U.S. DEPARTMENT OF COMMERCE
National Technical Information Service
PB-288 135

Quantitative Hydrogeology of the Upper Republican Natural Resources District Southwest Nebraska

(U.S.) Geological Survey, Lincoln, NE Water Resources Div

WRI 78-38

Jun 78

QUANTITATIVE HYDROGEOLOGY OF THE UPPER REPUBLICAN NATURAL RESOURCES DISTRICT, SOUTHWEST NEBRASKA

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations 78-38

Prepared in cooperation with
Conservation and Survey Division,
Institute of Agriculture and Natural Resources,
The University of Nebraska-Lincoln and
Upper Republican Natural Resources District

BIBLIOGRAPHIC DATA SHEET	1. Report No. USGS/WRD/WRI-78/077	2.	3. Recipient's Accession No.

4. Title and Subtitle	5. Report Date
QUANTITATIVE HYDROGEOLOGY OF THE UPPER REPUBLICAN NATURAL RESOURCES DISTRICT, SOUTHWEST NEBRASKA	June 1978
	6.
7. Author(s) Eric G. Lappala	**8. Performing Organization Rept. No.** WRI 78-38
9. Performing Organization Name and Address U.S. Geological Survey, Water Resources Division, Room 406, Federal Bldg. and U.S. Courthouse, 100 Centennial Mall - North, Lincoln, Nebraska 68508	**10. Project/Task/Work Unit No.**
	11. Contract/Grant No.
12. Sponsoring Organization Name and Address U.S. Geological Survey, Water Resources Division, Room 406, Federal Bldg. and U.S. Courthouse, 100 Centennial Mall - North, Lincoln, Nebraska 68508	**13. Type of Report & Period Covered** Final
	14.

15. Supplementary Notes

Prepared in cooperation with the Conservation and Survey Division of the University of Nebraska-Lincoln and the Upper Republican Natural Resources District.

16. Abstracts

Ground-water use for irrigation from the Ogallala Formation has increased rapidly in the Upper Republican Natural Resources District in southwest Nebraska. Water levels declined as much as 16 feet between 1952 and 1975. Discharge of the aquifer to streams was reduced by as much as 19 percent between 1967 and 1975. Quantification of the hydrogeologic system was provided by the development, testing and use of simulation models describing the soil zone and ground-water/surface-water system. Models were linked through source-sink terms. The ground-water/surface-water model was used to predict changes in water levels and streamflow caused by unrestricted irrigation-well installation and by no new wells after 1975. Water levels may decline as much as 140 feet in two areas by 2000 if installation of new wells is unrestricted. By 2000, water-level declines over the remainder of the area would be less than 60 feet under continued development and less than 40 feet with no new wells after 1975. The base flow of Frenchman, Stinking Water, and Spring Creeks could be reduced by more than 90 percent with no new wells and eliminated by 1992 under continued development.

17. Key Words and Document Analysis. 17a. Descriptors

*Ground-water mining, *Surface-water/ground-water relationships, *Soil-water-plant relationships, *Computer model, irrigation.

17b. Identifiers/Open-Ended Terms

*Ogallala Formation, *Southwest Nebraska, *Stream depletion, sensitivity analyses.

17c. COSATI Field/Group

18. Availability Statement	19. Security Class (This Report)	21. No. of Pages
Prepared for NTIS by U.S. Geological Survey, WRD. No restriction on distribution	UNCLASSIFIED	212
	20. Security Class (This Page) UNCLASSIFIED	**22. Price** MF PCA10/A01

QUANTITATIVE HYDROGEOLOGY OF THE UPPER REPUBLICAN NATURAL RESOURCES DISTRICT, SOUTHWEST NEBRASKA

By Eric G. Lappala

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations 78-38

Prepared in cooperation with
Conservation and Survey Division,
Institute of Agriculture and Natural Resources,
The University of Nebraska-Lincoln and
Upper Republican Natural Resources District

June 1978

ia

ATTENTION

AS NOTED IN THE NTIS ANNOUNCEMENT, PORTIONS OF THIS REPORT ARE NOT LEGIBLE. HOWEVER, IT IS THE BEST REPRODUCTION AVAILABLE FROM THE COPY SENT TO NTIS.

1b

UNITED STATES DEPARTMENT OF THE INTERIOR

CECIL D. ANDRUS, Secretary

GEOLOGICAL SURVEY

H. William Menard, Director

For additional information write to:

U.S. Geological Survey
Room 406, Federal Bldg. and U.S. Courthouse
100 Centennial Mall North
Lincoln, Nebraska 68508

CONTENTS

ILLUSTRATIONS

CONTENTS

CONTENTS

TABLES

SELECTED FACTORS FOR CONVERTING U.S. CUSTOMARY UNITS TO INTERNATIONAL SYSTEM (SI) UNITS

A dual system of measurements--U.S. customary units and the International System (SI) of units--is given in this report. SI is a consistent system of units adopted by the Eleventh General Conference of Weights and Measures in 1960. Selected factors for converting U.S. customary units to SI units are given below.

Multiply U.S. customary units	By	To obtain SI units
acre	0.0040	square kilometer (km^2)
	.4047	square hectometer (hm^2)
acre-foot (acre-ft)	1233	cubic meter (m^3)
acre-foot per year (acre-ft/yr)	1233	cubic meter per year (m^3/yr)
cubic foot per second (ft^3/s)	.02832	cubic meter per second (m^3/s)
foot (ft)	.3048	meter (m)
foot per day (ft/d)	.3048	meter per day (m/d)
foot per year (ft/yr)	.3048	meter per year (m/yr)
inch (in)	25.4	millimeter (mm)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)
British thermal unit (Btu)	1055	joule (J)
pound per square inch (psi)	6895	pascal (Pa)

QUANTITATIVE HYDROGEOLOGY OF THE
UPPER REPUBLICAN NATURAL RESOURCES DISTRICT,
SOUTHWEST NEBRASKA

By Eric G. Lappala

ABSTRACT

Ground-water use has increased rapidly in the Upper Republican Natural Resources District in southwest Nebraska with recent irrigation development. The principal aquifer being developed comprises saturated sand and gravel of the Ogallala Formation of Tertiary age. Water levels in this aquifer have declined as much as 16 feet between 1952 and 1975. Natural discharge of the aquifer to perennial streams has been reduced by as much as 19 percent between 1967 and 1975.

Good management of the water resources of the District requires quantitative knowledge of the operation of the hydrogeologic system. Quantification was provided through the development and use of simulation models describing the operation of the land surface-plant-soil and ground-water phases of the hydrologic cycle. An integrated approach to simulation was used wherein models of the soil and saturated zones were linked through source-sink terms.

Both models were tested against documented hydrologic conditions, and sensitivity analyses were utilized extensively in the testing process. After the models were considered sufficiently representative of the operation of the actual hydrogeologic system, they were used to predict future rates of water-level changes and streamflow depletions caused by two possible futures. One was continued unrestricted private irrigation-well development, and the other allowed no additional development after 1975.

These analyses indicate water-level declines of as much as 140 feet in the Grant and Lamar areas by 2000 under unrestricted ground-water development for irrigation. Water-level declines over most of the remainder of the study area would be less than 60 feet under continued development and less than 40 feet under no additional development from

1975 to 2000. Ground water in storage would be reduced by about 3.7 percent by 2000 under continued development and by about 2.8 percent by 2000 under no futher development. The analyses also show that the base flow of Frenchman, Stinking Water, and Spring Creeks would be reduced to less than 10 percent of the 1975 values under no further development and eliminated by about 1992 under continued development.

INTRODUCTION

Purpose and Scope

Use of ground water for irrigation in the Upper Republican Natural Resources District (NRD) in southwestern Nebraska has increased rapidly during the past few years, causing water levels to decline and streams to be depleted. Described herein are the hydrogeologic system, changes to that system over the period from 1952 to 1975 caused by agricultural development, and the development and use of mathematical models to assess the effects of future regulated and unregulated private development of irrigation wells.

The Upper Republican Natural Resources District comprises Chase, Dundy, and Perkins Counties in southwestern Nebraska. The study area shown in figure 1 was extended north and west of the NRD boundaries and includes 3,340 mi^2. The Platte and Republican Rivers, which act as hydrologic boundaries, are used as the northern and southern study area boundaries. The study was extended westward into Colorado about 5 miles to ensure continuity of results across the State line.

The economy of southwest Nebraska is agriculturally based. Development of irrigation has stimulated the local economy by inducing many types of agribusinesses to locate there. Concern over the economic consequences of limits to the water supply was one of the motivations for initiating this investigation but is not explicitly addressed herein.

Principal population centers in and adjacent to the study area include Imperial, Grant, Ogallala, and Benkelman, Nebr., and Holyoke and Wray, Colo. Total population of the Nebraska part of the study area as of the 1970 census was 10,606 (Nebraska Natural Resources Commission, 1976).

