A Hydrogeologic Study of the Ground-Water Reservoirs Contributing Base Runoff To Four Mile Creek East-Central Iowa

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GEOLOGICAL SURVEY WATER-SUPPLY PAPER 1839-O

Prepared in cooperation with the Iowa Geological Survey

A Hydrogeologic Study of the Ground-Water Reservoirs Contributing Base Runoff To Four Mile Creek East-Central Iowa

By GEORGE R. KUNKLE

CONTRIBUTIONS TO THE HYDROLOGY OF THE UNITED STATES

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A HYDROGEOLOGIC STUDY OF THE GROUND-WATER RESERVOIRS CONTRIBUTING BASE RUNOFF TO FOUR MILE CREEK, EAST-CENTRAL IOWA

By George R. Kunkle

ABSTRACT

Four Mile Creek basin is an agricultural watershed in east-central Iowa. The base flow of the stream comes from a system of two interconnected ground-water reservoirs, one composed of upland loess and the other of bottom-land alluvial sand. Together, the two reservoirs occupy the entire watershed of 19.5 square miles upstream from the principal gaging station.

The loess is a water-table reservoir, 5–12 feet in saturated thickness, which drains either to tributaries heading in the uplands or downslope to the alluvial sand reservoir. The alluvial sand is a semiartesian reservoir averaging 15 feet in thickness; it underlies the alluvial silts and clays of the modern flood plain and drains to the stream.

Both reservoirs leak to an underlying till. The till is a bouldery clay loam deposit ranging in thickness from 170 to 360 feet. Leakage to the till is estimated to be about 1.8 inches annually. Water in the till moves downward into Middle Devonian limestones and is lost from the basin. The limestones are incised by deep bedrock channels infilled by permeable sands. Ground-water movement in the limestones is toward the bedrock channels which act as underground drains.

During the 2 water years of record, 1963 and 1964, recharge to the loess reservoir averaged 2.1 inches, and discharge as base flow averaged 0.2 inch. Most of the recharge, about 1.4 inches per year, was eventually lost from the basin as vertical leakage to the deeply buried bedrock. Although the base-flow contribution from the loess averaged only 10–15 percent of the annual base flow, the loess contributed as much as 35 percent during the late fall and winter but none during the summer.

The alluvial sand reservoir received an average of about 3.2 inches of recharge annually. A small amount of recharge came from the loess reservoir; the remainder came from precipitation recharge and bank storage inflow. Quantitatively, precipitation recharge and bank storage inflow could not be determined separately, although ranges may be placed on each component. For example, during 1963 the bank storage component of inflow had a range of 0–1.7 inches, and recharge by precipitation was 1.6–3.3 inches.

Discharge from the alluvial sand reservoir averaged 1.4 inches annually. Because of the intimate connection between the alluvial sand and the stream, bank-storage outflow was significant, accounting for 37–42 percent of the annual base flow. It was found that the rate of bank-storage recession could be predicted from the aquifer characteristics; however, the amount of bank-storage inflow or outflow could not be theoretically determined.

The base flow characteristics of Four Mile Creek are related principally to the alluvial sand reservoir. Because of the wide range of aquifer characteristics possible for reservoirs of this type, reservoir coefficients, and hence, base-flow characteristics will vary widely from basin to basin.

INTRODUCTION

PURPOSE AND SCOPE

An important factor in evaluating the water resources of a drainage basin is a knowledge of the magnitude, duration, and frequency of base flow, or runoff derived from ground-water reservoirs. In areas where base-flow data are incomplete, definition of the base-flow characteristics by extrapolation from basins with record is uncertain. Much of the uncertainty centers around problems such as defining the sources of ground-water runoff, the mechanics of ground-water discharge to streams, and the extent to which storage within a single ground-water reservoir might affect the base flow of a stream receiving discharge from more than one reservoir.

The purpose of this report is to describe the hydrogeologic characteristics of the ground-water reservoirs discharging to Four Mile Creek and the quantitative manner in which those reservoirs operate to control the base-flow regimen of the stream.

Initially, the scope of the study was limited to a 3-mile downstream reach of Four Mile Creek. Gaging stations were established at each end of the reach. Later, information collected from the study reach was used to analyze the stream regimen within the entire basin.

The descriptions of the base-flow regimen should have considerable carryover value because Four Mile Creek is representative of many small streams in east-central and southern Iowa, and has factors in common with many streams throughout the humid eastern part of the United States.

ACKNOWLEDGMENTS

The investigation described in this report was under the general supervision of R. W. Carter, chief, research section, Surface Water Branch, U.S. Geological Survey, succeeded by R. A. Baltzer, acting chief, research section.

Many other individuals or groups contributed to the project. Messrs. V. R. Bennion and S. W. Wiitala, district engineers, Surface Water Branch, U.S. Geological Survey, Iowa City, provided administrative supervision of the project. Mr. E. C. Pogge, hydraulic engineer, U.S. Geological Survey, Iowa City, worked jointly with the author on almost all phases of the field studies and was solely responsible for the installation and rating of the stream gages. Mr. L. E. Betts, hydraulic engineering technician, U.S. Geological Survey, Iowa City, maintained the stream gages, ground-water observation wells, and rain tained the stream gages, ~~ground-water observation wells, and rain~~ U.S. Geological Survey, Iowa City, provided valuable assistance and guidance. The Iowa Geological Survey under H. G. Hershey, State Geologist, supplied most of the records used in the regional aspects of this study; Mr. P. J. Waite, Iowa State Climatologist, Des Moines, assisted in the selection of rain-gage sites, supplied four nonrecording rain gages, and advised in the analysis of the rainfall records; Dr. R. L. Morris, Assistant Director, Iowa State Hygienic Laboratory, Iowa City, provided assistance in the analysis of the chemical characteristics of ground and surface waters; Dr. R. V. Ruhe, research geologist, U.S. Department of Agriculture, Ames, assisted and guided the author in geologic studies of the Pleistocene stratigraphy of the study area and provided valuable borehole and soils information; Messrs. M. I. Rorabaugh, research engineer, U.S. Geological Survey, Tacoma, Wash., H. H. Cooper, research engineer, and R. C. Riggs, hydraulic engineer, U.S. Geological Survey, Washington, D.C., provided technical assistance.

To all these individuals or groups, the author expresses his sincere appreciation.

GEOGRAPHY

LOCATION AND EXTENT OF AREA

Four Mile Creek basin is in northwestern Tama County, east-central Iowa (fig. 1). The study reach has an approximate lat of 42°12′ and a long of 92°35′. The extent of the basin draining to the study reach and the location of the inflow and outflow gaging stations are shown in plate 1.

The upstream gaging stations, Four Mile Creek near Lincoln and Half Mile Creek near Gladbrook, have upstream basins of 13.78 and 1.33 square miles, respectively; the downstream station, Four Mile Creek near Traer, has a drainage area of 19.51 square miles. Between the three stations there is 4.40 square miles draining to the study reach.

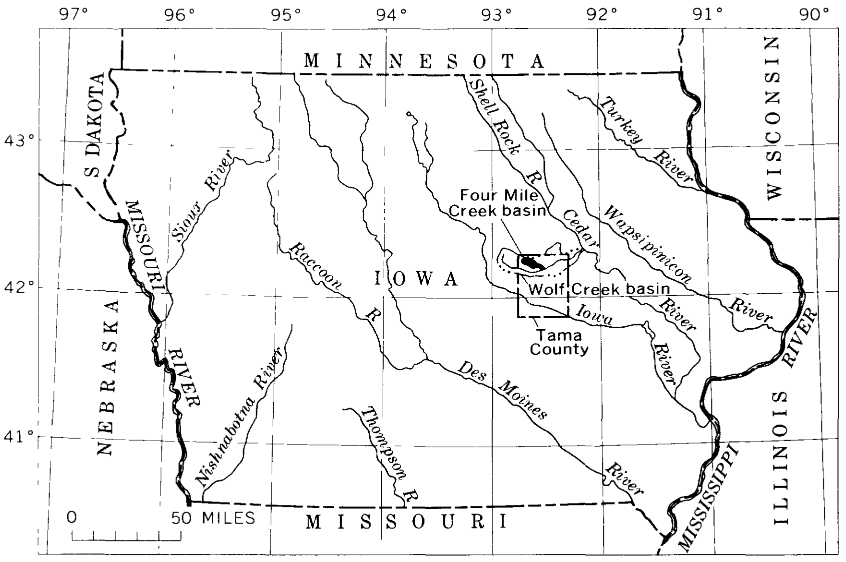

FIGURE 1.—Location of Four Mile Creek and Wolf Creek basins.

CULTURE

The study reach along Four Mile Creek is extensively farmed as is the entire basin. The principal crops are corn, beans, and alfalfa. Slightly less than one-third of the reach is in permanent pasture, and a small part of the upland areas is in timber. During the study period there were no significant changes in land use.

CLIMATE

Iowa has a humid-temperate climate subject to a wide variety of weather conditions with rapid day-to-day changes. The summers are hot and humid, and the winters are cold and damp. Thunderstorms account for most of the rainfall during the summer and for the highly variable streamflow during that season.

The 30-year (1935–64) average annual temperature at Grundy Center (17 miles northwest of the reach) is 47.6°F. Figure 2 shows the average maximum and minimum monthly temperatures at Grundy Center for the 1963 and 1964 water years—the 2 years during which hydrologic data were collected for this report.

The average length of the growing season is 154 days, generally from about May 15 to about October 15.

The 18-year (1947–64) mean annual precipitation at Traer (8 miles east of the study reach) was 32.4 inches; however, the monthly distribution of precipitation for any 1 year is quite variable. Figure 3 shows monthly precipitation for the two water years of study. Annual precipitation for 1963 was 30.97 inches and for 1964, 31.81 inches, which was slightly less than the mean annual precipitation during the period of record.

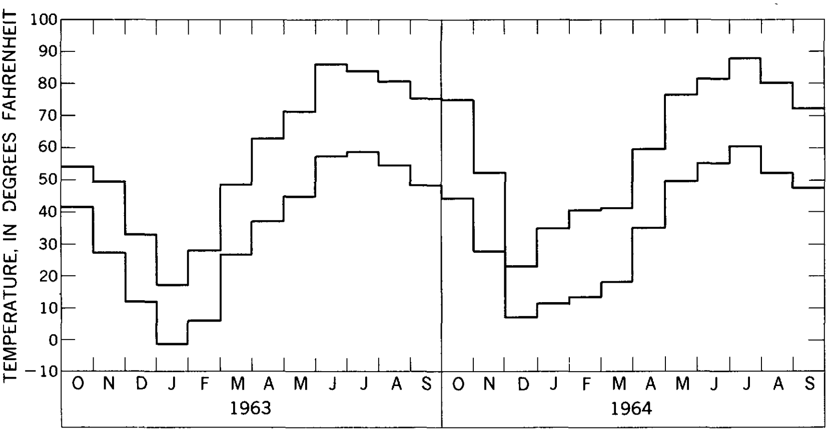

FIGURE 2.—Monthly average maximum and minimum temperatures at Grundy Center, 1963 and 1964 water years.

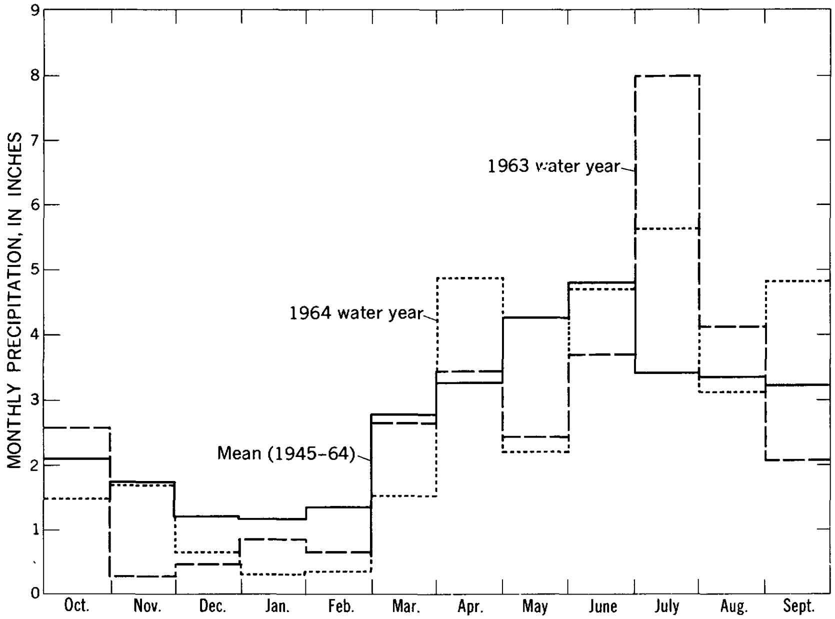

FIGURE 3.—Monthly distribution of precipitation at Traer, 1963 and 1964 water years.

