigation would require the operation of many costly recording instruments over a long period of years before meaningful analysis of the hydrology could even be attempted. Despite the infrequency of hydrologic events, operating costs are usually high because of the distant location and relative inaccessibility of many of the basins. As for detailed studies of the geohydrology, the cost of a desirable observation-well program, including the monitoring of pumping withdrawals, can be a very appreciable budgetary item, and the cost of a complete exploration of ground-water conditions in alluvial fans will usually be prohibitive. There are, of course, all gradations of intensity with which a detailed study may be made.

A reconnaissance investigation, on the other hand, usually lasts from a few months to a year or two. During that time all pertinent data available are assembled and are supplemented with whatever information is needed for at least a rough appraisal of the local water supply. Obviously if a great deal of data are available--as for example, in an area where the geology had been previously mapped in detail for an oil exploration, or where considerable well data and water-level data are available--the scope of the geohydrologic reconnaissance may, in certain aspects, approach that of a detailed investigation.

For the purpose of this paper, areal investigations can be grouped into five general categories in which the level of intensity, in the past at least, has been governed largely by the degree to which the ground-water body has been developed and by the problems resulting from that development. Existing wells usually provide most, if not all, of the subsurface information used in the investigations.

*Reconnaissance evaluations.*--Preliminary or reconnaissance studies commonly involve field investigations aimed primarily at observing the physical environment as related to basic hydrologic and associated physical principles. The amount of specific data available for the different classes of information varies widely but is usually scant. Therefore reliance must be placed on the observable features of geology, geomorphology, climate, natural vegetation, streams, and ground water. For such observation, reconnaissance from the air by light plane may be most helpful. Subsurface information on geology or ground water may be negligible or absent. The investigator bases his evaluation on the observed or obtained information. The experience and ability of the investigator are the most important factors in the reliability of the resultant evaluation of the ground-water system.

*Intermediate evaluations*.--These studies commonly are undertaken, in Nevada for example, after at least five years of development. Principally, they seek to describe the cause and effect relations of development of the ground-water system. Data on water withdrawals and the water-level changes obtained from existing wells form the basis of analysis. These studies afford an early check on the effect of development on the system. The studies provide an opportunity to identify possible problems before they become critical and to identify the potential for additional development. If the study is the first for that area it may also include a reconnaissance or higher level study of the overall hydrologic system.

*Intensive evaluations*.These studies seek to describe the hydrologic system comprehensively and to analyze the ground-water development and its effect on the system. Past and current data on all facets of the hydrology provide the controls for the analysis. Ideally, these studies would obtain or provide adequate data to fully support and sustain the quantitative and qualitative interpretations.

*Continuing inventories*.--In areas of significant ground-water development, continuing inventories provide data on the withdrawals and their effect on the ground-water system. These data record changes with time, an important factor in hydrologic analysis. Continuing inventories usually include repetitive measurements of pumpage, water levels, and chemical quality of water. Also, related measurements of streamflow losses or gains, precipitation, and other appropriate information commonly are obtained. Ordinarily, continuing inventories are most effective and useful in areas where intensive studies are in progress or completed. However, well-planned, long-range programs should include sound continuing inventories which would be started several years in advance of an intensive study.

*Special studies*.--Various types of studies are included in this group. They are characterized by their emphasis on one or two aspects or phases of surface and (or) ground water. For example, these studies may be concerned with variations of transmissivity in relation to the yields of wells, changes of water quality in response to withdrawals, or the effect of withdrawals on nearby streamflow, an understanding of which is needed to alleviate a current operational problem.

The individual elements in the basic hydrology of a basin have been discussed in preceding sections of this report. Separate treatment of these elements is not always ideal, however, because such treatment obscures the interrelations among the elements. A logical place to summarize and integrate the earlier material is in a discussion of methodologies for conducting both reconnaissance and detailed intensive hydrologic investigations in arid regions. Such a discussion follows. Impressions and conclusions reached in the course of this bibliographic study have been interspersed in the preceding material and they appear again in the discussion of recommended investigational procedures.

Although, as stated earlier, there are all gradations of intensity with which areal studies may be conducted, the two hypothetical studies that will be used as illustrations are somewhat typical. However, emphasis on some of the facets of a study will vary regionally with local or State water-rights law. In some regions the safe-yield concept is followed and withdrawals of ground water are limited to the average volume of perennial yield less the unintercepted natural discharge; in other regions ground water is "mined" from storage with little concern for the magnitude of perennial yield; in still other regions a concept of optimum ground-water withdrawal is being developed which considers the concepts of both safe yield and mining, along with the conjunctive use of surface or imported water (Maxey, 1968).

## Reconnaissance Investigations

The method of making reconnaissance investigations that is described below is based largely on the methodology used by the U.S. Geological Survey in its studies in Nevada. Several of the Nevada reports are cited. A reconnaissance investigation of a basin will usually include the following elements:

Description of the hydrogeology
Chemical and thermal quality of water
Water development
Areal distribution of average annual precipitation and runoff
Time distribution of precipitation and runoff
Flood frequency
Average annual hydrologic budget
Source, occurrence, and movement of groundwater
Perennial yield and potential development

The various facets of the investigation are discussed on the pages that follow.