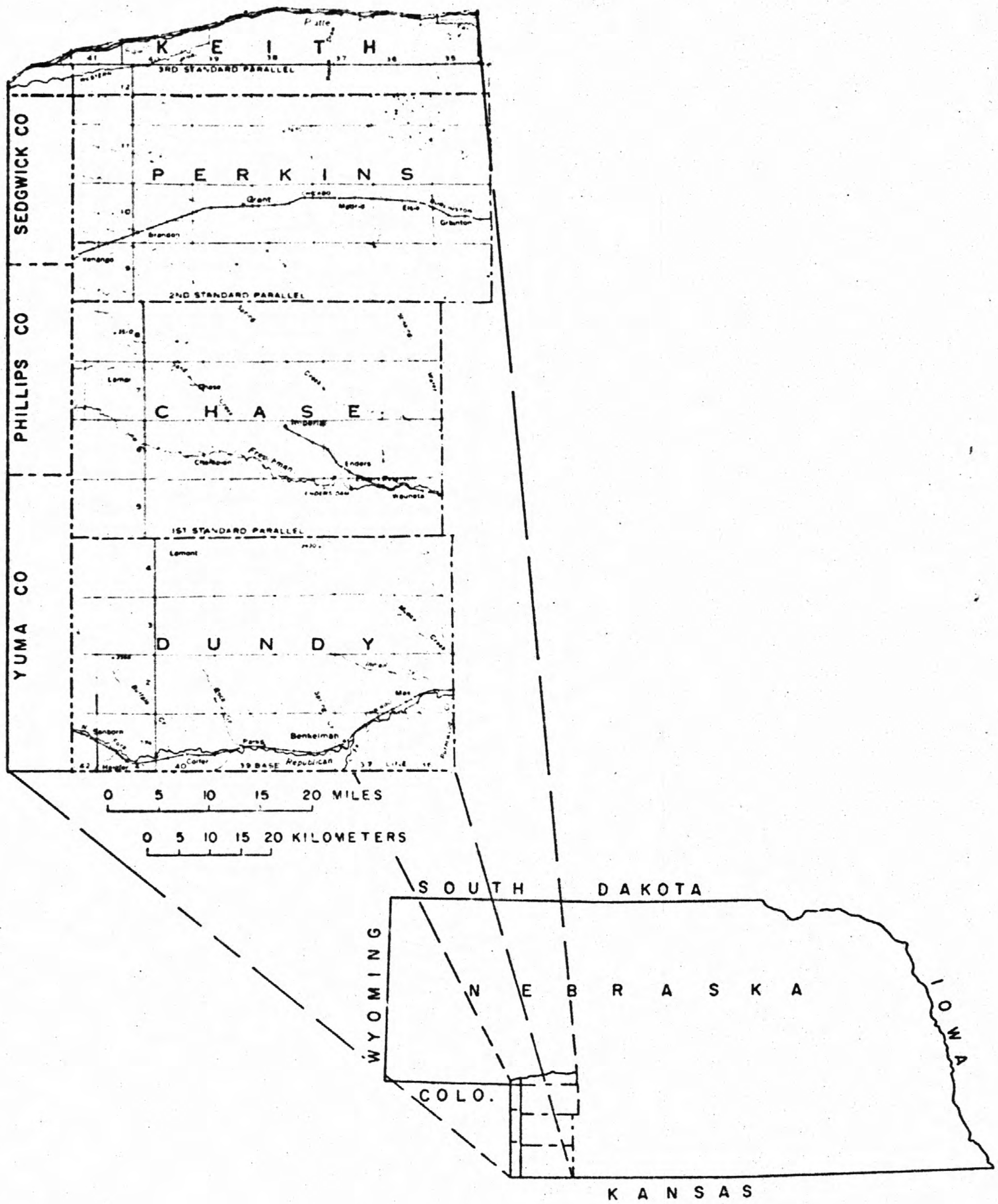


Figure 1.--Location of the study area.

Objective

The objective of this investigation was to develop, test, and apply mathematical models that adequately simulate the operation of the hydrogeologic system to enable evaluation of future effects on the water supply of (1) various levels of unrestricted private irrigation development and (2) management plans determined by the NRD to meet specified objectives under the authority given them by the Nebraska State Legislature.

Acknowledgments

The U.S. Soil Conservation Service provided manpower and materials to map soils and land use. The managers and personnel of the Highline, Midwest, and Southwest Electrical Cooperatives were particularly helpful in compiling power-use data on irrigation wells. Local well drillers contributed valuable test-hole information. The assistance of all of the above as well as the cooperation of local residents during test drilling and water-level and streamflow measurements is gratefully acknowledged.

Physiography and Drainage

The study area lies in the High Plains subsection of the Great Plains physiographic province (Fenneman, 1931). Altitudes range from 2,840 ft where the Republican River crosses the County line in Dundy County, Nebr., to 3,750 ft in the northeastern part of Sedgwick County, Colo. The land surface slopes generally to the east at about 10 ft/mi. The principal physiographic features of the study area are divided into five categories which are described below.

Loess plains and tablelands.--This landform category includes 7.7 percent of the study area, principally in western Chase, Perkins, and Keith Counties and northeastern Dundy and southeastern Chase Counties. Surface drainage is evident, slopes are less than 5 to 7 percent and surface drainageways are broad, grassed swales. Rapid surface runoff is not common.

Rolling uplands.--This physiographic feature occupies 57.4 percent of the study area. Slopes range from 0 to 15 percent. Flat areas within this category contain numerous small closed depressions. Most of these exist in northeastern Perkins and southeastern Keith Counties. The origin of the depressions has been discussed by Darton (1905). These depressions may serve as recharge basins if the bottom sediments are sufficiently permeable. Surface runoff is more rapid than under the loess plains and tablelands. Erosional features comprise broad valleys, swales, and stairstep slumping on valley sides where they are underlain by loess.

Sandhills and interdune valleys.--This landform category includes 28.3 percent of the study area, with the largest areas being found in western Dundy and southwestern Chase Counties. Local relief between valleys and dune crests and saddles is on the order of 50-150 ft. Northwest-trending sand-dune ridges may extend from as much as 10 mi. Slopes range from 0 to as much as 50 percent. Surface drainage is generally poorly defined with closed basins or interior drainage being common. Many interdune valleys contain lakes and swamps that are maintained by ground-water discharge.

Dissected plains and uplands.--This physiographic feature covers 1.7 percent of the study area, principally in eastern Chase and Dundy Counties. Slopes may exceed 50 percent in some areas. Surface drainage is well defined. Surface runoff is rapid and erosional potential is high when precipitation is intense. Local flooding problems in the Wauneta area are the result of direct surface runoff from these areas.

Bottom lands along major watercourses.--This physiographic feature comprises 4.9 percent of the study area along the valleys of Frenchman, Spring, and Stinking Water Creeks and the Republican and South Platte Rivers. Slopes are less than 5 percent. Surface drainage may be altered locally due to changing surface flow regimes of the watercourses.

Land Location System and Altitude Control

This report uses a system of referencing well locations to the U.S. Bureau of Land Management's numbering system as shown in figure 2.

Well numbers are based on the land subdivisions within the U.S. Bureau of Land Management's survey of Nebraska. The numeral preceding the N (north) indicates the township, the numeral preceding the E (east) or W (west) indicates the range, and the numeral preceding the terminal letters indicates the section in which the well is located. The terminal letters denote, in respective order, the quarter section, the quarter-

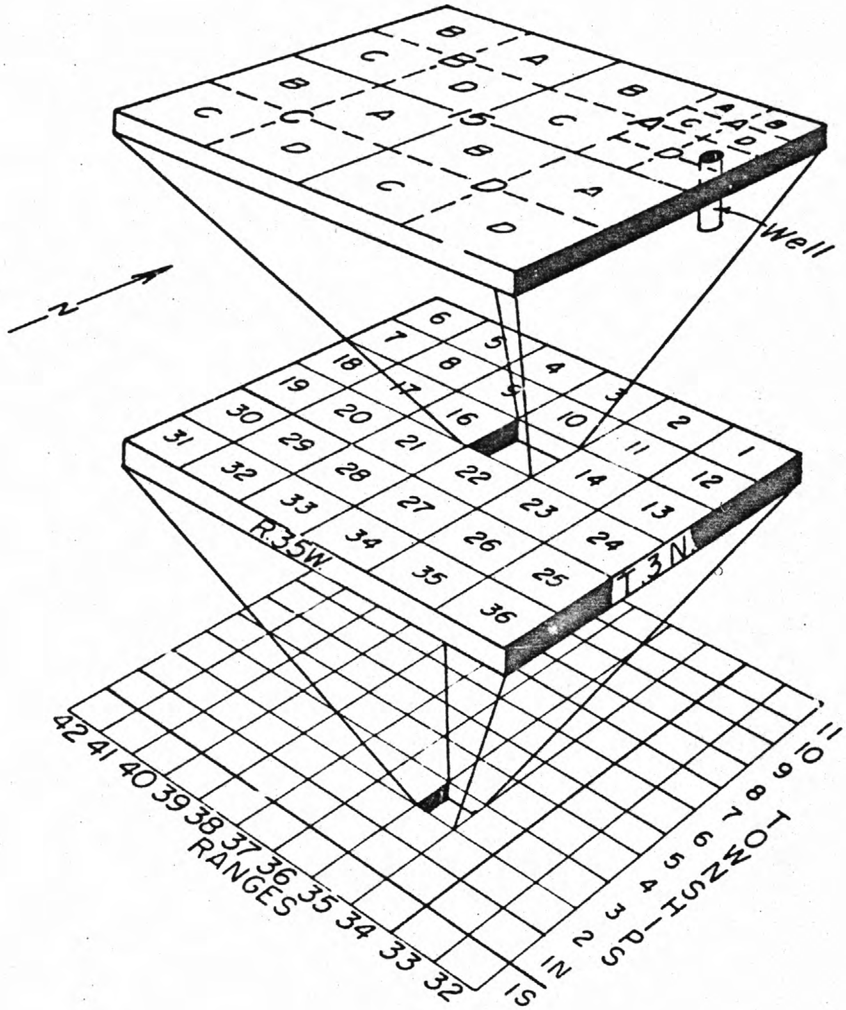


Figure 2.--Well-numbering system.