PLEISTOCENE HISTORY AND GEOMORPHIC DEVELOPMENT

The present topographic features and drainage of the area relate to glaciation during the Pleistocene and subsequent modification during Recent time. Glacial deposits mantle the bedrock in the study reach along Four Mile Creek to a maximum thickness of 360 feet. For the most part, till and an overlying mantle of loess compose the drift. Along the drainage divides, the loess is commonly 30 feet or more thick but thins along the valley sides.

From drill and core samples, two of the major glacial stages are recognized, the Nebraskan and the Kansan. Both are documented by buried soil profiles (paleosols) termed gumbotils by Kay and Apfel (1929, p. 139–179). Logs of holes drilled on the drainage divide in the Four Mile Creek area show a stratigraphic section composed, in descending order, of about 30 feet of loess, the Kansan gumbotil, a relatively thin Kansan till, a Nebraskan gumbotil, and a relatively thick section of Nebraskan till (see figure 4). The altitude of the Kansan gumbotil is approximately 1,020–1,030 feet, and the altitude of the Nebraskan gumbotil, 1,000–1,015 feet.

Another stratigraphic break is recorded at an altitude between 840 and 900 feet, as indicated by organic silt, wood chips, charcoal, leached silt and clay, and sand and gravel. The break appears to be correlative

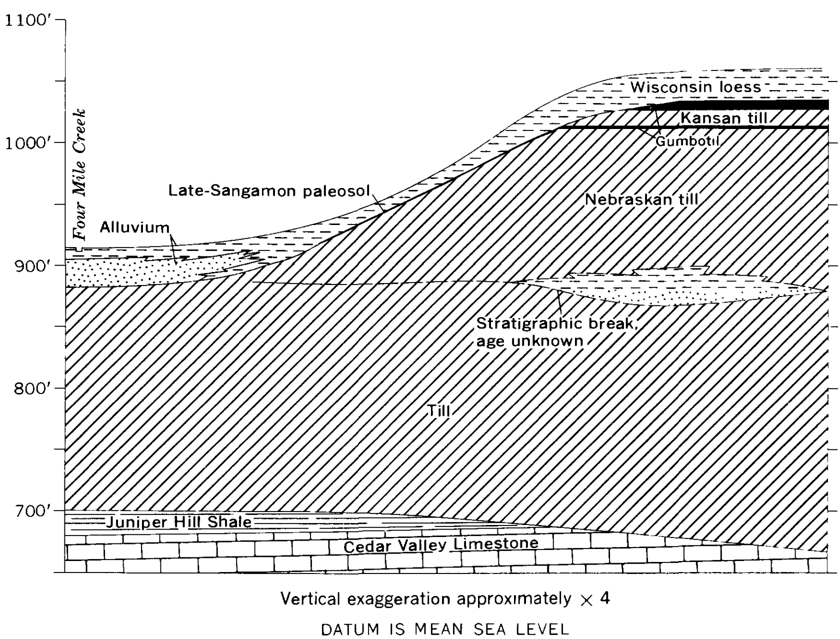

FIGURE 4.—Generalized geology of the Pleistocene deposits.

with one recorded in southern Iowa (W. Steinhilber, oral commun., U.S. Geol. Survey, Iowa City), although its stratigraphic age and significance have not yet been determined.

Erosion, deposition, and soil development occurred in the area from the end of the Kansan Glaciation until Wisconsin time. Both Kansan gumbotil, Yarmouth to Sangamon in age, and late Sangamon paleosol formed during this interval. Because of the general accordance of altitude of the Kansan gumbotil, it is inferred that the Yarmouth to Sangamon surface was much flatter than the present landscape. Erosion dissected the flat surface, and development of the late Sangamon paleosol occurred subsequently.

The late Sangamon paleosol occupies a side-slope position which truncates Nebraskan and Kansan gumbotils. The erosion immediately preceding the development of the late Sangamon paleosol probably represents the inception of the Four Mile Creek drainage system.

During early Wisconsin time the topography of northeastern Iowa was extensively modified and resulted in a surface considerably flatter than the topography of southern Iowa. Until recently, the flat landscape, termed Iowan topography, was attributed to any early Wisconsin glacial advance, the Iowan stade. Ruhe, Dietz, Fenton, and Hall (1965, p. 11) rule out an Iowan ice advance because they find no Iowan till mantling the Iowan topography. Instead, they attribute the featureless landscape to fluvial erosion.

Iowan topography characterizes the upstream part of Four Mile Creek basin. In contrast, mature topography with sharp divides characterizes the lower part including the study reach. This topographic discontinuity is not reflected in the Pleistocene stratigraphy; in fact, the Pleistocene stratigraphy of the study reach persists throughout the Wolf Creek drainage basin, about 300 square miles.

Loess deposits, as much as 35 feet in thickness, document the late Wisconsin Glaciation. Coincident with loess deposition, stream erosion cut into and removed parts of the late Sangamon paleosol and older deposits. The occurrence of simultaneous deposition and erosion is responsible for thinner increments of loess on the side slopes than on the ridge tops (Ruhe and others, 1965, p. 15).

Recent erosion and deposition account for alluvium found in the present stream valley. The alluvium consists of an average of 15 feet of sand overlain by an average of 12 feet of silt and clay. Since farming began here, as much as 2 feet of sediment has been deposited on the modern flood plain.

Present relief in the reach is a maximum of 155 feet (pl. 1). Tributaries are well developed, and the drainage pattern is mature. The northward-facing slopes are the steeper as the creek impinges along the south flank, a common characteristic of east-west flowing streams in Iowa. The valley bottom is fairly broad, averaging about one-quarter of a mile in width. Several coalescing levels of old and modern alluvial fans compose the microtopography of the valley bottom and merge gradually with the side slopes. The simultaneous deposition of loess and erosion of the valley have combined to smooth and round the topography in the basin.

The stream is incised into its flood plain. Almost vertical banks, as much as 6 feet in height, are cut into alluvial silts and clays. The streambed is relatively flat with only a few minor nick points along the stream profile.

Present-day erosion in the area consists of sheet flow along side slopes, headward erosion in tributaries, and mass wasting of the main-stem banks into the creek. Integration of the natural drainage is aided by field draintile in side-slope draws and across the relatively flat bottom lands. No record is available of the amount, location, depth, or size of the draintile installed.

GROUND-WATER FLOW SYSTEMS

Ground-water movement in east-central Iowa may be divided into a regional flow system and many small local flow systems. The local flow system is characterized by recharge, movement, and discharge to local streams within the confines of individual surface drainage basins. In local systems, the ground-water divides closely approximate the surface-water or topographic divides. In the regional flow system, there is very little correlation between ground-water and surface-water divides. Recharge in one drainage basin may ultimately discharge in another drainage basin many miles away. Generally speaking, water in the regional ground-water flow system discharges to the major rivers of east-central Iowa (the Iowa and Cedar Rivers) whereas water in the local flow systems discharges to the tributaries in addition to the major rivers.

REGIONAL FLOW SYSTEM, WOLF CREEK BASIN

Although the regional ground-water flow system does not contribute base runoff to Four Mile Creek, a brief description of this system will aid in understanding the mechanics of the local flow system.

Regional ground-water flow in the Wolf Creek basin begins as vertical leakage through thick Pleistocene tills, and locally shales, to underlying limestone aquifers. These aquifers, Silurian, Middle Devonian, and Mississippian in age, are cut into by deep buried bedrock channels. The channels are infilled with permeable sands and have a discharge outlet considerably lower than the level of the local streams. Consequently, water in the limestones moves into the channels which divert the flow to outlets at great distances from the Wolf Creek basin. In effect, the bedrock channels act as large underground drains preventing discharge from the bedrock to local streams.

An exception to the regional flow pattern is found in a 5-mile reach of Wolf Creek approximately 8 miles upstream from the confluence of Wolf Creek with the Cedar River (fig. 1). At this site the contribution from the limestone bedrock is computed to be about 3.2 cfs (cubic feet per second) (Kunkle, 1965, p. D209). From a piezometric map of the bedrock, the discharge to this reach of Wolf Creek is found to be supported by recharge from a 23-square mile area. Dividing the discharge by the area of recharge gives a rate of leakage to the regional flow system of 0.0004 foot per day. The errors involved in computing this value cannot be accurately determined but are estimated to be in the range ±20 percent. Because no better estimate of the leakage rate is available, the value of 0.0004 foot per day is assumed to be correct.

From hydrographs of wells in the drift and the bedrock, it can be demonstrated that there is very little variation in the vertical hydraulic gradient with time. Therefore, the leakage rate determined may also be considered a constant.

LOCAL FLOW SYSTEM, FOUR MILE CREEK BASIN

Superimposed on the regional ground-water flow system within the bedrock are many local flow systems within the drift. Each of the local flow systems is confined to a particular surface drainage basin and contributes base runoff to the stream occupying that basin.

Four Mile Creek basin is hydrologically typical of the many local flow systems in the Wolf Creek basin. The following description of Four Mile Creek basin includes a discussion of the geologic framework in which the flow system operates, the hydrologic properties of the geologic materials, estimates of the rates and amounts of ground-water movement, and a hydrologic budget of the local flow system for the 2 water years of investigation.

Three geologic units compose the framework of the local flow system. These are the till and associated paleosols, the loess and associated loesslike deposits, and the alluvial sands. The role of the till and its associated paleosols is that of an aquitard. The loess and the alluvial sands constitute the two ground-water reservoirs contributing to runoff. Ground-water movement in the local flow system is diagramed in figure 5.

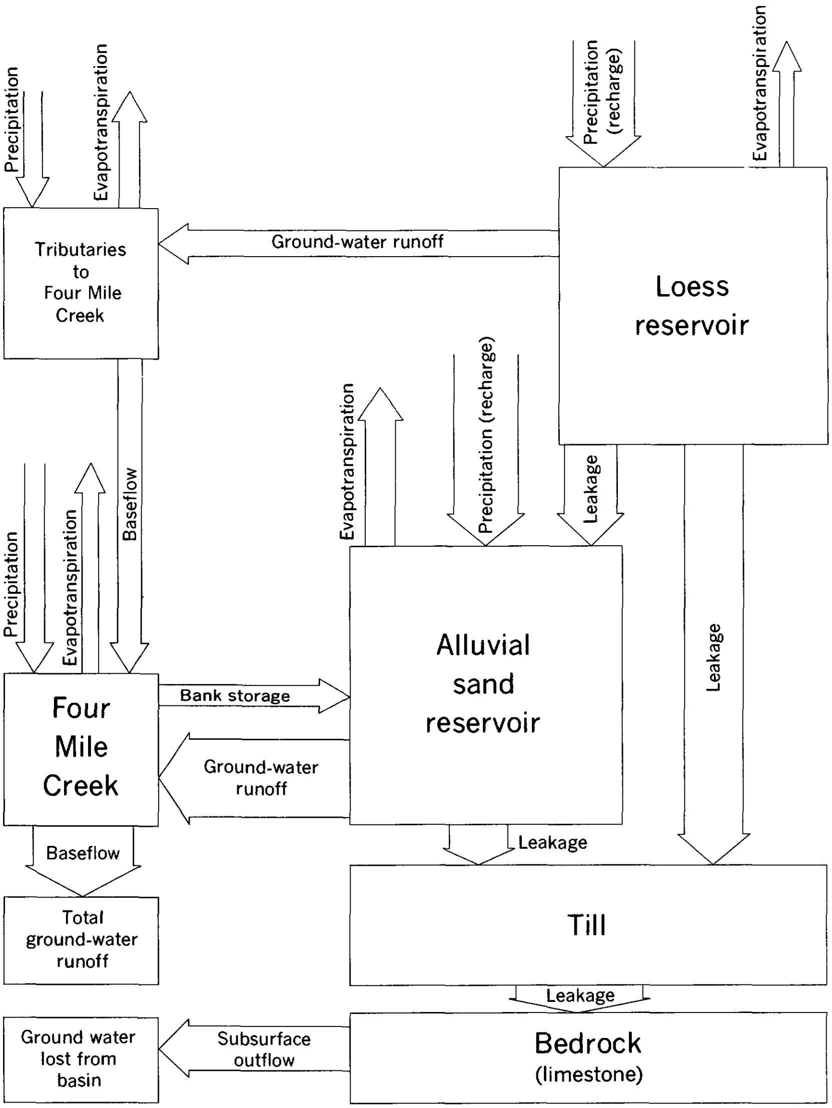

FIGURE 5.—Ground-water movement.