*Description of the hydrogeology.*--The conventional methods of hydrogeologic investigation are used in desert regions, and those methods, therefore, will not be described here. However, because of the limitations of time and money, great dependence is placed on whatever data are already available from all sources, and additional exploratory work is kept to the minimum needed to fulfill the objectives of the investigation. It is necessary that the location, extent, and boundaries of aquifers and aquicludes be determined or inferred. Altitudes and depths below land surface of the water table or potentiometric surface likewise must be determined or inferred. The valley should be classified as to whether it is topographically open or closed; and if closed, the ground-water data should be used to identify the valley as being either drained or undrained (p. 8-9). Aquifer coefficients such as the coefficients of permeability or transmissivity and the specific yield or storage coefficient also must be determined or inferred. The location and discharge of springs, and fluctuations of their discharge, often offer indication of the permeability of the aquifer, the depth of the water table, and the maximum quantity of recoverable water in the immediate area.

Areas of evapotranspiration from the ground-water body, if any, should be mapped. Davis and De Wiest (1966), in discussing phreatophytes, state:

> "Phreatophytes are one of the most useful surface indicators of ground-water conditions (Meinzer, 1927). The total area of vegetation gives some rough indication of the total amount of water being discharged at the surface. By identifying the species and measuring the density of vegetation (and obtaining the depth to the water table), a rather close estimate can be made. Even without this work, a simple measurement of area from aerial photographs can be multiplied by a reasonable value for transpiration to obtain a first estimate of quantity (Mann, 1958). For example, if 250 acres are covered with dense vegetation, the transpiration most probably lies between the values of 2.0 feet and 7.0 feet of water per year. If a reasonable value of 3.5 feet per year is assumed then the estimated discharge would be 875 acre-feet per year, or a constant discharge of about 540 gpm, enough to supply a town of more than one thousand inhabitants.
>
> "Phreatophytes give some indication of depth to water. Grasses thrive where the watertable is generally less than 10 feet below the surface, shrubs where the watertable is less than 30 feet below the surface, and trees where the watertable is less than 90 feet below the surface. These are rough figures at best. Actual depths depend on the width of the capillary fringe, amount of surface rain, actual plant species, and other factors.
>
> "Phreatophytes also give some suggestion of water quality. Willows and cottonwood usually grow where the water is potable and pickleweed usually grows in places where the soil is saturated with saline water. Many other phreatophytes such as palms, mesquite, and salt grass may grow with roots in either potable or brackish water."

The presence of phreatophytes, however, does not assure the availability of recoverable ground water at the site, because phreatophytes can abstract water from saturated clays and fine silts which would not contribute water to a well.

*Chemical and thermal quality of water.*--The extent to which water dissolves chemical constituents from the rock materials is governed largely by the solubility, volume, and distribution of the rock materials, the time the water is in contact with the rocks, and the temperature and pressure in the ground-water system. Consequently the chemical constituents and temperature of the surface and ground waters not only indicate the suitability of the water for various uses, but are also helpful in providing clues to the geology and source of the waters. The investigation should therefore provide temperature and chemical quality data.

*Water development.*--Water development, as used here, refers to any activities by man that affect the natural regimen or hydrologic budget of the basin. Included are withdrawals or diversions from the ground-water or surface-water supply, impoundment of water in stock ponds or reservoirs, and irrigation activities. The objectives of the investigation require that pumpage and diversion be inventoried, that sufficient data be collected concerning stock ponds and reservoirs so that evaporation and seepage can be estimated, and that sufficient data be collected concerning irrigation practices so that the quantities of water applied, consumed, and returned to the ground-water body or surface stream can be estimated.

*Areal distribution of average annual precipitation and runoff.*--An isohyetal map of average annual precipitation or a relation of average annual precipitation to elevation should be prepared for the basin. If reconnaissance studies are eventually to be made in many basins in a region, it is recommended that an isohyetal map be prepared for the entire region, because it is much more economical in time and labor to prepare one large map than it is to prepare several small ones, and more consistent results are assured. The previously described "altitude-anomaly" method is recommended for preparing the isohyetal map. The relatively few precipitation records usually available for constructing the map can often be supplemented by snow-course data, and on occasion the pattern of vegetation in the more humid or higher altitude zones of the region may provide a clue to the probable amount of precipitation that occurs in a given altitude zone (p. 12).

The areal distribution of precipitation having been estimated, we next estimate the areal distribution of average annual runoff. As a preliminary step the basin should first be divided into contributing and noncontributing areas. Areas elevated above the basin floor contribute a part of their precipitation to the supply of recoverable water. Areas on the basin floor are generally considered to be noncontributing, in that the precipitation they receive is not recoverable; it is assumed that this precipitation is used to satisfy soil-moisture deficiencies on the basin floor and is subsequently transpired by native vegetation or evaporated from the bare ground. If there is an irrigation development on the valley floor it will be necessary to consider the possibility of some recharge to the ground-water body from rain falling on previously watered soil.