quarter section, and the quarter-quarter-quarter section and are assigned in counterclockwise direction beginning with A in the northeast corner of each subdivision. If two or more wells are located in the same tract, they are distinguished by appending a sequential digit to the well number. Thus, the second well inventoried in NE¼SE¼NE¼ sec. 15, T. 3 N., R. 35 W. would be assigned the number 3N-35W-15ADA2.

Altitude control for construction of maps of water levels and base of the ground-water system was obtained from two sources. Altitudes for test holes and wells measured prior to 1955 for the Frenchman Creek Basin report (Cardwell and Jenkins, 1963) and for test holes and wells measured in 1975 in the east-central part of Chase County were determined by instrument leveling. Altitudes of other data collection points, including most wells and test holes measured between 1973 and 1975, were determined from 7½-minute series topographic maps prepared by the Topographic Division of the U.S. Geological Survey. These maps show contours of the land surface at an interval of 10 ft with supplemental 5-ft contours. In addition, the surveyed altitude of many wells is given on these maps. Areal coverage of the study area with these maps is shown in figure 3.

Previous Studies

This report draws heavily on hydrologic data and interpretations from many previous investigations covering parts of the area. These studies define conditions that existed prior to and during ground-water development for irrigation. General hydrologic conditions and ground-water availability along river valleys in and adjacent to the area have been described by several authors.

The Republican River valley was discussed by Condra (1907); Waite, Reed, and Jones (1946); Waite and others (1948); and Bradley and Johnson (1957). Hydrology of the Frenchman Creek valley was described by Waite and others (1948) and Bradley and Johnson (1957). Hydrologic conditions in the South Platte River valley were discussed by Bjorklund and Brown (1957). These studies included general descriptions of water levels, aquifer boundaries, and potential well yields.

Areal hydrologic studies that were not confined to river valleys were described by Wenzel and Waite (1941) for Keith County, by Johnson (1960) for the northeastern part of the area, and by Cardwell and Jenkins (1963) for that part of the area lying within the Frenchman Creek basin

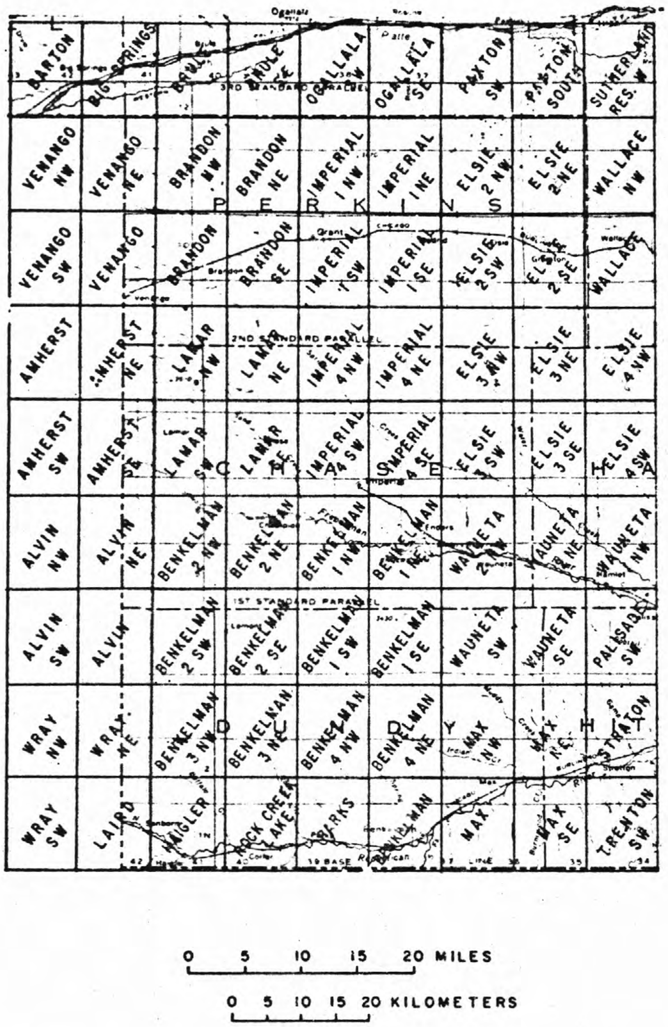


Figure 3.--Topographic mapping of the study area.

above Palisade, Nebr. These studies included information on aquifer boundaries, saturated thickness, water levels, bedrock geology and configuration, and water use. Distribution of ground-water recharge and discharge for the Colorado part of the study area was discussed by Redell (1967).

Development of ground water for irrigation in the area and the effects of this development on the water supply have been described by Cardwell (1953), Cardwell and Jenkins (1963), Boettcher (1966), Leonard and Huntoon (1974), Luckey (1973), Luckey and Hofstra (1974), the U.S. Bureau of Reclamation (1974), and Lappala (1976). Cardwell and Jenkins (1963) included analyses of stream depletion through the year 2000 caused by pumping of ground water in and adjacent to the valleys of Frenchman, Spring, and Stinking Water Creeks. Their predictions were reliable until use of center-pivot irrigation systems became widespread. These systems permit irrigation of land that was not considered irrigable in their stream-depletion analyses.

The future effects on water levels due to pumping for irrigation have been described for that part of the area lying in the Platte River basin in Nebraska by the Missouri River Basin Commission (1975) and for the Colorado part of the area by Luckey (1973), and Luckey and Hofstra (1974). Techniques of simulating the hydrologic system used in this study are essentially identical to those used in the earlier studies. Additional previous studies that served principally as surveys of basic data are referenced, where appropriate, in subsequent sections of this report.

General Methodology

Analysis of the use and management of both ground and surface water for irrigation requires the use of accounting procedures to assess the availability in space and time of water as it moves through the hydrologic cycle (fig. 4). The accounting procedures used for this investigation are referred to as simulation models.

Application of the models is separated into four phases for this study: (1) Describing the operation (or developing conceptual models) of those parts of the hydrogeologic cycle which apply to the problem, (2) formulating the conceptualized models into applicable mathematical equations and a sufficiently accurate method for their solution, (3) testing of the response of the mathematical models against measured