HYDROGEOLOGIC UNITS

TILL

Although the till does not contribute a measureable amount to local ground-water runoff, the conditions under which the till transmits leakage to the bedrock constitute one of the important boundary conditions of the local flow system. The till in Four Mile Creek basin is 170–360 feet thick and is fairly uniform loam to clay loam in texture. Many sand and gravel pockets occur in the till but for the most part do not interconnect and have little influence on the hydrologic behavior of the till. As indicated previously, the upper part of the till contains three paleosols; the Nebraskan and Kansan gumbotils, and the late Sangamon paleosol (fig. 4). These paleosols constitute the remains of the illuviation zone of old soils, and consequently have higher percentages of clay and a lower permeability than the till itself.

Cores taken at a depth of 58 feet below the drainage-divide surface indicate that the till is unsaturated beneath the gumbotils. Moisture content in the cores averaged 33 percent by weight and porosites between 40 and 47 percent. The unsaturated zone exists despite the presence of saturated loess above the gumbotil. Perched water in the loess is attributed to the relatively impermeable nature of gumbotil. The thickness of the unsaturated zone is not precisely known.

Along Four Mile Creek, and beneath the alluvium, the paleosols are absent, and the till section is completely saturated. Wells A–5, –6, –7 and F–2, –3, –4, show a uniform decrease in head with depth equal to 0.21 foot per foot. If the section were not saturated or if the till did not have uniform hydraulic characteristics, the constant change in potential with depth would not exist. Changes in the vertical hydraulic gradient with time are not more than 10 percent because stage changes in the overlying alluvium and underlying limestone are small compared to the saturated thickness of the till.

Vertical leakage through the till was estimated to average 0.0004 foot per day (p. O9) for an area downstream from the confluence of Four Mile and Wolf Creeks. The estimate is believed applicable to Four Mile Creek basin because of close similarities between the geology (Ruhe and others, 1965, p. 19–21) and the hydrologic conditions of the two sites.

It is important to recognize that the leakage rate varies areally. For example, beneath the flood plain of Four Mile Creek the paleosols are absent, the complete till section is saturated, and the measured vertical gradient is 0.21 foot per foot. However, beneath the drainage divides, the paleosols are present, retard ground-water recharge to the till, and the average vertical hydraulic gradient is somewhat greater, 0.35 foot per foot.

An estimate of the vertical permeability of the till may be obtained by using the area underlain and that not underlain by paleosol as weighting factors in the following equation.

$$0.70(P'I_u)+0.30(P'I_s)=0.0004 \text{ foot per day,}$$

where

P' = vertical permeability of the saturated till in feet per day'

I_u = vertical hydraulic gradient in the till underlying the divides, and

I_s = vertical hydraulic gradient in the till underlying the bottom lands.

Substitution in the above equation gives an estimated vertical permeability of 0.0013 foot per day for the till. The leakage rate through the incompletely saturated upland till is PI_u, or 0.0005 foot per day; for the completely saturated bottom land till, $P'I_s$ is 0.0003 foot per day. This solution gives a slightly greater leakage rate for the divide areas despite the presence of paleosols there retarding ground-water recharge to the till.

An indication of the accuracy of the computed leakage rates and vertical permeability is obtained by estimating the leakage rate through the unsaturated till zone. The permeability of the unsaturated till is obtained from an empirical relation between the ratio of unsaturated to saturated permeability and the degree of saturation as determined by Irmay (Todd, 1959, p. 72). For the till cores, saturation is about 76 percent, and the permeability of the unsaturated till is estimated from the relation to be 0.31(0.0013) equal to 0.0004 foot per day. For a hydraulic gradient of unity, the leakage rate is also 0.0004 foot per day. Considering the assumptions which are made in both computations, the agreement is good, and the estimated upland and bottom-land leakage rates are taken to be reasonably correct.

THE LOESS RESERVOIR

As shown in plate 1, loess or loesslike colluvium and alluvium occupy most of the reach area. The boundary between the loess and loess-derived colluvium is transitional and is arbitrarily defined by the boundary between upland and bottom-land soils. Likewise, the boundary between colluvium and alluvium is transitional. For the purpose of this report, the loess and loesslike colluvium and alluvium are grouped together because of their similar hydrogeologic properties and are referred to as the loess reservoir.

The loess reservoir occupies 3.15 square miles of the 4.40 square miles drained by the reach (pl. 1). Furthermore, the area of the loess reservoir can be divided into small subbasins which drain to tributaries of Four Mile Creek, and side-slope areas where subsurface drainage has no surface outlet and ground water drains into the alluvial sands. The area of the reservoir contributing to tributary streamflow is 1.45 square miles, whereas the area contributing leakage to the alluvium is 1.70 square miles.

The importance of the loess as a ground-water reservoir was not realized at the time observation wells were installed. Later, it was discovered that some of the observation wells completed below the loess leaked along the annulus, and their water levels responded to hydrologic conditions in the loess. When the hydrographs for the leaky wells were matched with hydrographs of piezometers in the loess, it was determined that well E–1, a recording leaky well, 30 feet deep and completed just beneath the loess, produced a hydrograph representative of storage conditions in the loess reservoir (fig. 6). For the purposes of this report, well E–1 is used exclusively as an index to storage conditions in this reservoir.

Water levels in the loess reservoir respond to rainfall and snowmelt, but they lag in time compared with water levels in the alluvial sand. Storage in the reservoir shows a strong seasonal variation. The reservoir is recharged in early spring, but water-level highs are not reached until early summer. Despite depletion of soil moisture during the growing season, rains in late summer and early fall, if of sufficient magnitude, can cause appreciable recharge. Water-level lows occur in the winter, generally after a long autumn recession.

Examination of the daily stage records of well E–1 shows no diurnal fluctuations indicative of evapotranspiration. The 10–20 feet of unsaturated loess overlying the reservoir probably precludes the loss of significant quantities of ground-water storage by evapotranspiration through vertical movement to the root zone or land surface. In contrast, tributaries draining the loess show strong diurnal changes in discharge during the growing season. From this and the fact that the tributaries are lined with phreatophytes and, during base-flow conditions, become dry before reaching Four Mile Creek, it is concluded that most of the evapotranspiration losses from the reservoir occur in the riparian zone of the tributaries. Similar observations on the importance of riparian vegetation as related to base flow in small basins are reported by Dunford and Fletcher (1947).

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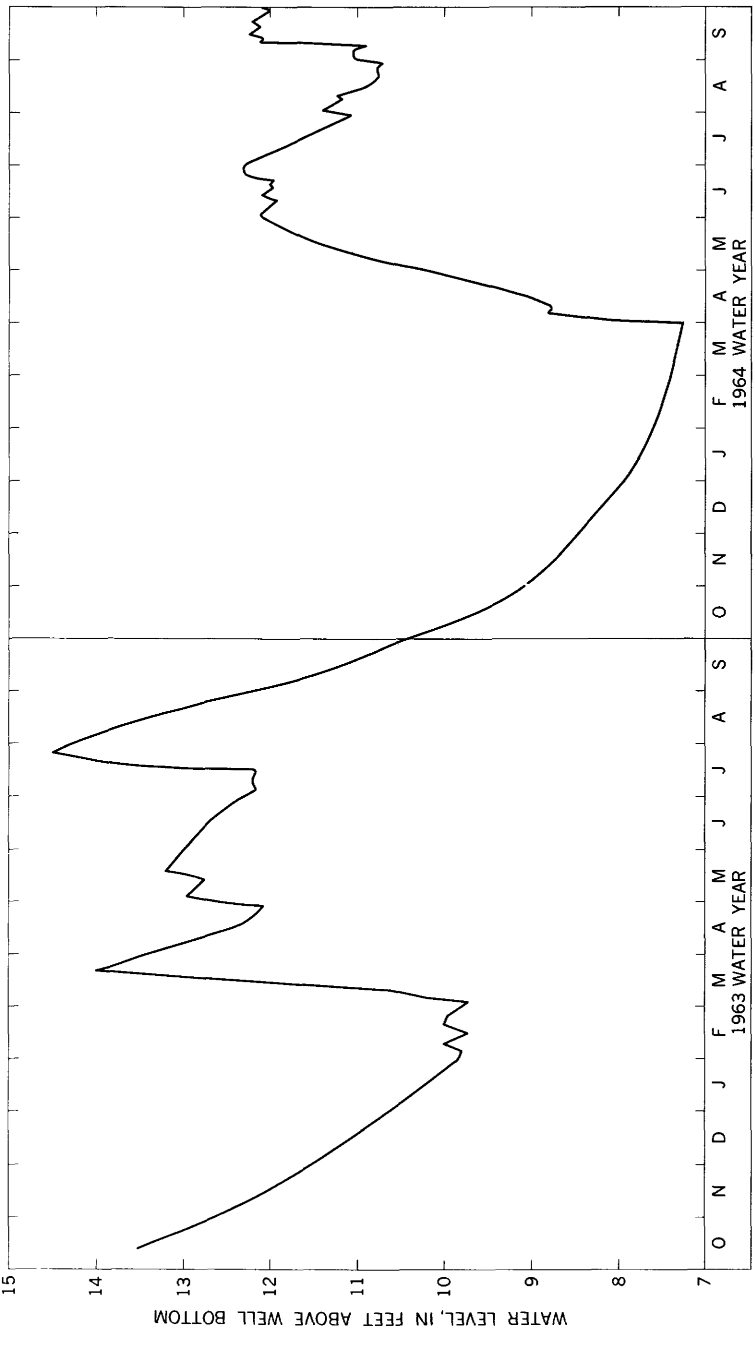

FIGURE 6.—Storage conditions in the loess reservoir as shown by well E–1.

Because of the irregular shape of the loess reservoir and the many discharge outlets from it, the only practical method of working hydrologically with the reservoir is by use of the storage depletion equation,

$$Q = A(\Delta h / \Delta t) S', \tag{1}$$

where

Q=average discharge,

A=drainage area of the reservoir,

$\Delta h / \Delta t$=average rate of water level recession, and

S'=effective yield of the reservoir.

Solution of equation 1 requires that the effective yield of the loess reservoir be determined. As used in this report, effective yield, S', is equal to the measurable amount of water discharged from a reservoir to a particular outlet divided by the volume of reservoir dewatered during the discharge period. Where the measurable discharge is equal to the total discharge, the effective yield is a maximum and is equal to the storage coefficient, S, or specific yield for the water-table reservoir.

The effective yield of the loess to the tributary streams was determined by measuring the discharge of a tributary draining the loess and the amount of water-level recession in the reservoir. The tributary basin selected for this determination is the 0.085-square-mile basin above the low-flow station shown on plate 1. Loess and loess-derived colluvium mantle the entire basin.

In early November 1963, 10 days after the first killing frost, volumetric discharge measurements made at the low-flow station averaged 0.0018 cfs, while the corresponding water-level recession in well E–1 wos 0.025 foot per day. Correcting for units and substituting in equation 1, the effective yield of the loess reservoir is computed to be 0.003.

An estimate of the storage coefficient is obtained by adding to the measured runoff the estimated vertical leakage rate from the loess reservoir areas, 0.0005 foot per day, and resolving equation 1 for S. This gives an S equal to 0.02, which is reasonable for a silt-loam deposit.