For the sake of simplicity in this discussion, it will be assumed that there is no reason to expect any yield from precipitation on the valley floor. In the absence of any readily recognizable evidence to the contrary, it would be assumed that there was no appreciable underflow into or out of the contributing area and that the entire yield of the contributing area was in the form of runoff. Subsequent computation of the hydrologic budget should reveal whether or not the assumption of no underflow was erroneous. Average annual runoff, or yield, would then be computed for the contributing area by a few of the methods described in the section of this report titled, "Water Yield." The average of the computed values of yield, weighted perhaps in accordance with the experience or judgment of the hydrologist, should produce a more reliable value of yield than any single computation.

The authors recommend using the Mundorff method (p. 44) for the first estimate of yield, and the channel-geometry method (p. 48) as a check on the results. Where feasible, any records of runoff that are available for the basin, or any miscellaneous measurements of discharge that have been made, should be correlated with concurrent records for nearby streams to provide an additional estimate of runoff. (See p. 37.)

A final check on the reasonableness of the computed value of mean annual runoff will be automatically made when a hydrologic budget for the basin is computed. After a final value of average annual runoff is decided on, the relation of runoff to precipitation, as used in the Mundorff method, is adjusted to give total runoff figures that agree with their final value and the areal distribution of runoff is thus established. In the event that man-made developments are present in the contributing area, suitable adjustments must be made for diversion, evaporation from stock ponds or reservoirs, or consumptive use by irrigated crops. Evaporation from stock ponds or reservoirs was discussed earlier in this report; consumptive use by irrigated crops may be computed by the Blaney-Criddle formula. Values of $K$ in that formula, for various crop types and localities, are given in a report by Blaney and Criddle (1962).

*Time distribution of precipitation and runoff.*--A within-year distribution of precipitation, should be computed based on records for the nearest precipitation station. A graph or table would be prepared for that purpose showing the average monthly distribution of precipitation in percent of the average annual precipitation. A time distribution of annual precipitation would be prepared, based on the assumption that the time distribution of annual precipitation in the study basin approximates that of the nearest long-term precipitation station, when annual values of precipitation for each location are expressed as percentages of their respective means. This assumption does not imply that there is any close relation between concurrent annual values of precipitation at the two locations. The distribution of annual precipitation would be shown on probability graph paper.

Similar presentations of the within-year distribution of runoff and the time distribution of annual runoff should be prepared. This would be done on the basis of the two precipitation distriubtions and the relation between zonal precipitation and water loss mentioned earlier. Greater refinement of the runoff values might be achieved in some areas where it is feasible to differentiate between winter and summer precipitation and to compute separate precipitation-runoff relations for each season.

*Flood frequency.*--A flood-frequency relation for the contributing part of the basin can be estimated from the regional relations discussed on page 61.

*Average annual hydrologic budget.*--Basically, any hydrologic budget states that inflow plus or minus change in storage equals outflow. In preparing the average annual hydrologic budget for an arid basin, the task can often be simplified by considering the contributing and noncontributing areas separately. The contributing area will be examined first. The storage item in the budget is usually negligible when the time interval considered is the water year. The inflow is the precipitation; the outflow is the water yield (adjusted for the effect of significant manmade development) plus natural water loss. The computation of water yield was described in the preceding paragraphs.

In considering the hydrologic budget for the noncontributing valley area, the inflow is the water yield from the contributing area. If we consider first a noncontributing area that has no man-made developments, this water yield is the only source of inflow. If the valley area is topographically closed and undrained, and if all the inflow apparently becomes subsurface water, the evapotranspiration by vegetation and evaporation from the bare soil are the only means of discharge or outflow. Methods of computing this type of discharge were discussed earlier in the section of the report titled "Evapotranspiration." The net change in ground-water storage in the course of years would probably be negligible. If the computed discharge is significantly greater than the computed inflow, there is the possibility that either or both of the computed values are in error, and there is also the possibility that ground water is moving into the basin beneath topographic boundaries, either through rock fissures or faults or as a result of difference between topographic and phreatic divides. All of these possibilities should be investigated and the values in the hydrologic budget should finally be adjusted to give consistent results.

If in this undeveloped basin the evapotranspiration discharge is significantly less than the yield, there is a possibility that storm runoff ponds on the desert floor and later evaporates. There may be evidence of this in the form of salt incrustations on objects above the ground surface, and the depth and volume of water involved may therefore be estimated. The water that seeps below ground either to be held as soil moisture or to reach a high water table could be estimated from a determination of the porosity (water-holding capacity) of the soil. The additional evaporation from the soil and ponded water would then be added to the evapotranspiration discharge in the hydrologic budget. The subject of ponded water on the desert floor was treated earlier in this report (p. 35-36). Another possibility, in the event that evapotranspiration discharge is significantly less than the inflow into a topographically closed basin, is that the assumption of an undrained basin is in error. There may be underground movement of water out of the basin. Several examples of interbasin underflow through Paleozoic carbonate rocks have been reported in Nevada (Eakin, 1961, 1966; Walker and Eakin, 1963).