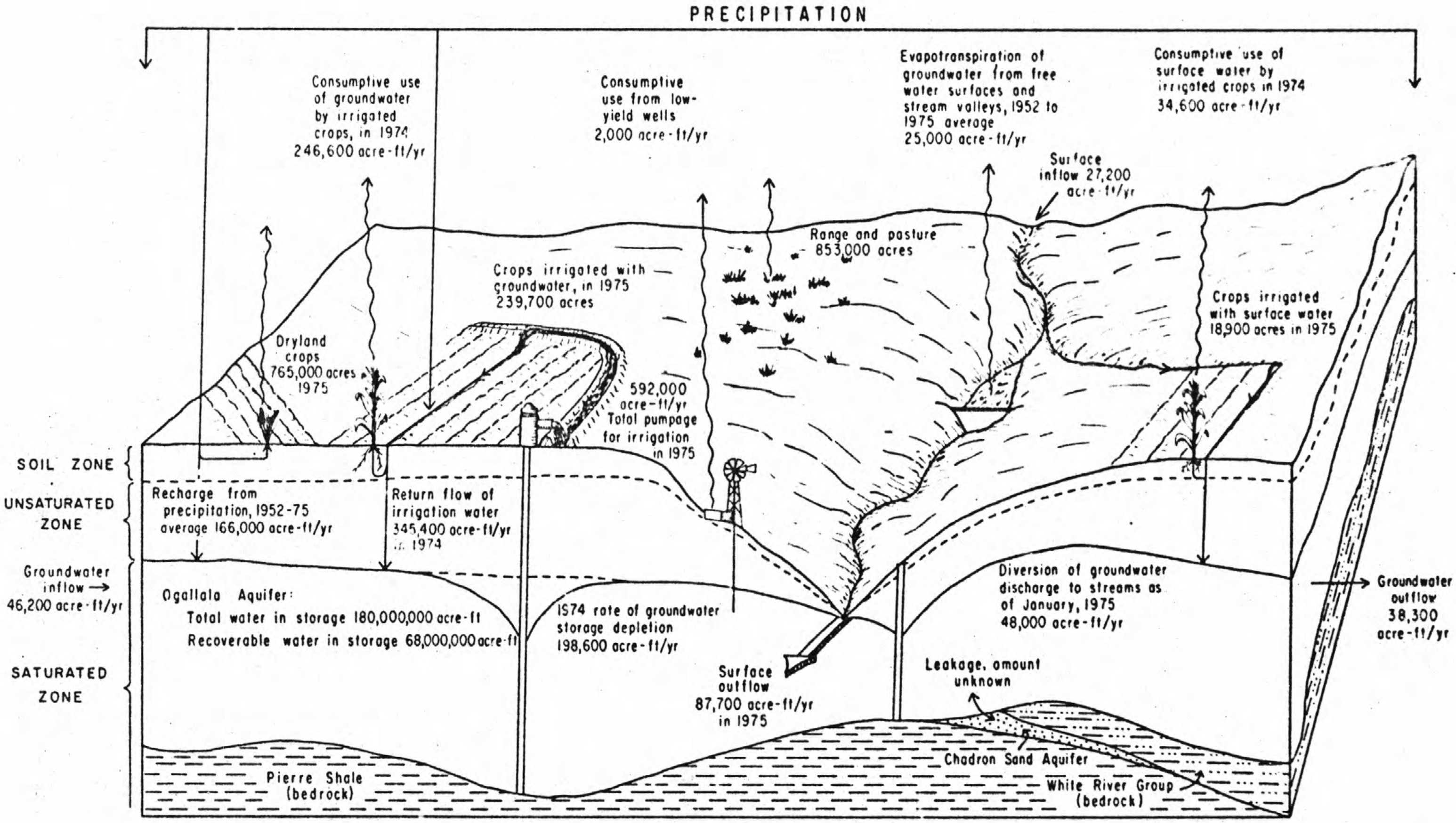


Figure 4.--The conceptual model of the hydrogeologic system.

past response of the simulated system, and (4) applying the tested model to evaluate future response of the system under specified conditions. All four phases require describing operation of the hydrogeologic system over specified time intervals. These intervals are: (1) the pre-irrigation development period, represented by averages over the period from 1925 to 1952; (2) the irrigation development period from 1952 to 1975; and (3) a future analysis period from 1975 to 2000.

Definition of Hydrogeologic Terms

The following concepts and terms used in the report are defined here for simplicity of presentation. Where possible, the definitions are those of current usage of the U.S. Geological Survey. Others are taken from recent hydrologic literature.

An aquifer is a geologic formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs. A strata of low vertical permeability which restricts the vertical movement of water from or to an aquifer is a confining bed. If not overlain by a confining bed, the uppermost aquifer at a given locality is referred to as an unconfined or water-table aquifer.

Base flow of a stream refers to the component of total streamflow attributable to ground-water discharge into the stream channel.

Surface runoff refers to that component of streamflow that enters the stream channel by flowing over the land surface.

Hydraulic head (h) is an expression for the potential energy of a fluid, and when used in discussing isothermal and isohaline fluids occurring in porous media is expressed as:

$$h = \psi + Z,$$

where h = total hydraulic head, L,

ψ = pressure head, L, and is equivalent to P/γ,

wherein P = fluid pressure, $MT^{-2}L^{-1}$, and

γ = fluid specific weight, $MT^{-2}L^{-2}$.

Atmospheric pressure is used as the datum from which P is measured.

Z = elevation head, L, taken relative to some arbitrary horizontal datum, commonly sea level.

Aquifers or parts of aquifers in which the hydraulic head is above the top of the aquifer or the base of the confining bed are said to be confined. The water table is the point near the top of an unconfined aquifer at which the total hydraulic head (h) equals the elevation head (Z), or at which the pressure head (ψ) is zero. Immediately above the water table lies the capillary fringe, in which saturated conditions exist, but in the lower part the pressure is less than atmospheric (Lohman, 1972, p. 2).

The water table is considered the top of the saturated zone in this report. The unsaturated zone extends from the top of the saturated part of the capillary fringe to the land surface. The use of the term unsaturated is a convenience only, as parts of this zone are often temporarily saturated.

Porosity (ε) is the volume of pore space per unit volume of a porous medium.

Storage capacity (S) of a porous media is a measure of the volume of fluid that can be added to or taken from storage in the pore space of a porous medium. It is a function of volumetric moisture content (θ) in the unsaturated zone; a function of the drainable porosity, fluid, and aquifer compressibility in unconfined aquifers; and a function of fluid and aquifer compressibility in confined aquifers. The drainable porosity is commonly referred to as the specific yield (S_y) and is the the difference between total porosity and the volume fraction of fluid retained in the pore spaces against the forces of gravity (specific retention).

Specific storage (S_s) is the volume of water derived from fluid expansion and aquifer compressibility per unit volume of aquifer for a unit decline in hydraulic head. The artesian storage coefficient (S) is the product of specific storage and aquifer thickness and thus is a measure of the water derived from storage over the entire aquifer thickness.

Field capacity (F_c) is an arbitrary value of specific retention corresponding to a pressure head of minus one-third of an atmosphere or 11.31 ft of water (Linsley, Kohler, and Paulhus, 1958). At moisture contents less than field capacity, no significant drainage of water by gravity is assumed to occur.

Wilting point (W_p) is an arbitrary value of moisture content corresponding to a pressure head of minus 15 atmospheres or 508.9 ft of water (Linsley, Kohler, and Paulhus, 1958). Permanent wilting and consequent cessation of transpiration by most plants occurs at moisture contents less than the wilting point.

Hydraulic conductivity (K) is a measure of the volume of fluid that will move in unit time under a unit gradient in total hydraulic head through a unit area orthogonal to the direction of flow. It is a property of both the porous media and the fluid and has the dimensions of velocity (LT^{-1}). In the unsaturated zone, hydraulic conductivity is also a function of moisture content.

Transmissivity (T) is the product of the thickness of the saturated zone and the hydraulic conductivity of that zone and has dimensions L^2T^{-1}.

The soil or root zone is that part of the unsaturated zone in which most active root growth and moisture extraction occur.

Evapotranspiration (ET) is the combined process of evaporation from free water and bare soil surfaces and transpiration by plants. Evapotranspiration is a consumptive use of water at a given locality because the water vapor is usually removed from the immediate area by atmospheric circulation.

Potential evapotranspiration (PET) is defined by Linsley, Kohler, and Paulhus (1958) as the amount of water that would be evaporated from bare soil and transpired by plants if neither is under moisture stress. That is, water availability is not a constraint on the amount of potential evapotranspiration. As such, it is a measure both of the energy available to drive the evaporative processes and of the efficiency of the mechanisms that remove evaporated water molecules from the immediate vicinity of an evaporative surface such as leaf stomata or bare soil.

Consumptive-irrigation requirement (CIR) is the amount of water required to meet ET demand and to maintain soil moisture at an arbitrary level after soil moisture and infiltrated precipitation have been drawn upon.

Infiltration (I) is that part of precipitation and applied surface water that enters the soil zone.

Deep percolation is water that leaves the soil zone into the underlying part of the unsaturated zone.

Recharge to an aquifer as used in this report is that part of deep percolation that reaches the aquifer. It also includes seepage from free water sources such as streams and lakes that are in direct hydraulic connection with the aquifer.

Discharge from an aquifer is the water that is transferred from the aquifer to the unsaturated zone or to the land surface. Natural discharge includes seepage to lakes, springs, and streams. Artificial ground-water discharge includes abstraction of ground water by wells, drains constructed to intercept the aquifer, and evaporation from lakes or ponds excavated below the water table.

Leakage is the transfer of water between aquifers, or in some usage in this report between surface-water sources and an aquifer.

Underflow of ground water is lateral movement of ground water across a specified boundary.

DESCRIPTION OF THE HYDROGEOLOGIC SYSTEM

Figure 4 shows the movement of water through the hydrogeologic system applicable to the study area. Mathematical models which rigorously consider all parts of this system are precluded by limited data availability, by limited computational facilities required for solution of the mathematically formulated problem, and in some cases by a poor understanding of the physical phenomena involved. Consequently, most simulation studies of this type have been designed to describe parts of the hydrogeologic cycle with linkages to other components being specified as inputs or boundary conditions to the modeled segment. Recent examples of these are described by Pikul and others (1974) and Knapp and others (1975). A similar procedure is followed in this study by separating the hydrogeologic system shown in figure 4 into the following four segments: (1) the surface-water system, (2) the soil zone, (3) the lower unsaturated zone, and (4) the saturated zone.

Surface-Water System

The surface-water system in the study area comprises natural streams, canals, and both on-channel and off-channel storage reservoirs.