Areal variations in the effective yield will depend primarily on the local influence of the underlying paleosols because the paleosols control the ratio of vertical to lateral discharge from the loess. Geologic evidence suggests that the distribution of these paleosols in the tributary basin above the low-flow station is representative of their distribution elsewhere in Four Mile Creek basin. Little areal change in effective yield is to be expected from variations in the physical properties of the loess, which are remarkably uniform over wide areas.

Variations with time in the effective yield will occur as the S' computed is related to S just as gravity yield is related to specific yield. Insufficient data, unaffected by evapotranspiration, are available for study of any variations with time.

Further information on the hydrologic characteristics of loess was obtained from a recharge test performed near Iowa City, about 70 miles southeast of the reach along Four Mile Creek. Because the physical properties as well as boundary conditions of the loess at the site of the recharge test are the same as in Four Mile Creek basin, the results of the recharge test are believed to be applicable to the Four Mile Creek area.

The test was made using a recharge well and two observation wells. Observed data, corrected for previous trend, were analyzed by the watertable theory proposed by Boulton (1963) with type-curve solutions by Prickett (1965). The type-curves were modified for partial penetration and anisotropy by the method proposed by Weeks (1964). The test gave a horizontal permeability of 1.6 gpd per sq ft (gallons per day per square foot), a vertical permeability of 1.3 gpd per sq ft, and a storage coefficient of 0.02.

The agreement of the storage coefficient computed from the Iowa City test with that computed for the loess of the Four Mile Creek tributary basin also indicates that the rate of vertical leakage used in computing the storage coefficient for the tributary basin is of the right order of magnitude.

THE ALLUVIAL SAND RESERVOIR

The alluvial sand constitutes the only other reservoir contributing significant amounts of ground water to Four Mile Creek. This deposit occupies the flood plain and extends for some distance up the tributary valleys. It is surrounded by the loess reservoir and is underlain by till.

PHYSICAL PROPERTIES

To determine the physical properties of the alluvium, 52 test holes and (or) observation wells were drilled which penetrated the entire alluvial section. The location of the holes and the distribution and thickness of the alluvial units are shown in plate 1.

The alluvium in the study reach occupies a trough commonly 30 feet in depth underlying the bottom lands. The alluvial sediments consist of two distinct hydrogeologic types: permeable sand and relatively impermeable silts and clays.

The alluvial silts and clays cap the sand and form extensive lenses along the boundaries of the alluvial trough. In the study reach Four Mile Creek cuts into but not through the silts and clays, except for a very short distance near the confluence of Four Mile and Half Mile Creeks. At that site approximately 1–2 feet of sand is exposed in the creek bank.

Between the gaging stations the alluvial sand deposit averages 15 feet in thickness and 1,500 feet in width; it occupies 1.25 square miles of the total reach area of 4.40 square miles. Samples indicate that the deposit grades from fine sand near the top to a fine gravel at the bottom of the trough. Sieve analyses of 55 samples show excellent to good sorting with sorting coefficients ranging between 1 and 2. A histogram of median grain size (fig. 7) shows that the bulk of the sand is uniform in texture. The good sorting and uniform distribution in median grain size indicates that the sand is homogeneous. However, because of textural changes with depth, the sand is not isotropic.

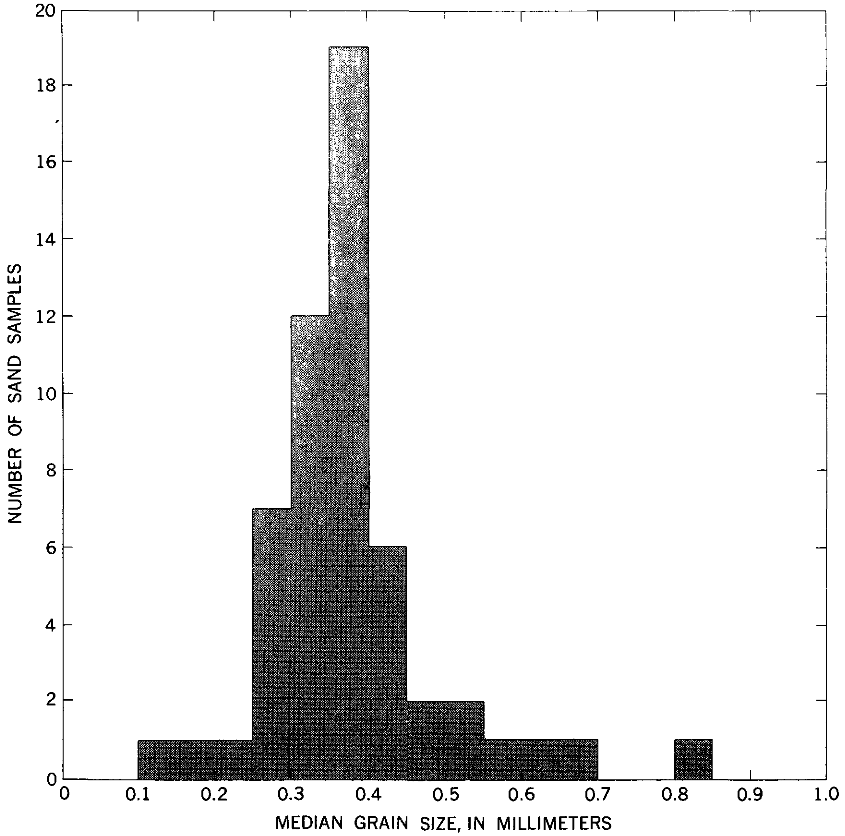

FIGURE 7.—Distribution of median grain size in the alluvial sand.

GROUND-WATER MOVEMENT

To determine the direction of ground-water movement in the alluvial sand, approximately 25 sand-point piezometers were installed in addition to three recording observation wells: well C–1, NE¼NE¼NW¼, sec. 3, T. 85 N., R. 15 W.; well D–5, SE¼NW¼SE¼, sec. 33, T. 86 N., R. 15 W.; and well F–2, SE¼NE¼NW¼, sec. 33, T. 86 N., R. 15 W. Some piezometers were used in well nests to define the vertical distribution of head while others were installed along lines approximately normal to the stream to define horizontal gradients.

The piezometers indicate that ground-water movement in the sand is mainly horizontal except in an area of variable width adjacent to the stream where an upward component of flow exists, and in another area along the alluvial border where downward components of flow exists. The situation is diagramed in figure 8.

In the zone where downward components of flow exist, precipitation recharges the alluvial sand by infiltration through the overlying

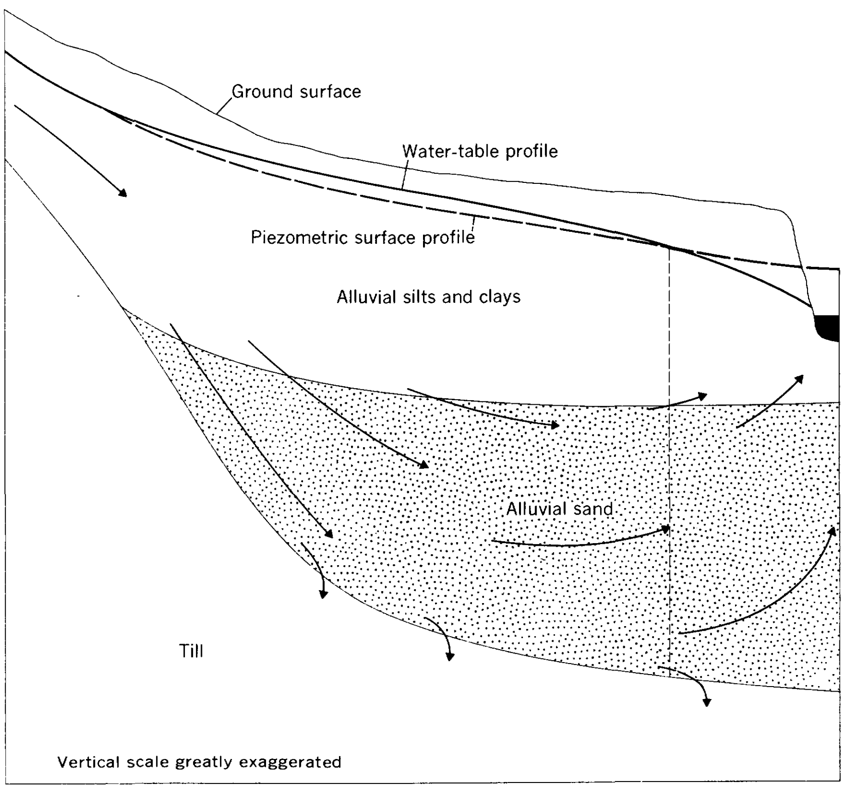

FIGURE 8.—Direction of ground-water movement in the alluvial sand reservoir.

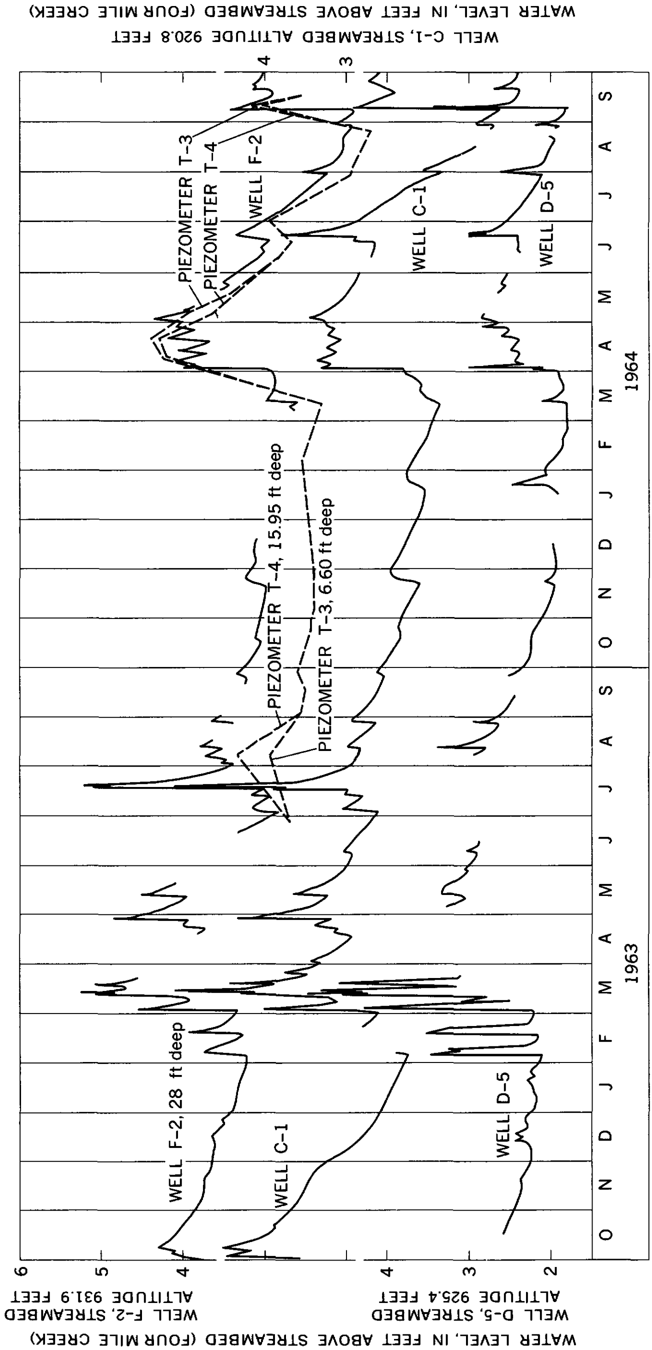

FIGURE 9.—Storage conditions in the alluvial sand reservoir, 1963 and 1964 water years.

silts and clays. Measured vertical gradients range from 0.75 foot per foot, several days after a heavy rainstorm, to 0.05 foot per foot at the end of a month-long recession.

In the zone of upward movement adjacent to the stream, very little recharge by precipitation occurs in the sand. Figure 9 shows the water levels in three closely spaced wells (F–2, T–3, T–4) completed at different depths in this zone. During most of the period of record, the potential in the two shallow wells is below the potential in the deeper well. Only during April 1964 did the potential in the shallow wells exceed that of the deeper well and thereby allow recharge to the sand. Thus, except for wet conditions, generally occuring in the spring, upward movement is maintained and the area is primarily one of ground-water discharge.