So far we have been discussing a topographically closed basin with an undeveloped noncontributing valley area. If the valley area has some development--for example, irrigation activity--a few complications are added. Precipitation falling on watered cropland may make some contribution to the underlying ground-water body. Pumpage, the consumptive use of water by crops, and return flow are additional items that must be considered in the hydrologic budget.

If the noncontributing valley area is topographically open we have a drained basin. The outflow from the valley might occur naturally as streamflow, surface effluent such as springs or seeps, evapotranspiration, evaporation of water ponded on the valley floor, or as underflow. Several of these methods of outflow may operate concurrently. Under any constant stage of development that required less water than the average inflow to the valley, the net change in ground-water storage over a long period of years would probably be negligible. However, it would be necessary to evaluate the discharge represented by this use of water. Methods of measuring each of the forms of outflow were discussed earlier in this report.

*Source, occurrence, and movement of ground water.*--The arid and semiarid basins of the Southwest that require reconnaissance study are usually basins whose ground-water bodies are the most feasible sources for development of a water supply. Any surface water that might be diverted for use will generally be ground-water effluent at springs. Under those conditions the entire supply may be considered ground water. If, for the purpose of this discussion, we assume our basin to be of that type, the reconnaissance study should include a presentation of the salient features of the ground-water system as interpreted from the field investigation and computations that have been made. The source of recharge will usually be described as the higher-altitude precipitation, possibly some precipitation on the valley floor, and possibly some underflow. The areas where this underflow occurs will be delineated. The recharge areas--generally somewhat downstream from the apex of the alluvial fan and at constricted reaches of the ephemeral stream channel (Davis and De Wiest, 1966, p. 397, 425)--will likewise be delineated. That part of the valley fill that is the principal ground-water reservoir will be described. From water-level data and the location of hydrogeologic barriers, the quantity of water in storage will be estimated and the movement of ground water will be deduced, including underflow, if any. The location of springs will also be described.

*Perennial yield and potential development.*--Malmberg and Eakin (1962), in discussing the perennial yield at Sarcobatus Flat, Nev., make the following statements, which are applicable to most small ground-water basins in the Southwest where the concept of safe yield governs withdrawals.

> "The perennial yield of the water-bearing deposits in Sarcobatus Flat is the rate at which water can be pumped from wells year after year without decreasing the storage to the point where the rate becomes economically unfeasible or the rate becomes physically impossible to maintain. It is ultimately limited by the amount of water available to the system through natural and artificial recharge and infiltration of any excess irrigation and waste water.
>
> "The net amount of water that can be pumped perennially is limited to the total natural discharge and underflow that can be diverted to wells, and the amount of water that infiltrates to the ground-water reservoir that can be salvaged and is suitable for reuse.
>
> "In an estimate of perennial yield, consideration should be given to the effects that ground-water development by wells may have on the natural circulation in the ground-water system. Development by wells may or may not induce recharge in addition to that received under natural conditions. Part of the water discharged by wells may re-enter the ground-water reservoir by infiltration of excess irrigation water. Ground water discharged by wells usually is offset eventually by a reduction of the natural discharge. In practice, however, it is difficult to compensate fully for the discharge by wells by a decrease in the natural discharge, except when the water table has been lowered to a level that eliminates ground-water underflow from the basin and transpiration in the natural area of discharge. Ground-water underflow out of the drainage basin further complicates the evaluation of perennial yield."

As for potential development, it depends not only on the quantity of water available, but also on the chemical quality of the water. Where the supply of water of good chemical quality is limited, optimum development may often be brought about by the planned utilization of inferior water for secondary uses, such as irrigation, while the better water is reserved for human consumption. If the basin under investigation is a drained basin--that is, a basin hydraulically connected to another--plans for its development must take into consideration the effect of that development on the water supply of the downstream basin.

## Detailed Investigations

Detailed investigations differ from reconnaissance investigations not so much in scope as in intensity of study. The principal shortcoming of the reconnaissance study is the lack of refinement in the results obtained; the study provides a rough estimate only of the water-supply potential in the study area. This lack of precision can be overcome only by a long-term comprehensive observational program to obtain basic hydrologic data. The intensity of a detailed investigation will depend, of course, on the particular environment of the study area, the specific objective of the study, and the time and funds available. However, a generalized description of a detailed investigation will serve the purpose of integrating much of the material presented earlier. A description of that type follows under the headings listed below.