Streamflow

Most of the streamflow leaving the study area originates as ground-water discharge. Twenty gaging stations have been operated in and adjacent to the study area since 1894 by the U.S. Geological Survey, the Nebraska Department of Water Resources, and the U.S. Bureau of Reclamation; eleven of these are currently operating. The character of flow at selected stations in the study area is shown on plate 1. Also included on plate 1 is a series of regularly timed measurements on Spring and Horse Creeks, tributaries to the Republican River. Although these two streams have not been gaged continuously, the magnitude of their flow regimen is shown by the graphs. The flow of the remainder of the streams is typical of streams draining ground-water systems in western Nebraska. A fairly uniform year-to-year peak is followed by a recession caused by increased ET. This pattern is particularly evident on the records for Frenchman Creek near Imperial, Buffalo Creek near Haigler, and Rock Creek near Parks. The records for Stinking Water Creek near Wauneta exhibit a somewhat more flashy character. Significant parts of the drainage of this stream comprise rough topography and silty soils. This flashy runoff has resulted in occasional flooding of the town of Wauneta as reported by the Upper Republican Natural Resources District (1976). However, a significant portion of the flow of both Stinking Water Creek and its tributary Spring Creek, in Chase County, is ground-water discharge. The annual peaks of Frenchman Creek near Enders and Frenchman Creek near Palisade shown on plate 1 are reservoir releases from Enders Reservoir.

Much of the runoff in the upper ephemeral reaches of the stream valleys outside of the area of rough topography seeps into the streambed before reaching the point of effluence of these streams. Some of the stream valleys are discontinuous, being blocked by sand dunes. The end points of these valleys often contain shallow lagoons or playas that serve as recharge basins if their bottoms are permeable. Most of these basins occur in the sand-dune areas. Some of the playas such as those in northeastern Perkins and southeastern Keith Counties have relatively impermeable bottoms, and the water impounded in them evaporates.

In addition to continuous recorders and regularly timed measurements, several sets of seepage measurements have been made on perennial streams in the study area. Measurements made on the Frenchman, Spring, and Stinking Water Creeks in October 1952 (pl. 2) by Cardwell and Jenkins (1963) are used in this study as being representative of the ground-water discharge distribution along the stream valleys under preirrigation development conditions.

Discharge measurements were made in May, July, and October 1975 on the above three streams as well as on Indian and Rock Creeks in Dundy County (pl. 2). Measurements were also made at about the location of the aquifer-bedrock contact on Buffalo, Horse, Spring, and Muddy Creeks in Dundy County (table 1). The October 1952 and November 1975 measurements on Frenchman, Spring, and Stinking Water Creeks in Chase County were used as one measure of the change in ground-water discharge to these streams. No measurements were made during the 1950's for the tributaries to the north side of the Republican River. Conditions as determined in 1975 probably are representative of the preirrigation development conditions on Indian and Buffalo Creeks as withdrawal of ground water adjacent to these streams has been minimal. However, flows measured in 1975 on Spring, Horse, and Rock Creeks probably reflect at least some depletion due to ground-water withdrawals. Depletions of spring discharge into Rock Creek during the irrigation season have become a cause for concern by the Nebraska Game and Parks Commission which operates a trout hatchery on this stream. A detailed study by D. Thomssen (1976) documents the hydrogeologic conditions around the hatchery and the stream depletions that have occurred.

Table 1.--Discharge measurements, Republican River tributaries, 1975

Stream	Location of measurement	Date	Discharge (ft^3/s)
Buffalo Creek	2N-41W-26DCC	05/19/75	9.0
		07/15/75	7.4
		10/06/75	7.2
Horse Creek	1N-38W-2CCD	05/19/75	.11
		07/14/75	.02
		10/06/75	.20
Spring Creek	1N-38W-2DDC	05/20/75	.55
		07/14/75	.19
		10/06/75	.28
	2N-38W-34ABA	05/20/75	.07
		07/14/75	1/ .02
		10/06/75	.06
Muddy Creek	2N-36W-1CCB	05/19/75	.52
		07/14/75	.23
		10/06/75	.20
	3N-36W-27BCC	05/19/75	.47
		07/14/75	.43
		10/06/75	.27

1/ Estimated.

The discharge of the Republican and South Platte Rivers bounding the study area on the south and north is controlled by surface runoff and reservoirs in Colorado, as well as ground-water discharge in both Colorado and Nebraska. Both streams receive ground-water discharge from the study area, but no seepage measurements have been made on either river. The amount of base flow or ground-water discharge to these rivers not entering through tributaries was determined by yearly reach-gain computations. Inflow to the South Platte reach was the measured flow at Julesburg, Colo., and outflow was the measured flow at Paxton, Nebr., reported by the U.S. Geological Survey (annual). The period of record used was 1940-69. Annual diversions used in the computations were those of the Western Canal (1940-69) and the Sutherland Supply Canal (1947-69) reported by the Nebraska Department of Water Resources (annual). The results of these computations indicated an average reach gain from ground-water discharge and unmeasured tributary inflow of 33 ft^3/s as shown in plate 3. Since canal returns and tributary inflow are not measured, the ground-water discharge to this reach was computed as the average of October, November, and December flows prior to 1967. This average is 12.8 ft^3/s. By examining water-table contour maps (Missouri River Basin Commission, 1975) north of the South Platte River, 80 percent of the reach gain or about 10 ft^3/s was assumed to originate as ground-water discharge from the study area.

A similar reach-gain study was made of the North Fork of the Republican River between the Colorado-Nebraska State line and the Republican River at Benkelman, Nebr., for the 1948-75 period. Gaged tributary inflow consisted of the Arikaree River near Haigler, Buffalo Creek near Haigler, Rock Creek near Parks, partial records of Horse Creek near Max, and Spring Creek near Max. This study showed a net loss of streamflow in the reach from the State line to Benkelman (pl. 3). The losses averaged 5.6 ft^3/s before 1968 and 8.7 ft^3/s since 1968. The losses in streamflow are due to ground-water pumpage in the Republican River valley and evapotranspiration by phreatophytes, principally Cottonwood (Populus). Phreatophytes along the north side of the valley consume at least 10,000 acre-ft/yr of ground-water discharge into the Republican River valley from springs and seeps along the north side of the valley, based upon 1975 hydraulic gradients and aquifer transmissivities in this area.

Streamflow has been reduced in Frenchman and Stinking Water Creeks due to diversion by wells of ground-water discharge to these streams. Single mass curves of monthly flows of these streams (pl. 3) show a continuing break in slope beginning about 1967. This is coincident with a significant increase in the development of ground water for irrigation. Single mass curves of streams such as Rock Creek (pl. 3) whose ground-water discharge areas have not been developed with irrigation wells until 1974 and 1975 show no such break coinciding with the rapid increase in the use of ground water for irrigation. The significant break in slope of the mass curve of Buffalo Creek in 1954 corresponds to the installation of a concrete control structure at the gage. The breaks in slope of the mass curve for Frenchman Creek near Champion in 1935 (pl. 3) are thought to be the result of changing operations on three small upstream storage reservoirs.

Reduced precipitation is not a cause of the depletion of the streams since 1967. This conclusion is based upon the following reasoning:

(1) The relationship between the flow regimen of two streams draining basins of similar hydrologic and geologic character under similar climatic conditions should remain constant.

(2) Long-term changes in precipitation over the basin would be widespread, affecting one station as much as the next if the basins are not too distant from one another. Statistical analyses of monthly precipitation data for the study area support this contention.

(3) The relationship between total annual precipitation and total annual streamflow should be fairly constant for basins whose discharge regimen is controlled by ground-water systems and therefore shows a uniform damped response to precipitation.

Double mass analyses (Searcy and Hardison, 1960) were made using Stinking Water Creek near Palisade (a depleted stream) against Rock Creek at Parks (a nondepleted stream) and Frenchman Creek near Imperial (depleted) against Rock Creek at Parks (pl. 4). Both these analyses show the same break in slope about 1967 as did the single mass curves (pl. 3). A double mass curve of monthly streamflow of Frenchman Creek near Imperial against precipitation at Imperial (pl. 4) shows the same break in slope. The relationship between annual precipitation and annual streamflow from 1941 to 1967 in the Frenchman Creek basin shows considerable scatter (pl. 4), (coefficient of determination = 0.35). However, the streamflow values for the years 1968 to 1975 are significantly less than those for the pre-1967 period for the same precipitation. These analyses lend further support to the conclusion that reduced precipitation has not been the cause of depletions to Frenchman and Stinking Water Creeks. The only remaining reasonable cause of the depletions is capture by irrigation wells of water normally discharged to streams.

Depletion of the Republican River between the State line and Benkelman due to ground-water withdrawals in the valley is apparent beginning about 1968. This is shown by an increase in reach loss using mass curve analysis (pl. 3). Extensive irrigation development on the North Fork of the Republican River in Yuma County, Colo., undoubtedly will increase streamflow depletions in the near future. Depletions to the South Platte River between Julesburg, Colo., and Paxton, Nebr., may also have occurred since 1969. However, a mass curve analysis of the computed reach gain (pl. 3) shows no detectable reduction in flow between 1940 and 1969. Mass curve analyses were done only through 1969 because the gaging station at Paxton was discontinued in 1970.