The width of the zone of upward movement is variable and appears related to the thickness of the confining silts and clays (pl. 1). Where the confining bed is thick, the zone is wide; where the confining bed is thin, the zone is narrow; and where the confining bed is absent, the width of the zone closely conforms to the stream-channel width.

Delineation of the zone of upward movement, as shown in pl. 1, is mainly by indirect evidence. From many borings and cores, it has been determined that areas of upward ground-water movement coincide in position with areas of glei-colored silts and clays immediately overlying the alluvial sand. Gleization, or the formation of glei colors, is attributed to reduction of ferric iron in the presence of organic matter in an oxygen-deficient environment. Correlation between areas of upward ground-water movement and glei-colored sediments suggests that ground water may be an environmental factor in the gleization process.

The movement of ground water as shown in figure 8 demonstrates how ground water could influence the formation of glei colors. Recharge, as oxygen-rich water, enters the sand primarily in the zone of downward movement. Water moving laterally through the sand and into the zone of upward movement must originate, during most periods of the year, from recharge in the zone of downward movement, or as leakage from the loess reservoir. As the water moves through the deposits, the dissolved-oxygen content will be diminished because of organic activity. Ground water entering the zone of upward movement will then be deficient in oxygen and allow the reduction of ferric iron.

The above hypothesis explaining the control that ground-water movement might have on the formation of glei colors was suggested, although not verified, by two water samples collected from the sand. One sample collected from well C–1, where no vertical flow exists, contained 0.5 ppm (part per million) dissolved oxygen. Another sample collected under identical conditions from a piezometer at the edge of the stream south of well C–1 contained no dissolved oxygen. More information is needed before the concurrence of upward ground-water movement and glei-colored sediments can be definitely substantiated.

HYDROLOGIC CHARACTERISTICS

The hydrologic characteristics of the alluvial sand are not easily determined because of the complex hydrologic conditions and boundaries of the reservoir. Many methods have been tried, but only those listed below have given reasonable and consistent results.

1. Aquifer test using constant discharge.
2. Hydrologic budget and storage depletion using equation 1.
3. Water-level profile, shape, and rate of decline during periods when profiles were stabilized.

Method 1 was employed using well D–5 as the pumped well and four 2-inch observation wells parallel to the stream and at a distance of approximately 90 feet from the stream. Early drawdown data were analyzed by use of the Theis (1935, p. 522) nonequilibrium formula because these data would be least affected by induced infiltration. Because only the first 10 minutes of data were used in the analysis, the effects of leakage into and out of the reservoir and of evapotranspiration were negligible. Type nonequilibruim curves were constructed for each well to correct for partial penetration and anisotropy (Weeks, 1964, p. D195); these represented a family of curves with horizontal to vertical permeability ratios between 0.20 and 0.08. The best fit was obtained with a ratio of 0.10, giving a solution for transmissibility, T, of 5,000 gpd per ft., and a horizontal permeability of 210 gpd per sq ft.

The coefficient of storage from the early-time drawdown data was about 10^{-4}, indicating that the reservoir had not yet drained by gravity and was unaffected by induced recharge. Late-time drawdown data showed the effects of leakage and induced infiltration; however, insufficient data were available to analyze the late data.

Method 2 is used to obtain the average transmissibility of the alluvial sand in the reach and the effective yield of the sand to the creek. Ground-water discharge to the reach is obtained by subtracting 10-day averages of inflow from outflow. The computed gain is considered to be accurate because the periods selected were ones of slow recession with little day-to-day variation in discharge. To eliminate the influence of evapotranspiration, a mid-November period in 1963 was selected that occurred about 10 days after the first killing frost.

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To estimate ground-water runoff from the alluvial sand, the tributary contribution must be computed and subtracted from the total gain in base flow. The tributary contribution is estimated by applying equation 1 to the area of the loess reservoir draining to the tributaries.

The average coefficient of transmissibility of the alluvial sand is estimated by

$$\frac{7.5(Q_b-Q_t)}{2IL}=2{,}850 \text{ gpd per ft,}$$

where Q_b is the base-flow gain in the reach, 24,200 cu ft per day; Q_t is the tributary contribution, 2,950 cu ft per day; I is the hydraulic gradient normal to the stream, 8 feet per mile; and L is the length of the stream reach between gaging stations, 3.5 miles.

The average coefficient of permeability is 2,850/15, or 190 gpd per sq ft, a value in excellent agreement with the results of the aquifer test. An average of the two methods, 200 gpd per sq ft, may be used with the isopach map of the alluvial sand (pl. 1) to determine the coefficient of transmissibility at any site in the reach.

The effective yield, S', to the stream is found by method 2. Rewriting equation 1 for the alluvial sand,

$$S'=(Q_b-Q_l)/(\Delta h/\Delta t)A, \tag{2}$$

where, Q_b is the base-flow gain in the reach; Q_l is the contribution from the loess reservoir, equal to the discharge to the tributaries joining the reach plus the leakage from the loess to the alluvial sand; $\Delta h/\Delta t$ is the average of the recession rates in wells C–1, D–5, and F–2 during the periods considered; and A is the area of the alluvial sand reservoir draining to the reach, equal to 1.25 square miles.

Periods in early November, when evapotranspiration losses are negligible, were selected in 1962 and 1963. Solution of equation 2 gives S' equal to 0.09 for the 1962 period and equal to 0.07 for the 1963 period.

Modification of equation 2 to include vertical leakage, Q_v, from the alluvial sand to the underlying till gives

$$S=(Q_b+Q_v-Q_1)/(\Delta h/\Delta t)\mathrm{A}. \tag{3}$$

Q_v from the alluvial sand was previously computed to be 0.0003 foot per day. By substitution in equation 3, the coefficient of storage is computed to be 0.12 for both the 1962 and 1963 periods.

Because the November periods selected for the above computations occurred after prolonged recessions, S' and S are maximums. These values are applicable only when water-level profiles have become relatively stable and complete gravity drainage occurs. There is reason to believe that during and immediately after recharge periods, or when the alluvial sand reservoir is subjected to a hydrologic impulse, S' and S are not maximums but are significantly less. Because these periods occur frequently and have a significant effect on the discharge recession, it is important to define the minimum values of the effective yield and storage coefficient.

The hydrologic events of April 29 and July 18, 1963, and September 8, 1964, represent impulses during which S' of the reservoir should have been a minimum. Ground-water discharge to the reach was obtained by subtracting 5-day averages of inflow from outflow for times 5–10 days following the peak discharge. The rate of ground water-level recession was obtained by averaging the concurrent recession rates observed in wells C–1, D–5, and F–2. The contribution from the loess reservoir was computed by use of equation 1 and subtracted from the total gain. Because each of the periods considered occurred during the growing season, losses to evapotranspiration had to be estimated (White, 1932, p. 81) and added to the ground-water runoff derived from storage in the reservoir. For each of the three periods, equation 2 gave an estimated effective yield of 0.02, and equation 3 gave a storage coefficient of 0.03.

As might be suspected, the base-flow gain could not be determined accurately during periods of fast recession even though 5-day averages are used. For this reason and because the estimates of evapotranspiration are not precise, the minimum values of effective yield and storage coefficient have been computed to only one significant figure. Even this refinement may be unjustified; however, the agreement among the values for the three periods suggests that the estimates are reasonably correct.

A third method (Rorabaugh, 1960, eq 11) uses the shape of the water-level profile and rate of water-level recession to determine the diffusivity, T/S, of the reservoir. For this method, well C–1 and two sand-point piezometers in line with well C–1 and approximately normal to the direction of ground-water movement were used to define the shape of the water-level profile. At this site the transmissibility of the reservoir is obtained by multiplying the thickness of sand (pl. 1) by the average permeability, 200 gpd per ft. The solution is, therefore, for the coefficient of storage.

A correction for leakage from the alluvial sand to the underlying till must be included, which gives the following modification to Rorabaugh's equation:

$$S=\frac{T}{(\Delta h/\Delta t)[cd(c+d)/2(bc-ad)]}+\frac{L}{(\Delta h/\Delta t)}. \qquad (4)$$

T, S, and $\Delta h/\Delta t$ are as previously defined; a, b, c, and d are defined in figure 10; and L is the estimated leakage rate to the underlying till, 0.0003 foot per day.

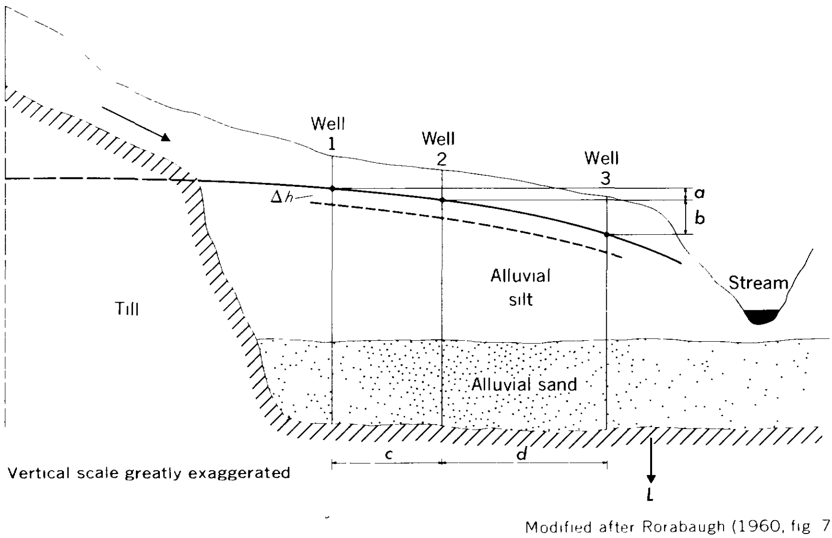


FIGURE 10.—Stabilized water-level profile with vertical leakage to the underlying till. Refer to equation 4 for use of indicated distances.

For solution of equation 4, three late fall and winter periods were selected when water-level data were unaffected by evapotranspiration. Each of the periods coincided with a slow recission. All the values of the storage coefficient determined are near 0.14 and in good agreement with those determined by method 2 for periods when the water-level profiles were stable.

In summary, the average reservoir coefficients for the alluvial sand within the study reach are:

Permeability	200 gpd per sq ft
Transmissibility	3,000 gpd per ft
Effective yield following an impulse	0.02
Storage coefficient following an impulse	.03
Effective yield after prolonged drainage	.08
Storage coefficient after prolonged drainage	.13