Hydrogeologic investigation
Precipitation-gage network
Stream-gaging network
Evaporation from reservoirs and stock ponds
Evapotranspiration where water table is low
Evapotranspiration where water table is high
Evaporation and evapotranspiration from large areas of shallow ponding
Data for studies of fluvial morphology
Analysis of data

*Hydrogeologic investigation.*--The geologic investigation in both reconnaissance and detailed studies has for its main purpose the determination of the location, extent and boundaries, and water-bearing characteristics of the hydrogeologic units. In the detailed investigation, however, more intensive effort goes into the collection of geologic data and into the construction of the geologic map, the cross sections, the fence diagrams, and the structure-contour maps needed for the refined study of the hydrogeology. The geology of arid lands is usually quite complex, but there are compensations. As Davis and De Wiest (1966) state,

> "Rock outcrops are fully exposed without the troublesome mantle of vegetation encountered in other regions, human habitations are sparse so that geophysical operations are not hampered by restrictions on the use of explosives or by metal pipes and the like that make resistivity surveys difficult, and points of natural discharge of potable water are clearly marked by vegetation. Conventional hydrogeologic methods of exploration are used in desert regions. Some modifications in operating procedures, however, are commonly needed because of greater depths to aquifers and because existing wells and bore holes are rarely sufficient in number to give the needed hydrologic and geologic control. More emphasis is given to seismic equipment coupled with extensive test drilling."

Geologic reconnaissance, together with a well canvass, will give a preliminary indication of where test drilling is needed. Continuing appraisal of water-level observations should indicate where additional test drilling is needed and where additions and changes in the network of observation wells are needed to provide water-level data that are representative of the regional water table or piezometric head. The intensity of coverage and frequency of water-level measurements will depend on the complexity of the hydrology and on the water problems in the area. The measurements should be made at a sufficient number of points and with sufficient frequency to insure the detection of significant changes in the hydraulic regimen.

Continuing estimates of the location, time, and volume of ground-water withdrawals, or pumpage, are necessary for a quantitative evaluation of ground-water conditions. As mentioned earlier, the records of total pumpage from a basin may be derived by one or more of the following methods: By summation of water-meter readings; from a relation between mechanical efficiency of pumps and total power consumption, coupled with a knowledge of the pumping lift; from a relation between total population and average use of water per capita; or from a relation between total production of some commodity and average use of water per unit of product.

Also mentioned previously was the need for determining the aquifer coefficients of permeability or transmissivity and storage or specific yield. Whether these coefficients are determined in the laboratory from samples in the field, by aquifer testing in the field, or estimated from well-log data, the results should be related to the geologic units shown on the maps and cross sections. Meyer (1962) has reported favorably on the use of a neutron moisture probe to determine the storage coefficient of an unconfined aquifer. In this method the neutron tube was used to determine the difference in moisture content of saturated material in a well and the moisture content of the same material after it had been drained following pumping.

Numerous wells and springs should be sampled for chemical analysis of the water. Each major aquifer system and subsystem should be sampled. This sampling is important not only for determining the suitability of the water for various uses, but it also provides clues to the interpretation of the ground-water hydrology. In developed areas periodic sampling is required to detect possible trends in the deterioration of the water. For example, where irrigation activities are concerned, the sampling should be adequate for studying the salt balance in ground-water reservoirs (Willets and McCullough, 1959).

*Precipitation-gage network.*--Because of the time limitation on reconnaissance studies, and because little is gained from a record of one or two years of precipitation or streamflow--a period which may or may not witness any events of hydrologic significance--precipitation or streamflow stations are usually not established for reconnaissance studies. Instead, the time and areal distribution of precipitation and runoff, as well as precipitation-runoff relations, are inferred from records for the nearest gaged basins that are adjudged to be hydrologically similar. In a detailed study, instrumentation would be provided to obtain such information more directly.

For a detailed study of the precipitation, a network of stations should be established to sample the precipitation at various altitudes and exposures. Because of the spotty character of summer storms, an extremely dense network of gages would be required to ensure sampling of each storm event. However, it will seldom be economically feasible to attain the required density, except in very small basins, and a compromise must invariably be made between what is desirable and what is feasible. Possibly two or three gages, depending on the range in altitude in the basin, should be installed on both the leeward and windward slopes of the watershed to gain a knowledge of the precipitation-altitude relation, and additional gages, as funds permit, should be installed elsewhere in the basin to increase the areal coverage. Some of the gages will be of the recording type and they should be installed at the site of each water-stage recorder in the contributing area of the basin. The stage and precipitation recorders would be operated by a single clock mechanism for synchronization of the records. One or more of the remaining precipitation gages would also be of the recording type.

In some basins, particularly those where snow is a factor in the hydrology, storage gages might be used at remote locations for increased areal coverage by the gage network, or snow courses might be established. After intense storms, "bucket surveys", which were described earlier (p. 17), might be used to obtain precipitation data to supplement those obtained from the gage network. If, by chance, the study basin is within the surveillance range of a Weather Bureau radar installation, supplementary storm data, at least of a qualitative nature, will be available. Aerial photographs might also be used to supplement the precipitation-gage network in delineating storm boundaries. If an area is flown shortly after rainfall has ceased there is usually a marked difference in color between wetted and unwetted soils. It is probable that even better results would be obtained by this method if infra-red photography or imagery were used.