Lakes and reservoirs

The total volume of water stored in lakes and reservoirs is not known. However, surface storage rights have been granted for 50,150 acre-ft by the Nebraska Department of Water Resources. With the exception of Enders reservoir, surface impoundments on perennial streams generally have a storage capacity of less than 1,000 acre-ft. Numerous farm ponds and natural closed depressions on ephemeral streams also exist, but their storage capacity is unknown. No permanent natural lakes exist in the study area.

Evaluation of seepage losses from surface-water reservoirs was possible only for Enders Reservoir owing to lack of measured inflow and outflow on other impoundments. The losses from Enders Reservoir were evaluated by conducting a monthly inflow-outflow study from 1949 to 1975 using the following equation:

$$Q_S = -Q_I + Q_E + (E-P) \cdot A \pm \Delta S/\Delta t, \tag{1}$$

where

Q_S = reservoir seepage, L^3T^{-1} (positive = ground-water recharge),

Q_I = average monthly reservoir inflow at the gaging station near Imperial, L^3T^{-1},

Q_E = average monthly reservoir release as measured at the gage near Enders, L^3T^{-1},

P = monthly precipitation at Enders, LT^{-1},

E = monthly lake evaporation at Enders, LT^{-1},

A = end-of-month reservoir area, L^2,

ΔS = monthly change in storage, L^3,

Δt = time increment of study (= 1 month), T.

Seepage into and out of the reservoir based on this equation is shown in figure 5. From this study it is concluded that significant reservoir losses to the aquifer occurred during the first few months of operation, but since 1961 annual lowering of reservoir stage regains all of the water lost to bank storage during the period of reservoir filling plus additional inflow, resulting in a net gain from ground water.

The approximate gain in this reach before construction of Enders Reservoir, based upon estimated hydraulic-head gradients and aquifer transmissivities, was about 5,800 acre-ft/yr. The average annual gain over the 1961-74 period was about 4,800 acre-ft/yr (fig. 5). Thus, although natural ground-water discharge has been reduced about 1,000 acre-ft/yr in the reservoir reach, the reach still gains. A similar surface water-ground water interaction exists in the valley of Frenchman Creek between Enders Dam and the eastern Chase County line. This reach gains ground water at stream stages corresponding to reservoir releases of less than about 200 ft^3/s and loses to the aquifer at stages corresponding to discharges more than 200 ft^3/s, (U.S. Bureau of Reclamation, 1974). The loss in the reach between Enders Dam and Palisade ranges from about 20 ft^3/s to 40 ft^3/s during peak releases shown on plate 4. Water lost to bank storage during the high stages returns to the stream after releases are stopped (U.S. Bureau of Reclamation 1974). Other reservoirs constructed in the valleys of streams which gain from ground-water discharge should exhibit seepage patterns similar to Enders Reservoir.

Total evaporation from free water surfaces amounts to about 10,500 acre-ft/yr in the study area. This value was derived using corrected pan evaporation records from 1950 to 1972 at Enders Reservoir (U.S. Bureau of Reclamation, 1974). The area of lakes and reservoirs with evaporating free water surfaces is estimated at 2,500 acres, with 1,700 acres accounted for by Enders Reservoir (U.S. Bureau of Reclamation, 1974).

Soil Zone

The segment of the hydrogeologic cycle comprising the land surface, plant community, and the plant root system will be referred to as simply the soil zone. Movement of water within the soil zone is described within the framework of the soils and vegetation. Water enters the soil zone from precipitation and any applied irrigation water and is withdrawn by evapotranspiration, direct surface runoff, and percolation to the underlying unsaturated zone. Input and output of water as related to the components of the soil zone are described in the following sections.

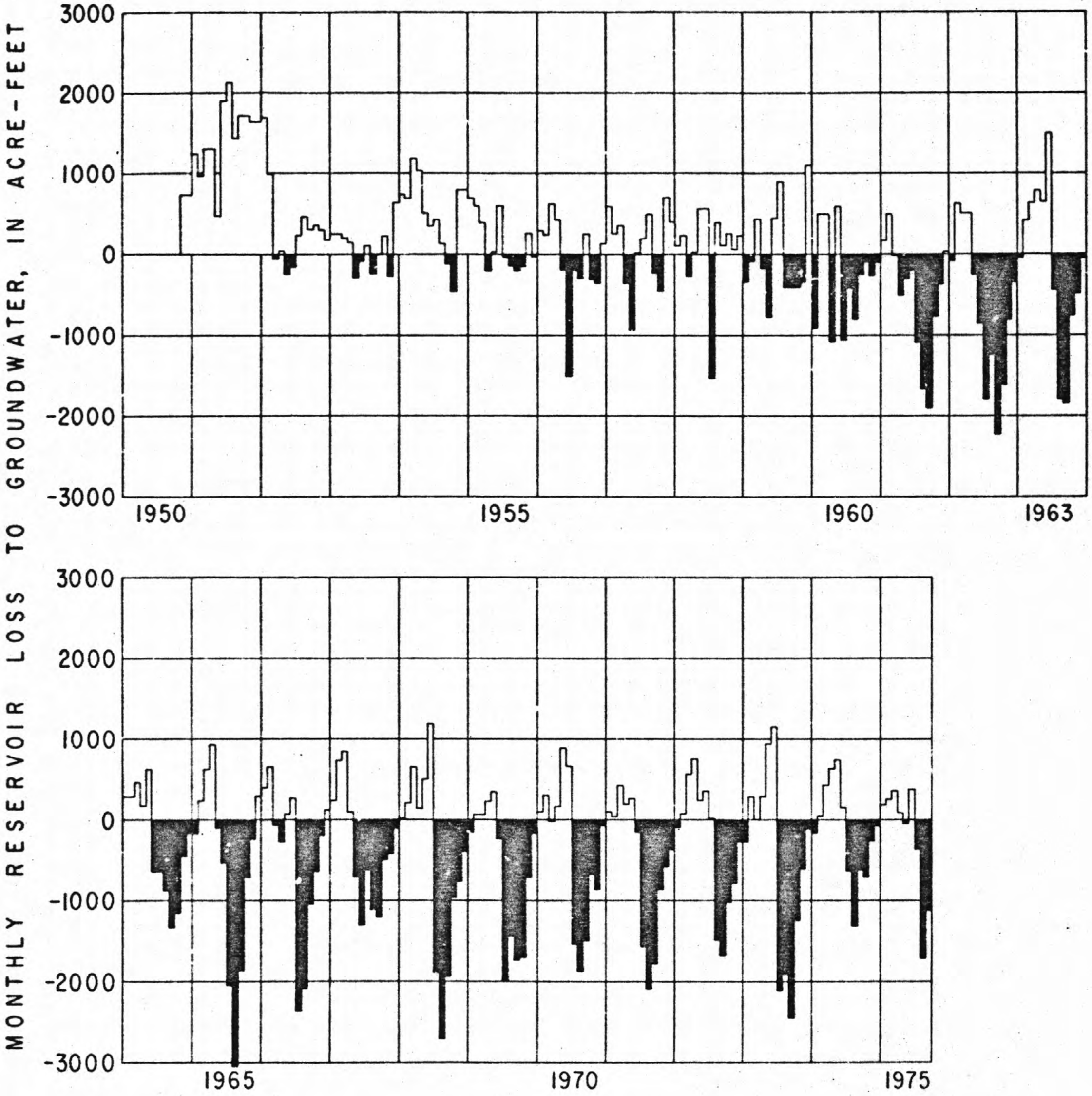


Figure 5.--Monthly seepage losses and gains for Enders Reservoir, 1949 through 1975.

Soils

The soil matrix acts as a storage reservoir of water from which releases are made to plant roots and to transpiration, bare-soil evaporation, and downward movement or deep percolation to the lower unsaturated zone.

The areal distribution of soil associations for this study was adapted from mapping by Bill McKenzie (retired) of the U.S. Soil Conservation Service (written commun., 1976) at a scale of 1:250,000 for Chase, Dundy, and Perkins Counties, supplemented by the author's interpretation of the 1926 survey in Keith County and recent surveys in Phillips, Sedgwick, and Yuma Counties, Colo. The mapped units were further lumped into nine categories based on slope, texture, saturated vertical permeability, and moisture-holding capacity. The resultant distribution is shown in figure 6 and the included soil series, textural classes, and hydrologic properties are summarized in table 2. Detailed descriptions of the soil associations can be found in the Dundy County Soil Survey by Brubacher and Moore (1971) and in a report by Elder (1969). Descriptions of individual soil series and their agricultural properties used later in this report are available in published form from the Soil Conservation Service.