GROUND-WATER RUNOFF FROM THE LOESS AND ALLUVIAL SAND RESERVOIRS

STREAMFLOW RECORDS

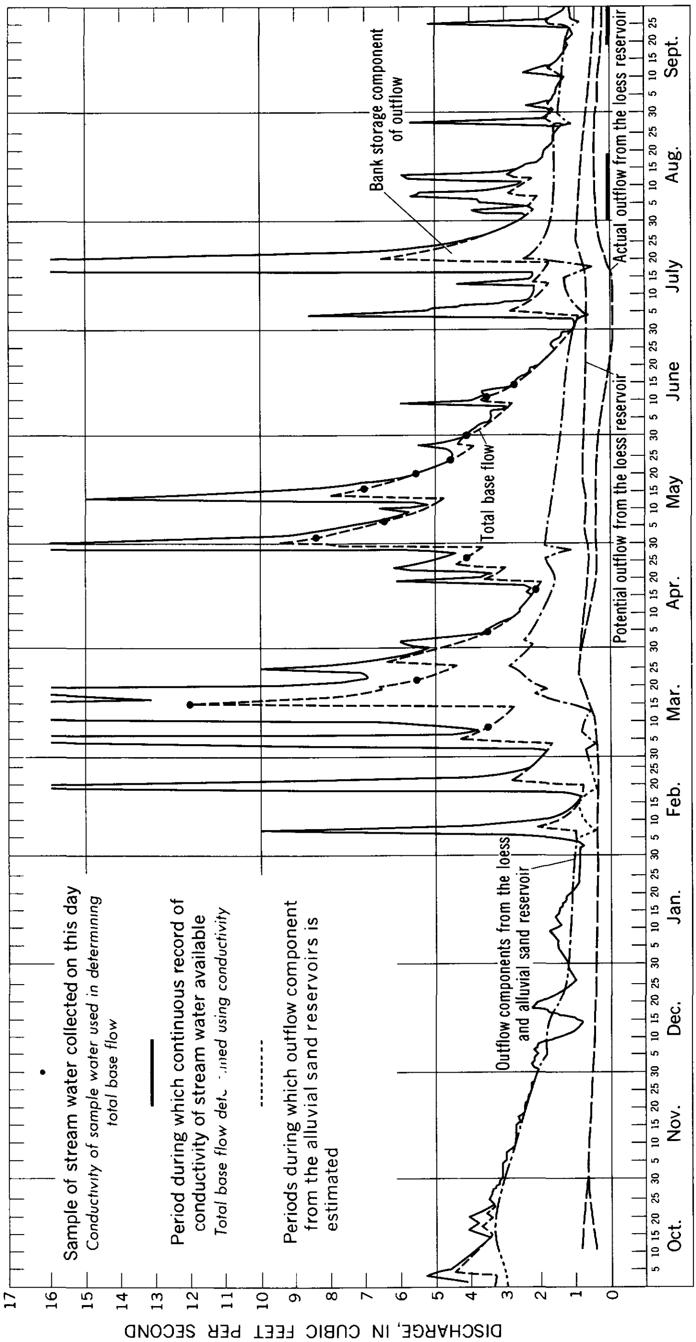
The streamflow stations are equipped with water-stage recorders and concrete controls with V-notch sharp-crested weir plates. Except for periods affected by ice or plugged intakes, the records obtained at the Four Mile Creek stations are believe to be accurate within 5 percent if streamflow is less than 3 cfs and within 3 percent if more than 3 cfs. For Half Mile Creek, the weir plate is rated by volumetric measurements up to 0.5 cfs, and the entire record is considered accurate within 3 percent. The hydrograph for Four Mile Creek near Traer is presented in figure 11. Hydrographs of Four Mile Creek near Lincoln and Half Mile Creek near Gladbrook are very similar in appearance to those shown for the Traer station.

One of the most difficult problems in studying a drainage reach is the accurate measurement of the gain or loss in the reach. Errors in determination of inflow and outflow, while small, may be additive and amount to more than 50 percent of the gain in the reach. Thus, no daily discharge hydrograph could be computed for the reach, although average gains in the reach could be determined during extended base-flow periods.

HYDROGRAPH SEPARATION

The ground-water outflow hydrographs (fig. 11) show the combined effects of the two reservoirs on the stream. By considering the hydrologic characteristics of the reservoirs, the total outflow may be divided into the outflow components from each reservoir. From the separated hydrograph the annual outflow can be computed for later use in a ground-water budget.

Because a hydrograph of discharge for the gaged reach could not be defined, the reservoir information is extended from the reach to include the entire Four Mile Creek basin. In thus extending the data, several assumptions are made. First, it is assumed that the percentages of the basin areally covered by the two reservoirs is the same as in the reach; that is, 30 percent of the basin is occupied by the alluvial sand reservoir and 70 percent by the loess reservoir. Second, it is assumed that the percentage contribution from each reservoir into the gaged reach applies to the entire basin. Third, it is assumed that the alluvial sand reservoir can be treated as a tetrahedron pinching out upstream. If, then, T in the downstream part of the reservoir averages 3,000 gpd per sq ft, the average T for the basin is 3,000/2, or 1,500 gpd sq ft. Likewise, if a, the half width of the tetrahedron base, is 1,500 feet, the average a of the reservoir in the basin is 750 feet. S and S' for the



Sample of stream water collected on this day
Conductivity of sample water used in determining total base flow
Period during which continuous record of conductivity of stream water available
Periods during which outflow component from the alluvial sand reservoirs is estimated
Bank storage component of outflow
Total base flow
Outflow components from the loess and alluvial sand reservoir
Potential outflow from the loess reservoir
Actual outflow from the loess reservoir
DISCHARGE, IN CUBIC FEET PER SECOND
Oct.
Nov.
Dec.
Jan.
Feb.
Mar.
Apr.
May
June
July
Aug.
Sept.

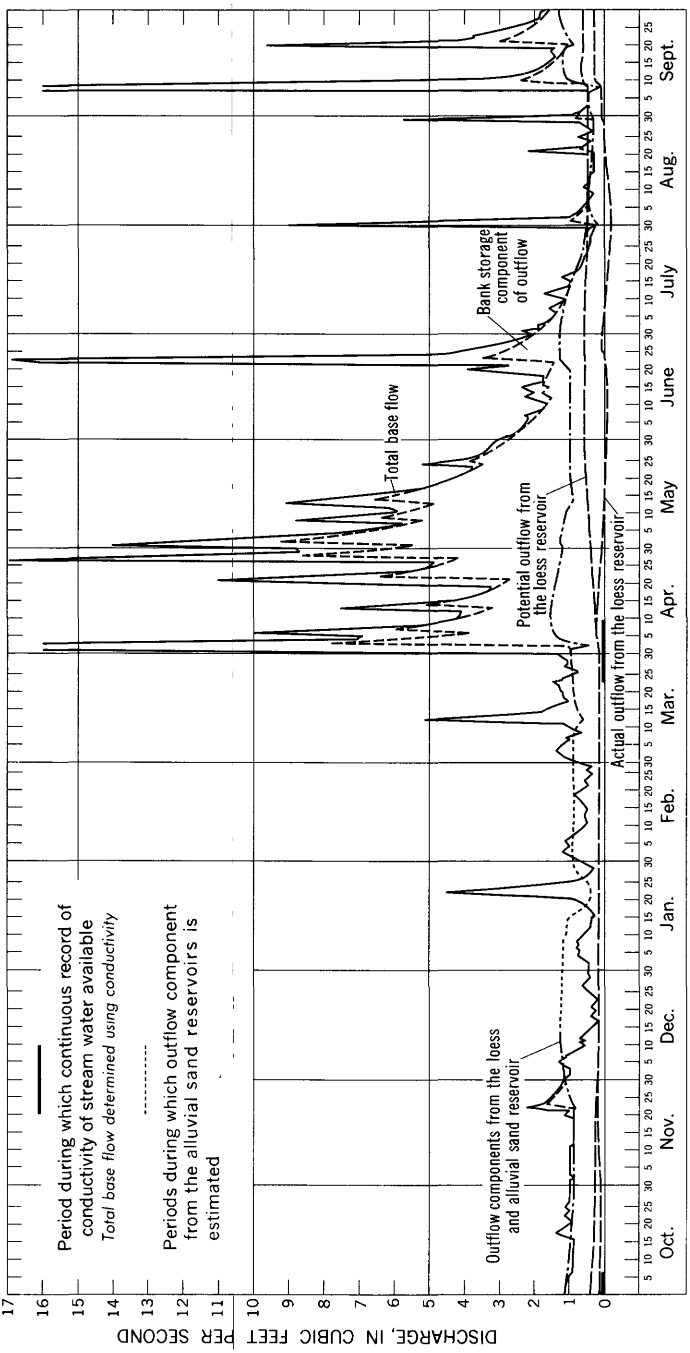


FIGURE 11.—Discharge of Four Mile Creek near Traer, 1963 and 1964 water years.

alluvial sand reservoir are assumed not to change in the basin. Lastly, it is assumed that the hydrologic characteristics of the loess reservoir, like the geologic characteristics, do not change significantly throughout the basin.

Base flow from the alluvial sand reservoir consists of the reservoir component of outflow and the bank storage component of interchange between ground water and surface water. Because water levels in the reservoir respond to stage changes in the stream, the bank-storage component will be a significant proportion of the total base flow from the reservoir.

Added to the two components of runoff from the alluvial sand reservoir is the component of outflow from the loess reservoir.

TOTAL GROUND-WATER OUTFLOW

To obtain the total ground-water outflow, an empirical approach was developed to separate ground-water runoff from surface runoff. The approach (Kunkle, 1965) utilizes the specific conductance of the stream water at the gage and representative conductance values of ground and surface water discharged by the stream.

Specific-conductance measurements of creek water at the gaging station near Traer show that during periods of base flow the conductance of the water, regardless of the magnitude of base flow, is 520 micromhos ± 1 percent. This value agrees well with conductance values of water samples collected from wells in the loess and alluvial sand reservoirs. From water samples of flood flows and surface detention collected during rainstorms, the conductance of surface runoff water (direct overland flow) is found to be in the range of 120–250 micromhos with a median value of about 160 micromhos. Using 160 micromhos as a representative value for water discharged as surface runoff and 520 micromhos as a representative value for water discharged as ground-water runoff, measurements of the conductance of the mixed streamflow water permit base flow to be determined by the following equations (Kunkle, 1965):

$$520x+160y=Cq$$

and

$$x+y=q$$

where

x=ground-water discharge, in cfs,

y=surface-water discharge, in cfs,

q=total streamflow, in cfs, and

C=specific conductance of the creek water, in micromhos.

From early March through June 1963 a number of water samples were collected at the Four Mile Creek near Traer gaging station; and beginning in August, a portable recording conductivity meter was installed. Owing to battery and clock trouble, a continuous record was not obtained with the recording meter. Days on which water samples were collected and periods during which conductivity was recorded are indicated on figure 11.

The records obtained using the recording conductivity meter indicate that immediately after a streamflow peak, ground-water runoff increases sharply with a ground-water peak within 1 day of the streamflow peak (Kunkle, 1965, fig. 1). The hydrograph separations based on conductance measurements were assumed to be valid and were used as models to separate the hydrograph for the entire period of streamflow record. Increases in ground-water outflows are shown on figure 11 to begin on the day of the streamflow peak, and ground-water peaks are shown to occur on the day following the stream-flow peak. On recession of flow the ground-water hydrograph is merged with the hydrograph of total streamflow to give a continuous hydrograph of ground-water outflow.

The ground-water peaks cannot quantitatively be verified using the ground-water theory as described by Cooper and Rorabaugh (1963) because of differences between the ideal theoretical stream reach and the actual integrated stream. However, the time of the ground-water peak, as herein derived, agrees closely with the theoretical time of ground-water peak.

ALLUVIAL SAND COMPONENT OF OUTFLOW

In estimating the alluvial sand component of outflow, the assumption is made that outflow is proportional to head in the reservoir. Rorabaugh (1964, p. 438) used a similar procedure and included a discussion of the theory.

Head in the reservoir (fig. 12) is obtained by plotting to a common datum and subtracting the water level in well C–1 and the stage in Four Mile Creek. Comparison of figure 12 with figure 9 shows that most of the water-level fluctuations in the reservoir are stream produced.

For selected periods when recharge and evapotranspiration are negligible, the outflow component to the gaged reach is computed by equation 1 and reduced to a percentage of the total base-flow gain in the reach. This percentage is multiplied by the concurrent base flow at the Traer gaging station to obtain the outflow component from the alluvial sand reservoir in the basin.

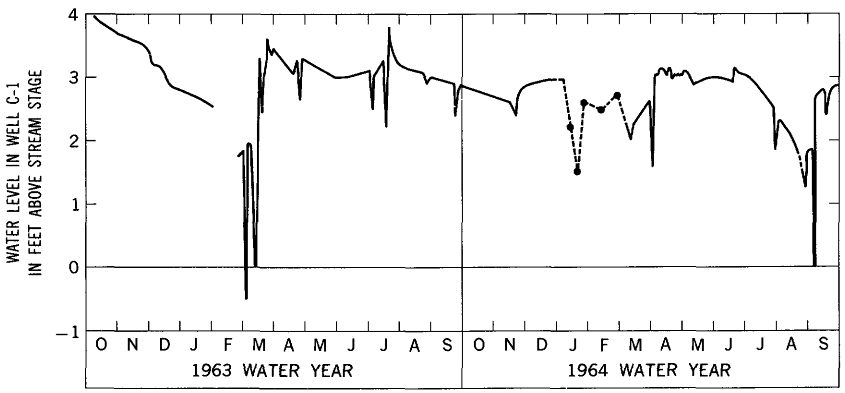


FIGURE 12.—Ground-water stage in well C–1 measured above stage in Four Mile Creek.