The analysis of precipitation data for arid regions is similar to that for humid regions and is described in any standard hydrology text (for example, Linsley and others, 1949).

*Stream-gaging network.*--The stream-gaging network to be established will consist of two subnetworks--one for the contributing area of the basin and the other for the noncontributing area. The network for the contributing area will be considered first. Gaging stations will be established at, or upstream from, the mouths of the canyons from which the streams debouch onto the alluvial fans. The water yield is most easily and reliably measured when it is in the form of streamflow, and the purpose of establishing stream-gaging stations at sites where little streamflow seeps underground is to gage as much of the yield as possible. Often it will not be economically feasible to gage all channels that contribute significantly to the water yield; crest-stage gages that provide data for peak stages only should be established on those ungaged streams to supplement the gaging-station network. A significant part of the yield is often discharged at springs or seeps. These discharges should be measured with sufficient frequency to obtain a meaningful record of discharge over a period of years.

The stream-gaging network for the noncontributing area will consist of a limited number of gaging stations installed downstream from the gages at canyon mouths, so that changes in major flows from the mountains can be traced across the alluvial fans and valley floor. If there is a single, more-or-less permanent channel downstream from the canyon mouth, a conventional water-stage recorder will be used; if there are several distributary channels, it will not be feasible to instrument all of them with water-stage recorders. The principal channel only would be so instrumented; crest-stage gages on the other channels would be used to estimate the discharge hydrograph for each.

The difficulties associated with stream gaging in arid regions and some of the measures recommended to overcome them were treated in the section of this report titled, "Surface Water," and will not be discussed here.

In addition to the network of so-called hydrologic stream-gaging stations, it will also be necessary to measure or gage all significant surface-water diversions, as well as any significant channelized return flow to the streams. During periods of relatively constant streamflow, series of discharge measurements should be made at numerous sites along the stream to define losing and gaining channel reaches and to obtain quantitative information regarding the seepage.

A sampling program to measure sediment and chemical quality of surface waters at the gaging stations should also be established. Water temperatures will usually be measured only in perennial streams or in those that flow most of the year. Where sediment sampling is required at remote sites, consideration should be given to the use of automatic samplers. As for the sampling for chemical analysis, it should be coordinated with similar sampling of ground water because of the relation that often exists between ground and surface water. The chemical quality of diverted surface water and return flow should also be analyzed for use in studies of salt balance in the basin. Sampling should originally be done at frequent intervals to determine if chemical quality fluctuates repidly with time (Hem, 1948). If the fluctuation is rapid, it may be advisable to install an automatic continuous monitoring system. The subjects of sediment transport and chemical quality of water were discussed in some detail in earlier sections of this report.

*Evaporation from reservoirs and stock ponds.*--If reservoirs or stock ponds in the basin have a significant effect on the hydrology, it is necessary that evaporation loss be determined. If the evaporation figures desired are merely totals for the entire year or for the 6-month periods, May through October and November through April, and if the reservoir or stock pond has water at all times, the evaporation can be estimated rather simply within an accuracy of 15 to 20 percent. The only instrumentation needed is one or more evaporation pans installed near the reservoir, and a water-stage recorder to indicate lake levels so that the area of the open-water surface at any time is known. Regional pan coefficients for the entire year and coefficients for the period May through October are given by Kohler and others (1959, pls. 3 and 4). These pan coefficients are applied to the observed pan evaporation to give reservoir evaporation. Use of these pan coefficients for shorter periods of time will invariably give erroneous results.

If evaporation figures for a reservoir or large stock pond are required for short periods of time, the mass-transfer method of computing evaporation would be used, and in place of evaporation-pan measurements, we would require a continuous record of air temperature, water temperature at the surface of the stored water, relative humidity, and windspeed. Computation of evaporation by the mass-transfer method was explained earlier.

If the amount of seepage from the reservoir or stock pond is also to be determined, it is necessary that the inflow to and the outflow from the impoundment be measured so that a water budget can be computed. With evaporation known, seepage is the residual item in the water budget.

Use of the specialized instrumentation needed for the mass-transfer method is not warranted if evaporation and seepage are to be determined for small stock ponds that contain water intermittently. They would be estimated by use of the method of Langbein and others (1951) that is described on page 34.

*Evapotranspiration where water table is deep.*--Where the water table is low enough to be beyond the reach of phreatophytes, soil moisture is the sole source of water that is transpired by vegetation or evaporated from bare soil. If the arid area is unirrigated almost all moisture retained in the soil after the occurrence of a storm is subsequently consumed by evapotranspiration. Consequently, in the annual hydrologic budget for such an area, the net change in soil-moisture storage is approximately zero and annual evapotranspiration is the residual in the budget when all other items in the budget are measured or estimated. If a hydrologic budget is to be prepared for short periods of time it is necessary to monitor changes in soil moisture and this requires a network of sampling sites. At each site a soil-moisture tube is installed and periodic readings are made with a neutron probe. Increases in soil moisture represent the amount of water added by infiltrating rainfall or storm runoff; decreases in soil moisture represent the amount of water consumed by evapotranspiration. Field capacity of the soil at a sampling site can be determined by digging a shallow depression around the soil-moisture tube, filling the depression with water several times to saturate the underlying soil, covering the depression with an elastic sheet to inhibit evaporation while water drains from the soil, and determining the soil-moisture content two or three days later with a neutron probe. The soil-moisture content at that time is reasonably close to field capacity.