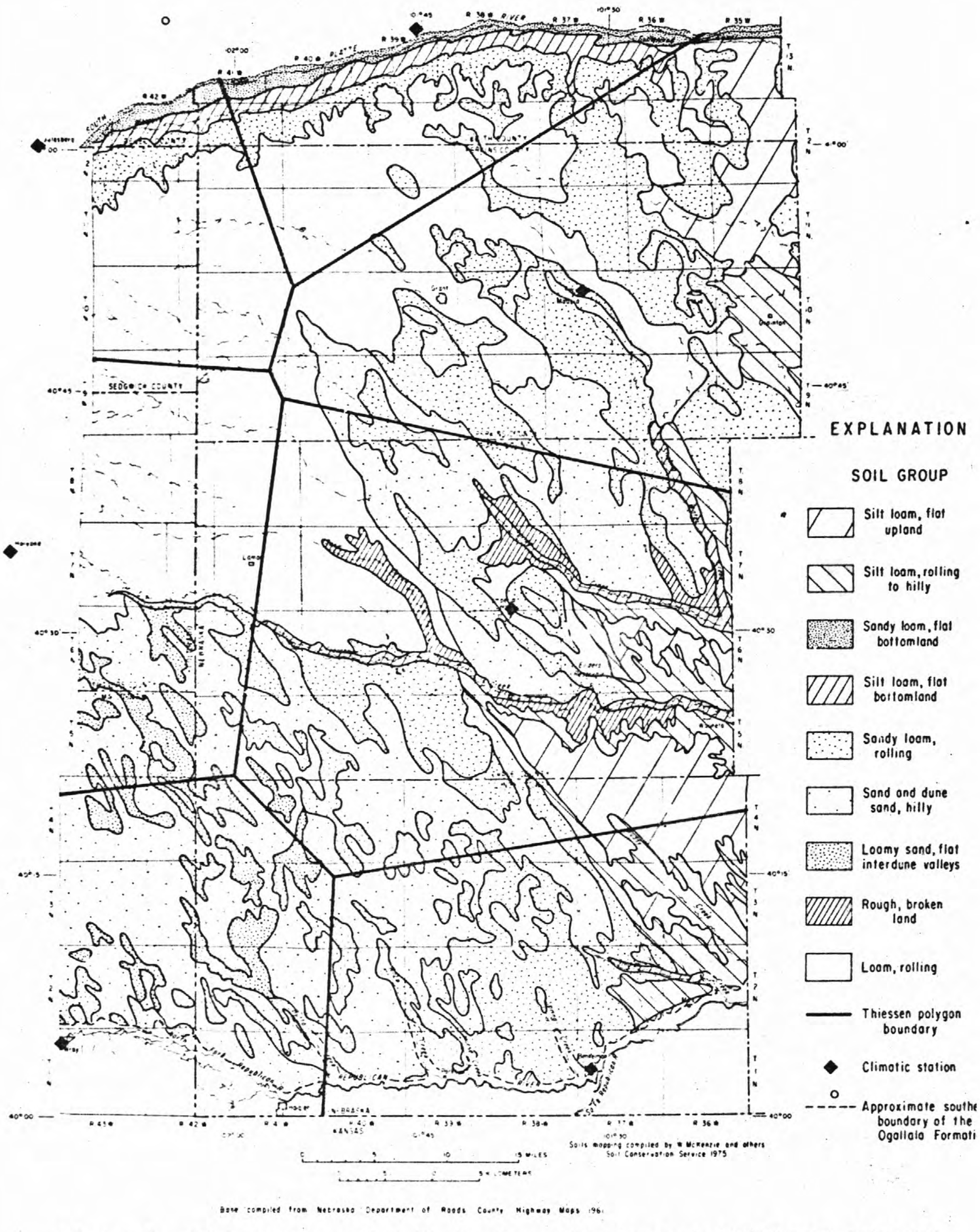


Figure 6.--Distribution of soil-physiographic types and Thiessen polygons used to distribute point climatic data.

Table 2.--Soil-topographic groups and pertinent hydrologic properties

Soil-topographic group	Included soil series	Textural class	Slope range (percent)	Available moisture Min. (dimensionless)	Available moisture Max. (dimensionless)	Available moisture Avg. (dimensionless)	Saturated vertical hydraulic conductivity (ft/d)	Acreage	Percent of area
Silt-loam, flat, upland	Keith, Kuma, Goshen	Silt-loam	0-5	0.16	0.24	0.21	0.4 - 4.0	161,920	7.7
Silt-loan, rolling to hilly	Colby,Ulysses, Keith	Silt-loam	0-30	.17	.24	.21	1.2 - 4.0	140,800	6.7
Sandy loam, flat bottom land	Bankard, Haverson, Glenberg	Sandy-loam	0-6	----	----	.17	----------	41,600	2.0
Silt-loam, flat, bottom land	Bridgeport, McCook	Silt-loam	0-7	.16	.22	.19	1.2 - 4.0	61,440	2.9
Sandy loam, rolling	Jayem, Haxtun, Rosebud	Sandy loam	0-20	.07	.20	.14	.4 - 12.0	525,440	25.0
Sand and dune sand, hilly	Valent, Valentine, Tassel	Fine sand	0-50	.05	.18	.08	4.0 - 40.0	532,480	25.3
Loamy sand, flat interdune valleys	Elsmere-Dailey	Loamy sand	0-4	.06	.17	.10	4.0 - 40.0	63,360	3.0
Rough broken land	Canyon	Loam	0-50	.13	.22	.18	1.2 - 12.0	35,840	1.7
Loam, rolling	Rosebud, Alliance	Loam	0-20	.11	.24	.19	0.4 - 4.0	540,160	25.7

Vegetation

The vegetative cover, both native and cultivated, is hydrologically important to this study because plants are the principal users of stored soil moisture. Also, the consumptive-irrigation requirement (CIR) of irrigated crops constitutes the major use of ground water.

The native vegetation (ca. 1850) in the study area was mapped by R. B. Kaul (1975). A part of this map and associated explanation is reproduced in figure 7. The sandsage prairie covers the largest area of remaining native vegetation. Most of the areas occupied by mixed prairie, flood-plain prairie-forest, and short-grass prairie have been cultivated since the turn of the century.

The cultivated area has remained fairly constant in the study area since at least 1950, as indicated by data on harvested crops compiled by the Nebraska Department of Agriculture (1941-74) for Chase, Dundy, and Perkins Counties (fig. 8). Yearly changes in crop types harvested reflect crop rotations and market conditions. Figure 8 shows that although the area under irrigated row crops and alfalfa increased, the total acreage in these crops did not change significantly until about 1973-74. This implies, at least on a countywide basis, that much of the irrigation development occurred on cropland which had grown the same crops under dryland conditions. Irrigated cropland increased only in the last 2 to 3 years due to cultivation of range or pastureland.

The distribution of both irrigated and dryland crops as of May 1975 was mapped for this project from color infrared imagery at a scale of 1:126,720 by David Loges of the Conservation and Survey Division of the University of Nebraska (written commun., 1975). Ten categories were mapped and aggregated into the following five groups based upon assumed similarities in water requirements:

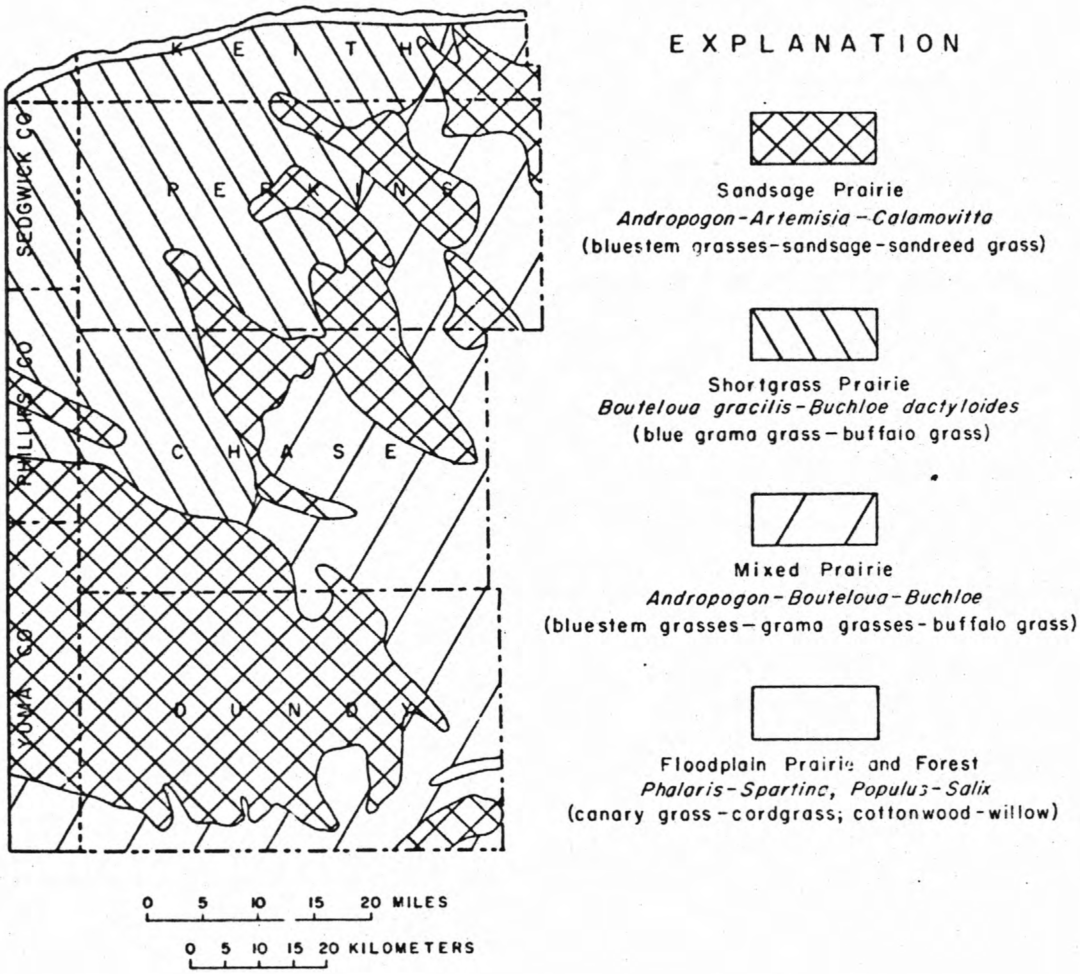


Figure 7.--Native vegetation complexes, circa 1850, after R. B. Kaul (1975).