Outflow from the alluvial sand reservoir for these selected periods is then correlated with the head in well C–1 to obtain a head-discharge relation curve (fig. 13). From this curve and the hydrograph of head (fig. 12), the outflow component from the alluvial sand reservoir for any period is obtained and plotted on the streamflow hydrograph.

Although the head-discharge relation is based on non-growing-season data, the relation is also applicable to the growing season. Discharge computed from the curve is partially corrected for evapotranspiration losses because the head also reflects these losses, assuming,

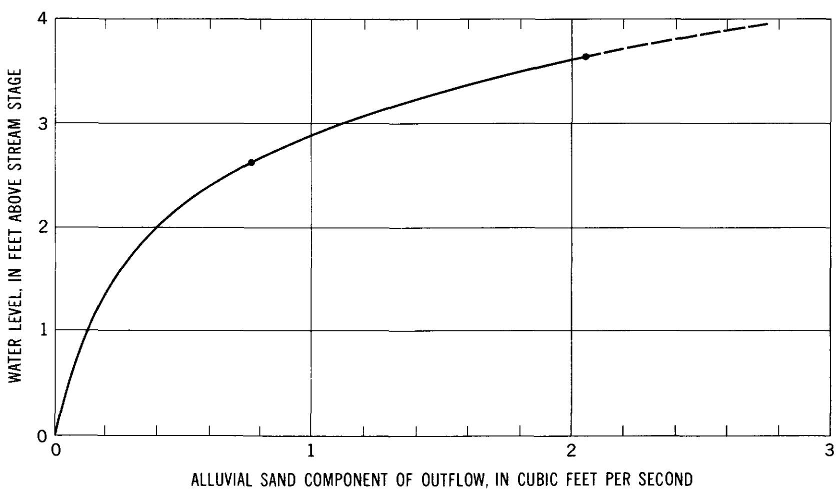


FIGURE 13.—Relation between ground-water stage in well C–1 and the alluvial sand component of outflow.

of course, that head adjustments in well C–1 adequately reflect the average losses from the reservoir. Within practical limits this assumption is believed to be reasonably correct.

THE LOESS COMPONENT OF OUTFLOW

Outflow from the loess reservoir is assumed to be proportional to the head in well E–1 (fig. 6). From equation 1, the outflow component to the gaged reach is calculated for selected periods when recharge and evapotranspiration are negligible. The component is reduced to a percentage of the total base-flow gain in the reach. This percentage is multiplied by the concurrent base flow at the Traer gaging station to obtain the outflow from the loess reservoir within the basin.

Outflow from the loess reservoir for these selected periods is then correlated with the head in well E–1 to obtain a head-discharge relation curve (fig. 14). From this curve and the hydrograph of well E–1, the outflow component from the loess reservoir for any period is obtained and plotted on the streamflow hydrograph (fig. 11).

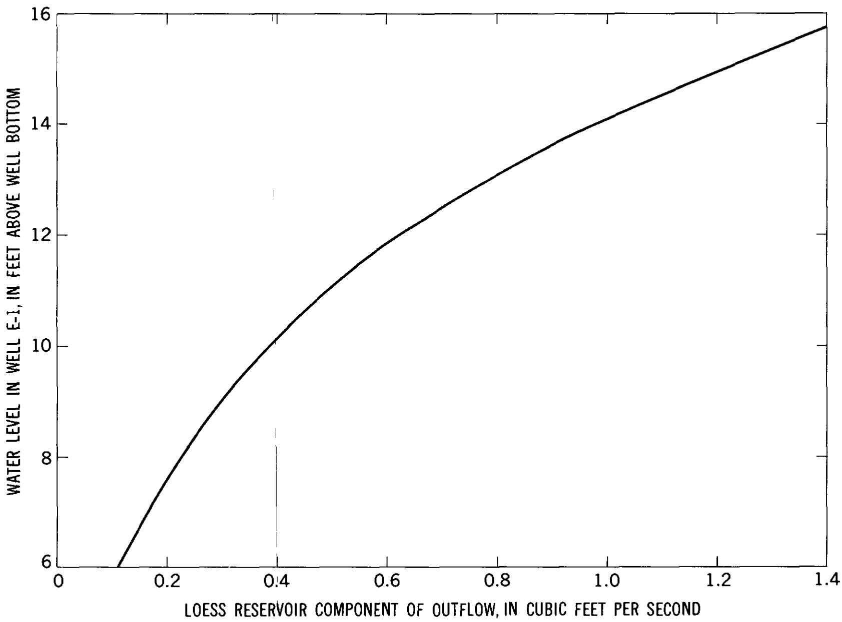


FIGURE 14.—Relation between ground-water stage in well E–1 and the loess component of outflow.

Addition of the loess and alluvial-sand components of outflow gives a potential base flow in excess of the observed base flow during parts of the growing season. The excess represents potential outflow from the loess reservoir lost by evapotranspiration. Prediction of excess discharge results from using well, E–1, whose water level is largely unaffected by evapotranspiration losses from the reservoir.

An empirical method for evaluating evapotranspiration loss is afforded by relating the average computed end-of-month excess to a monthly weather index (fig. 15). The weather index used is the monthly pan evaporation for Ames, Iowa, as reported by the U.S. Weather Bureau. With the weather index known, evapotranspiration loss is determined from figure 15 and subtracted from the computed potential outflow to estimate the actual outflow.

Note that the adjustments for evapotranspiration from the loess result in negative base flows during midsummer. Physically, it is possible for ground-water discharge to become negative or for evapotranspiration to exceed potential base flow to some reaches of the stream.

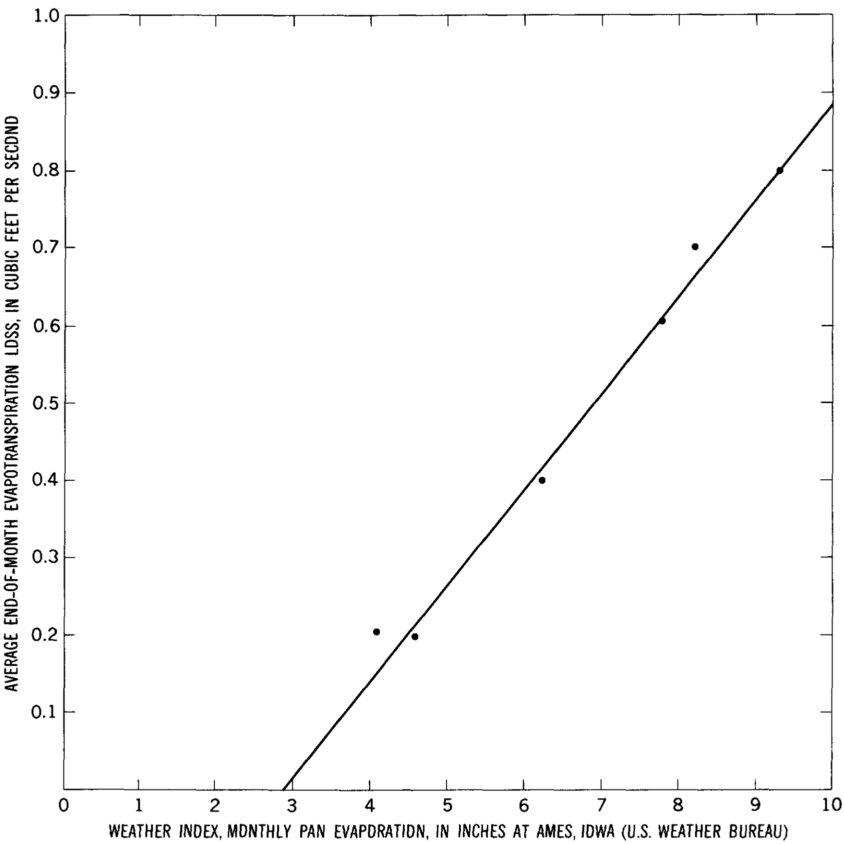


FIGURE 15.—Drainage basin evapotranspiration from the loess reservoir related to a weather index.

The negative flow may indicate that ground-water levels in the riparian zone of the tributaries have been depleted to a level below the stream, causing ground-water runoff to cease.

BANK STORAGE COMPONENTS

Bank storage was defined by Houk (1951, p. 179) as the stream water absorbed into the banks of the stream channel when the stream stage rises above the ground-water level in the adjacent bank. When the stream stage falls, the water stored in the banks returns to the stream. This definition of bank storage is restrictive in that it omits mention of an integral part of bank storage, which is the water stored in the reservoir that would have discharged to the stream had there been no rise in stream stage. This water may be termed ground-water backwater. In practice it is almost impossible to separate bank storage as defined by Houk from backwater. Theoretical equations and curves for computing all water stored in the ground as a result of a flood wave are given by Cooper and Rorabaugh (1963, figs. 102 and 103) and Rorabaugh (1964, p. 435). However, these equations assume ideal stream reaches and cannot be applied to an integrated stream basin such as that of Four Mile Creek.

For the purpose of this report, the concept of bank storage is enlarged to include all storage in the ground-water reservoir produced by stage changes in the stream. According to this definition the changes in streamflow due to bank storage consist of three components: the bank storage component of inflow, the bank storage component of outflow, and the component of backwater.

Quantitatively, the bank storage component of outflow is equal to the total ground-water component of outflow minus the components of outflow from the loess and alluvial sand reservoirs. As shown in figure 11, the bank storage component of outflow represents a significant proportion of the total base flow.

From the hydrologic characteristics of the alluvial sand reservoir, the effects of the bank storage component of outflow on the discharge recession curve can be calculated. T for the reservoir within the basin is 1,500 gpd per sq ft, or 200 ft^2 per day; a is 750 feet; and the effective yield S' is 0.02, when the reservoir is subjected to an impulse. The reservoir coefficient is T/a^2S' or 0.018 days^{-1}. According to Rorabaugh (1964, fig. 1), the exponential slope of the recession curve, after some critical time, t_c, will be $0.933/(T/a^2S')$ or 52 days per log cycle and t_c will be $0.2T/a^2S'$ or 10 days.

On the streamflow hydrograph the loess component of outflow was subtracted from the total ground-water component of outflow for the following 1963 periods: March 26 to April 15, April 29 to May 11, and

May 13 to May 20. The remainder, or the combined reservoir and bank storage components of outflow, was found to have recession slopes of 48, 54, and 54 days per log cycle, respectively. Furthermore, the critical time, t_c, ranged from 10 to 12 days. Because these values, determined from the hydrograph separations, agree very closely with those computed using ground-water theory, it is believed that the separations must fairly accurately portray the components of outflow.

The bank storage components of inflow and backwater cannot be separated. They are determined together from the effect of the stream on water levels in the reservoir. These components do not enter into the hydrograph separation and will be discussed later under the ground-water budget.

ANNUAL GROUND-WATER BUDGET

The local ground-water flow system of Four Mile Creek basin is duplicated in most respects throughout the Wolf Creek basin and, except for vertical leakage, is similar to many small basins in eastern Iowa. Within practical limits, then, a quantitative ground-water budget for Four Mile Creek basin will have carryover value to other areas of Iowa which are climatologically and geologically homogeneous.

A hydrologic budget is a balance between inflow and outflow and is quantitatively stated by the following equation:

$$W = R_g + ET_g + L_v \pm L_r \pm \Delta S_g,$$

where

W=ground-water recharge,

R_g=ground-water runoff,

ET_g=evapotranspiration of ground water,

L_v=vertical ground-water leakage to the underlying till,

L_r=lateral ground-water leakage from the loess to the alluvial sand (positive for the loess reservoir and negative for the alluvial sand reservoir), and

ΔS_g=changes in ground-water storage.

Each of the budget components on the right side of the equation can be determined separately for each of the ground-water reservoirs. Thus, recharge to the alluvial sand and recharge to the loess reservoir may be determined. The computations are made for Four Mile Creek basin above the Traer gaging station.

Ground-water runoff is evaluated from the streamflow hydrograph (fig. 11) and is divided into runoff from the alluvial sand reservoir and runoff from the loess reservoir.