Numerous sampling sites are needed to study the soil-moisture regime of a basin. The heterogeneity of the soil material and our lack of complete understanding of the hydrodynamics of soil-water systems are complicating factors in designing a network of sampling sites. Usually the sites are selected to sample the various soil complexes in the basin, but in the absence of obvious differences in soil, sampling may be random or by elevation zone. After a period of operation, study of the hydrologic budget of the instrumented basin will usually indicate a method of weighting the soil moisture observations, and unneeded sites can be eliminated or additional sites can be added as needed.

If the evapotranspiration from irrigated areas is an item in the hydrologic budget, the water use can be estimated from the Blaney-Criddle formula, or determined more accurately by monitoring soil-moisture changes with neutron probes. A record of air temperature is required for application of the Blaney-Criddle formula.

*Evapotranspiration where water table is shallow.*--If the water table is fairly close to the land surface, ground water may be consumed by the transpiration of phreatophytes. Additional ground water may be consumed by evaporation from bare soil if the water table is very close to the land surface. In narrow canyons the water use by riparian vegetation can be estimated from diurnal fluctuations in the discharge hydrograph of base flow. For basins drained by streams with wide flood plains, the water use by phreatophytes can be estimated by a study of fluctuations of the water table in wells. The latter method of determining evapotranspiration also requires that the specific yield of the water-bearing material be known, along with the size of the area occupied by phreatophytes. It is generally desirable to compute the transpiration by more than one method, if this is feasible, to increase the reliability of the determination. Results of evapotranspiration-tank studies in nearby areas may be transferred to the study basin, if natural conditions are similar to those in the tank studies. A third method of estimating evapotranspiration involves the use of the Blaney-Criddle formula, as described on pages 25-27. That method requires that the species of phreatophyte be identified, the areal extent and density of growth be known, a record of air temperature be available, and depth to the water table be known. A guide for surveying phreatophytic vegetation has been prepared by Horton and others (1964). An observation well is needed for obtaining the depth to the water table; the depth cannot be deduced reliably from plant measurements (Robertson, 1962).

Evaporation from the bare soil is computed as a percentage of the evaporation that would occur from an open-water surface under similar climatic conditions. The open-water evaporation may be estimated from the evaporation map prepared by Kohler and others (1959, pl. 2), or preferably from evaporation-pan observations adjusted by an appropriate pan coefficient as given in plate 3 of the Kohler report. Additional information needed to determine the evaporation from bare soil includes the soil type, depth to the water table, and the area of bare soil. This information, when used with figure 7, gives the evaporation figures that are sought.

*Evaporation and evapotranspiration from large areas of shallow ponding.*--If it is known that storm runoff ponds on the desert floor, one or more crest-stage gages should be installed to indicate the extent and volume of the ponding. The approximate rate of evaporation would be deduced from periodic observations of the ponded water surface or from observations of the nearest evaporation pan. The results obtained by this procedure will not achieve a high degree of accuracy, but more elaborate instrumentation maintained over a period of years for this phase of the hydrologic study is generally not warranted. After the surface water is evaporated, the subsurface water added by the storm runoff becomes subject to evaporation or evapotranspiration. The amount of subsurface water available for evaporation or evapotranspiration depends on the hydrogeology of the basin. If the water table was originally high the addition of storm runoff would have raised the water table to the land surface and the additional evapotranspiration from the augmented ground-water body can be computed on the basis of figures 6 and 7, as explained earlier. If the water table was originally low the added subsurface water would be perched above a soil layer of low permeability. This would be indicated by any well on the desert floor that reaches the regional ground-water body. If the subsurface water is perched, its volume, or the amount available for evapotranspiration, can be determined by soil-moisture sampling with a neutron tube. Or, if the geological reconnaissance had indicated the location of impermeable layers near the land surface, ring infiltrometers driven into the impermeable layer would be used to determine the infiltration rate of the ponded water and the porosity (water-holding capacity) of the soil above this layer.

*Data for studies of fluvial morphology.*--Although studies of fluvial morphology are beyond the scope of a basic-data program, it is recommended that a reach of stream in the vicinity of each gaging station be mapped at the time the gage is established. Pre-selected cross sections, where bed material had been sampled, would be surveyed after each major runoff event to monitor changes that occur. Scour chains should be installed at these same cross sections for monitoring streambed scour and fill. The information obtained will be extremely helpful in studies of fluvial morphology that may later be made.