Evapotranspiration from the loess reservoir must be determined in two steps. First, for those areas of the loess reservoir which drain to tributaries of Four Mile Creek, evapotranspiration is equal to the potential ground-water runoff minus the actual runoff. The annual values are evaluated from the streamflow hydrograph (fig. 11) and are equal to 0.2 inch for both the 1963 and 1964 water years. Second, for those areas of the loess reservoir which drain to the alluvial sand, evapotranspiration cannot be computed. However, evapotranspiration of potential leakage to the alluvial sand should be less than the losses from potential runoff to the tributary streams because there are fewer phreatophytes along the loess-alluvial sand contact to intercept leakage than along the tributaries. Because of this factor, it is estimated that the total evapotranspiration from the loess reservoir did not exceed 0.3 inch during each water year.

Evapotranspiration from the alluvial sand reservoir is obtained by plotting (fig. 16) the daily evapotranspiration as measured in each of the three recording observation wells, C–1, D–5, and F–2. The daily loss is equal to the diurnal change in stage caused by evapotranspiration times the specific yield of the deposits (0.13). An average of the losses recorded by the three wells is used to estimate annual evapotranspiration from the reservoir. To convert the losses into inches over

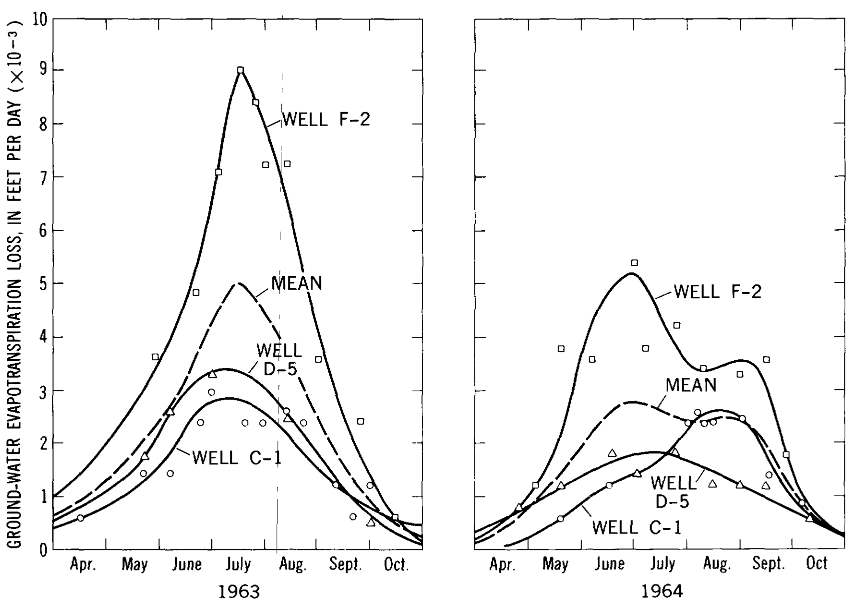


FIGURE 16.—Ground-water evapotranspiration from the alluvial sand reservoir determined by diurnal fluctuation of water levels in three recording wells.

the basin, the loss in inches on the reservoir had to be multiplied by an area factor, equal to 0.3.

This method of estimating evapotranspiration losses from ground water is the same as that proposed by White (1932, p. 81). Stallman (1961) noted that the analytical solution to White's method is incomplete and does not account for lateral ground-water movement. Conclusive evidence is presently not available to show that the complete solution is always necessary or entirely correct. Both the incomplete and complete solutions relate the position of the water table solely to flow in the saturated zone, whereas, as Stallman pointed out, the water table moves in response to water transfer in both the saturated and unsaturated zones.

Diurnal fluctuations in the alluvial sand are typified by small to no changes in stage during the night as opposed to large amounts of recession during the daytime. The small changes in stage at night indicate that lateral transfer of water is small. Therefore, the use of White's solution for estimating evapotranspiration is reasonable, but owing to a lack of precision in this and other approaches, components of the annual ground-water budget are reported to only one significant figure.

Vertical leakage rates are 0.0005 foot per day from the loess and 0.0003 foot per day from the alluvial sand. These rates are converted to inches per year and multiplied by an area factor (0.3 for the alluvial sand reservoir and 0.7 for the loess reservoir) to obtain inches for the basin.

Lateral leakage from the loess to the alluvial sand, $+L_r$, is equal to potential leakage minus evapotranspiration. Potential leakage was estimated to be 0.6 inch in 1963 and 0.4 inch in 1964 from the following relation:

$$\frac{\text{Area of reservoir draining to alluvial sand}}{\text{Area of reservoir drainage to tributaries}}\text{(potential runoff)}=\text{potential lateral leakage.}$$

Subtracting the estimated evapotranspiration, 0.1 inch in both 1963 and 1964, from potential lateral leakage gives the annual rate of lateral leakage—0.5 inch for 1963 and 0.3 inch for 1964.

Changes in storage for the loess reservoir are equal to the specific yield (0.02) times the change in stage measured in the key loess well, E–1, multiplied by the area factor. Changes in storage in the alluvial sand reservoir are equal to the specific yield (0.13) times the average change in stage in wells C–1, D–5, and F–2, multiplied by the area factor.

As previously noted, the storage coefficient is not a constant. Consequently, errors in the change in storage are introduced when large changes in stage occur. As the change in stage approaches zero, the error in storage change also rapidly approaches zero. By selection of the water year as the inventory period, changes in storage are minimized, as is the error in estimate.

Table 1 summarizes the components of the ground-water budget in Four Mile Creek basin above the Traer gaging station. The values of recharge also apply within a few percent to the basin above the Lincoln station. The budget for Half Mile Creek basin could not be estimated because no ground-water information is available upstream from the gage on that stream. In 1963 ground-water runoff from Half Mile Creek basin was 0.40 inch greater than, and in 1964 equal to, ground-water runoff at the Traer gaging station. R_g, ET_g, and ΔS_g for Half Mile Creek basin seem to have a greater annual range than for the larger Four Mile Creek basin.

TABLE 1.—*Annual Ground-Water Budget, Four Mile Creek Basin above Traer Gaging Station*

[Units are inches]

Ground-water Component	Water year 1963	Water year 1964
Alluvial Sand Reservoir		
Discharge:		
As reservoir component of outflow	0.9	0.6
As bank storage component of outflow	.7	.5
As vertical leakage to till	.4	.4
As evapotranspiration to atmosphere	1.9	1.3
Total	3.9	2.8
Storage changes	−.1	−.1
Recharge:		
From loess reservoir	0.5	0.3
As bank storage component of inflow	0–1.7	0–.7
From precipitation	3.3–1.6	2.4–1.7
Total	3.8	2.7
Loess Reservoir		
Discharge:		
As reservoir component of outflow	0.3	0.1
As lateral leakage to alluvial sand	.5	.3
As vertical leakage to till	1.4	1.4
As evapotranspiration to atmosphere	.3	.3
Total	2.5	2.1
Storage changes	−.6	+.2
Recharge from precipitation	1.9	2.3

Solution of the budget equation for the alluvial sand reservoir gives 3.8 inches of recharge in 1963 and 2.7 inches in 1964. The recharge is derived from three sources: infiltration of precipitation, lateral leakage into the alluvial sand from the loess reservoir, and bank-storage inflow. As previously shown, it is possible to account quantitatively for lateral leakage. However, infiltration of precipitation and bank-storage inflow cannot be determined separately.

Limits on bank-storage inflow may be assigned by computing storage changes in the reservoir produced by stream-stage changes. The storage changes may be all bank-storage inflow, all ground-water backwater, or, more probably, some combination of the two. By assuming that the storage changes are all bank-storage inflow, an upper limit to recharge from the stream is determined. Likewise, by assuming that all the storage changes are ground-water backwater, the maximum amount of precipitation recharge is arrived at.

The combined effect of bank-storage inflow and ground-water backwater can be estimated by summing the instantaneous stage rises in wells C–1 and F–2, averaging the two wells, and multiplying the cumulative average by the storage coefficient applicable to impulses, 0.03. This procedure assumes that all the instantaneous rises in well stage are steam produced or that recharge by precipitation does not influence water levels in the reservoir until sometime after the peak rise in ground-water stage. The assumption is substantiated by the response in ground-water levels to stream-stage changes during winter when recharge is nil.

The sum of the water-level rises attributed to bank-storage inflow and ground-water backwater totaled 1.7 inches in 1963 and 0.7 inch in 1964. Because bank-storage inflow could be all or nothing of the total, the limits on recharge by the stream are 0–1.7 inches and 0–0.7 inch, respectively.

At present there is no way to further distinguish bank-storage inflow from ground-water backwater. For this reason the concept of bank storage was enlarged to include backwater.

SUMMARY AND DISCUSSION OF THE SYSTEM

Ground-water movement in Four Mile Creek basin may be divided into a regional flow system and a local flow system. Regional flow is not confined within the surface drainage divides of the basin. Groundwater enters the regional flow system as downward leakage through 170–360 feet of Pleistocene drift. At depth, the water recharges limestone aquifers and then moves into buried bedrock channels infilled with permeable sand deposits. These channels act as buried drains and divert the water many miles from the point of recharge. It is estimated that about 1.8 inches a year is lost from the local flow system to the regional flow system.

The local flow system contributes base runoff to Four Mile Creek and is typical of many such systems operating in this part of Iowa.

Ground-water runoff in Four Mile Creek basin is derived from an interconnected two-reservoir system composed of upland loess and bottom-land alluvial sand. Both reservoirs are underlain by a thick deposit of till through which leakage from the reservoirs is lost from the basin.

The loess reservoir received an average of 2.1 inches of recharge during the 1963 and 1964 water years. Lateral discharge from the loess reservoir was small, averaging 0.9 inch of which 0.3 inch was lost by evapotranspiration. Most of the recharge, 1.4 inches, was discharged as vertical leakage to the underlying till. There was a net decrease in groundwater storage during the 2 years.

One-third of the lateral discharge from the loess is lost by evapotranspiration. The losses occur mainly in the riparian zones of the tributaries which drain the loess reservoir. These tributaries commonly become dry before the runoff reaches the main stem of Four Mile Creek. Consequently, runoff from the loess may become zero, and in some instances evapotranspiration may exceed potential runoff, such as during the midsummer months.

Only during late fall and winter does runoff from the loess compose a significant proportion of the total base flow. For example, during December 1963 and 1964 the loess contributed as much as 33 and 25 percent, respectively, of total base flow. The percentages are higher in late fall and winter because runoff from the alluvial sand reservoir is relatively small at that time compared to other seasons. Except for the growing season, runoff from the loess is relatively constant with a small peak in spring.

In contrast to the relatively stable but small base flow from the loess reservoir, base flow from the alluvial sand reservoir is highly variable. The principal reason for this is the intimate hydraulic connection between the reservoir and the stream coupled with the semiartesian nature of the reservoir. Base flow is augmented substantially by bank storage outflow, especially during spring. During bank-storage depletion, which is rapid, streamflow recedes about four times faster than after bank storage is exhausted. During the growing season the rate of recession is increased by evapotranspiration.

Quantitatively, bank-storage outflow accounted for 44–46 percent of the ground-water outflow from the alluvial sand reservoir. Of the total base flow from the basin, bank storage accounted for 37–42 percent.

Recharge to the alluvial sand reservoir was 3.8 inches in 1963 and 2.7 inches in 1964. Most of the difference is apparently related to difference in recharge from the stream, or bank-storage inflow. Quantitatively, bank-storage inflow cannot be determined but is known not to have exceeded 1.7 inches in 1963 and 0.7 inch in 1964. These maximum values indicate that bank-storage inflow is quite variable from year to year

In conclusion, base flow derived from the local flow system is dominated to a very great degree by runoff from the flood-plain reservoir. A similar observation is given by Kunkle (1962) for the Huron River basin in southeastern Michigan, and by Meyboom (1961) for the Elbow River basin in southwestern Alberta. In east-central Iowa the flood-plain reservoirs are composed of alluvium which varies from basin to basin as well as within basins. Because of the wide range of transmissibilities, storage coefficients, and reservoir widths possible for flood-plain reservoirs, the reservoir coefficients and, hence, base-flow characteristics will also be highly variable. To some extent, however, runoff from the loess, which is a relatively uniform deposit over much of Iowa, will tend to smooth out the differences in the base-flow characteristics between basins. The influence of the loess will be greatest in autumn and winter when base flow from the alluvium is minimal, and will be least in summer when evapotranspiration consumes a very large percentage of the potential runoff from the loess reservoir.

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