*Analysis of data.*--The additional data collected in the detailed study permits more sophisticated analysis of the basin hydrology than was possible in the reconnaissance study. For example, the geohydrologic information will be analyzed in basically the same manner as in the reconnaissance study to show the source, occurrence, and movement of ground water under conditions that exist at the time of the study. The availability of detailed data will result in an analysis of much greater accuracy, but in addition, with these data a model of the basin can be constructed to analyze the ground-water system under various conditions of water development. The model may be designed to study the effect of development on chemical quality of the water as well as on yield, thereby providing the information needed for optimum development of the basin. The model may utilize a digital computer (Remson and others, 1965), or the model may be of the analog type (Skibitzke, 1960; Brown, 1962).

The network of precipitation gages, after a period of years whose length depends on the frequency of hydrologic events, will provide valuable data that were not available for the reconnaissance study. These data, when supplemented by those for the sparse network of surrounding Weather Bureau stations, will provide an areal and time distribution of precipitation, which though generalized, is still better than that provided by the surrounding stations alone. For example, data will be available for defining the altitude-precipitation relation in the basin. Admittedly, the precipitation phase of the investigation will be weak; the vagaries of convective precipitation present a complication that makes satisfactory analysis of the precipitation regime difficult to attain.

Where the reconnaissance study provided crude estimates of water yield based on the estimated precipitation and on various climatic and physiographic characteristics, the stream-gaging network provides the data for defining precipitation-runoff relations for storm periods or periods of longer duration. These relations, used with the precipitation data, will give the time and areal distribution of runoff, including the flood distribution, or flood-frequency relation.

The benefit to the study accruing from the availability of detailed information concerning such hydrologic elements as soil moisture, storage in the snowpack, evapotranspiration, evaporation from open-water surfaces, chemical quality of water, and sediment transport, is self-evident. Among the advantages is the opportunity for study of the general hydrology using modeling techniques such as that of Crawford and Linsley (1966).

The data obtained will now be considered in relation to the three objectives of a basic data program, as listed under "Purpose and Scope" of this report. The first objective is a description of the hydrologic regimen of the basin involving the time and areal distribution of the various hydrologic elements; the second objective is a general inventory of the water resources involving hydrologic budgets for long- and short-term periods. It is apparent from the foregoing discussions that the detailed investigation fulfills these two objectives. The third objective--one that has not been discussed--is the collection of sufficient information for general study of both interrelations among various hydrologic elements, and relations of various hydrologic elements to the environment. The following incomplete list of relations that can be studied with the data collected indicates that this objective has been met.

Interrelations among the various hydrologic elements

1. Relation of long- or short-term runoff to precipitation
2. Relation of dissolved solids to streamflow
3. Relation of base flow or underflow to ground-water levels
4. Relation of base-flow recession characteristics to aquifer coefficients
5. Quantitative relations among dissolved constituents in surface or ground waters

Relations of various hydrologic elements to the environment

1. Relation of streamflow variability to geologic parameters
2. Relation of streamflow characteristics to physiographic and climatic parameters
3. Relation of aquifer characteristics to geologic parameters
4. Relation of sediment production to basin and climatic parameters
5. Relation of chemical quality to geologic parameters

## SUMMARY AND CONCLUSIONS

Current methods of collecting and analyzing the data required for a study of the basic hydrology of arid regions were summarized and discussed. It was concluded that despite the great variability of the many hydrologic elements, present methods of data collection and analysis will generally give results of acceptable reliability, but at a cost in time and money that far exceeds that for humid areas. Some uncertainty exists, however, concerning certain aspects of the hydrology, particularly in connection with convective precipitation and ground-water recharge through alluvial fans. Where convective rainfall is concerned, relations of intensity, duration, area, and frequency defy precise analysis and highly generalized relations must be used. For example, individual convective storm cells usually cover very small areas; we have no way of determining the probability of such storms developing over a multi-tributary stream system in such areal and time patterns as to produce large volumes and peak rates of runoff. It is even extremely difficult to determine the areal distribution of convective precipitation over an instrumented area, because it generally is not economically feasible to maintain a network of precipitation gages of the required density. As for ground-water recharge from storm runoff seeping into an alluvial fan, serious questions arise as to the amount of this recharge and the location of the recharge areas. These are not academic questions. If much of the seepage is held as soil moisture for subsequent evaporation, optimum management of the water resource might require that surface flows be carried across the fan in conduits to the place of use. The methodology for investigating the seepage problem is known, but because of the complex heterogeneous composition of most fans the cost of a complete exploration is usually prohibitive.

This report was intended to be comprehensive in its coverage of the various aspects of basic hydrologic investigations in arid regions. However, it was not possible to cover every ramification of the subject. Hopefully, the report will provide guidelines for investigational procedures and suggest approaches to unusual problems connected with basic studies. It should be emphasized that close observation of environmental factors can be of material aid in studying the hydrology of arid regions and may partially offset the general deficiency of direct hydrologic data.

An urgent problem, untouched in this study, concerns the need for guidelines to define the minimum level of data collection that is consistent with: (1) the particular environment of a study area, (2) the purpose of investigating the area, and (3) the immediacy of need for the results of the investigation. It is recommended that study of such guidelines be undertaken as soon as practicable.

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