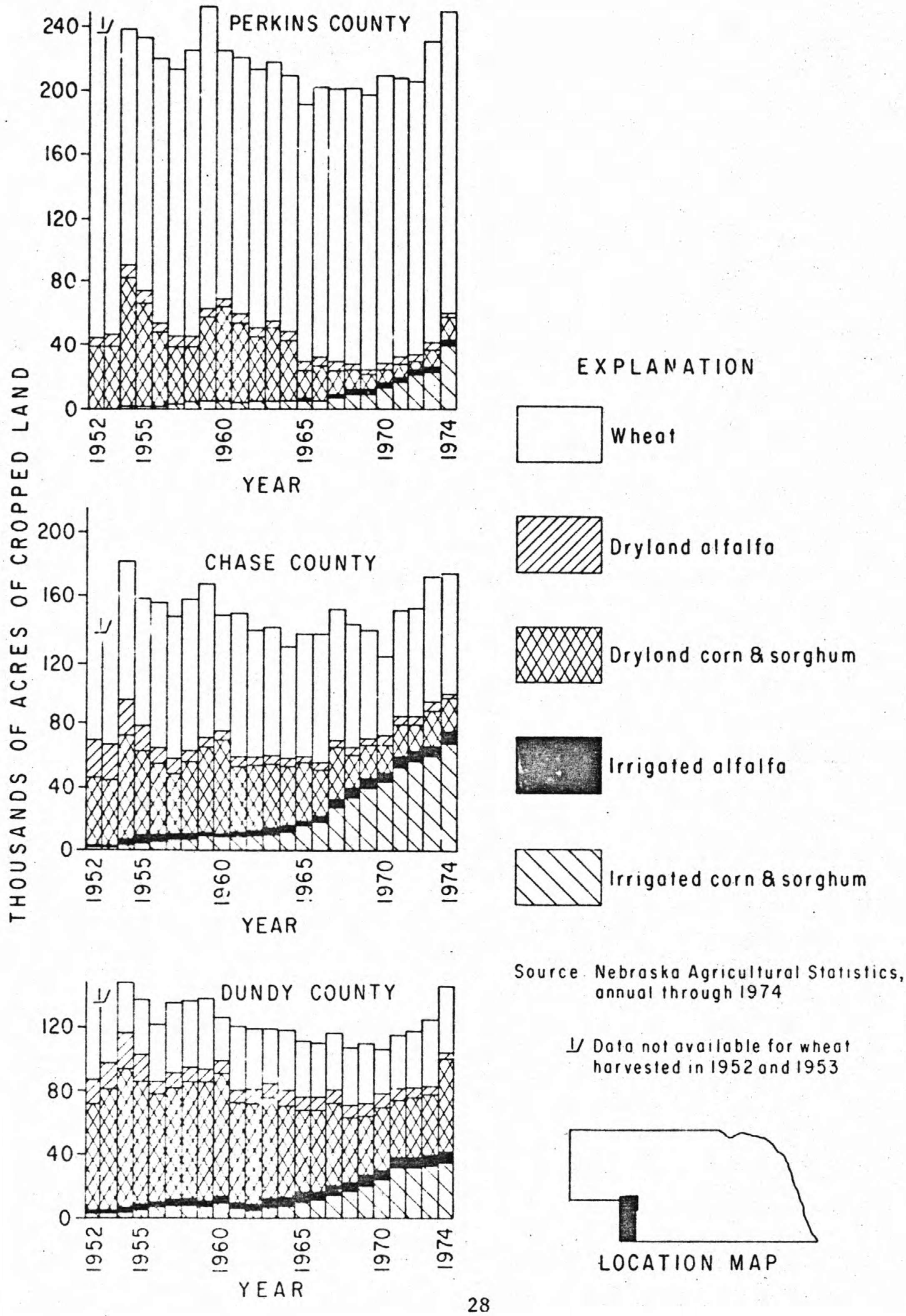


Figure 8.--Annual harvested crops for Chase, Dundy, and Perkins Counties, 1952-74.

Group	Mapped category
Forage crops	Dryland alfalfa Irrigated alfalfa Irrigated grass
Row crops	Irrrigated row crops (principally corn, sugar beets, and dry edible beans)
Small grains	Wheat Stripcropped wheat Fallow land
Pastureland	Range and pasture Range and pasture previously farmed
Noncultivatable	Rough, broken land (principally pasture and range)

Distribution of the five vegetative groups is shown on plate 5 and the area of each type is summarized by counties in table 3. Owing to the unavailability of imagery taken at more than one time of year, some errors are inherent in the data presented in plate 5. The principal error is due to the difficulty in distinguishing between dryland row crops and fallow land during May. This error is probably largest in Chase County and least in Perkins and Keith Counties, Nebr.

The 1975 distribution of vegetation groups shown on plate 5 was considered representative of the 1952-75 period and was used in the study to determine historical water requirements. This assumption was necessary because the 1975 inventory is the only detailed, consistent source of crop distributions. This assumption is considered reasonable because figure 8 shows little change in crop type since 1952, the period covering irrigation development; it simply indicates a replacement of dryland crops with the same irrigated crops.

Table 3.--Crop type inventory and irrigated land summary
Determined from color infrared imagery

Land-use category	Chase County Acres	Chase County Per-cent	Dundy County Acres	Dundy County Per-cent	Perkins County Acres	Perkins County Per-cent	Keith County[1/] Acres	Keith County[1/] Per-cent
All crops								
Row crops	92,490	16.6	52,810	9.1	48,760	8.8	28,060	15.2
Forage crops	16,160	2.9	13,350	2.3	13,300	2.4	6,650	3.6
Small grains	162,700	29.2	121,290	20.9	365,710	66.0	101,910	55.2
Pastureland	230,120	41.3	298,870	51.5	125,230	22.6	41,170	22.3
Rough broken land	55,720	10.0	94,010	16.2	1,110	.2	6,830	3.7
Total	557,190		580,330		554,110		184,620	
Irrigated crops only								
Total	106,980	19.2	60,350	10.4	56,520	10.2	34,710	18.8
	[2/] (81,300)		[2/] (40,930)		[2/] (47,020)			

[1/] Keith County, south of South Platte River only.

[2/] Reported by Nebraska Department of Agriculture, 1974.

Input and output to the soil zone

Water enters the soil zone from precipitation and applied irrigation water and leaves via evapotranspiration, surface runoff, and deep percolation.

Precipitation.--The study area is characterized by a semiarid climate where potential evapotranspiration exceeds precipitation (fig. 9). Most precipitation is received during the late spring and early summer. Precipitation in winter months is in the form of snow, often accompanied by high winds.

The precipitation falling on the study area is not evenly distributed in time nor in space. The following seven weather stations which have reasonably complete precipitation records (National Oceanic and Atmospheric Administration, annual) since 1925 were used for this study: Benkelman, Imperial, Madrid, and Ogallala, Nebr., and Holyoke, Julesburg, and Wray, Colo. Monthly precipitation records for these stations from January 1949 through June 1975 (330 months) were analyzed for both spatial and temporal homogeneity using the following statistical tests by J. T. Dugan, U.S. Geological Survey (written commun., 1975): (1) Simple linear correlation, (2) stepwise forward multiple regression, (3) standard and paired t-tests, and (4) analysis of variance with Duncan's multiple range test (Nie, et al., 1975, and Dixon, 1971).

These tests were used to evaluate the potential for grouping stations to represent the monthly precipitation over the study area. Simple linear and stepwise forward regression showed that records from all stations except Ogallala and Julesburg were strongly related to each other and could be grouped when the records as a whole were considered. The Ogallala and Julesburg stations lie in the South Platte River valley which is more subject to the influences of cold air drainage from Colorado on precipitation patterns than are the remaining stations. This test was not considered sufficient proof to establish the homogeneity or heterogeneity of the monthly records.

A students' t-test was run between all stations. This test compares records as a group with no pairing of individual monthly values. The results from this test showed no differences between pairs of stations at the 0.01, 0.05, or 0.10 significance levels.

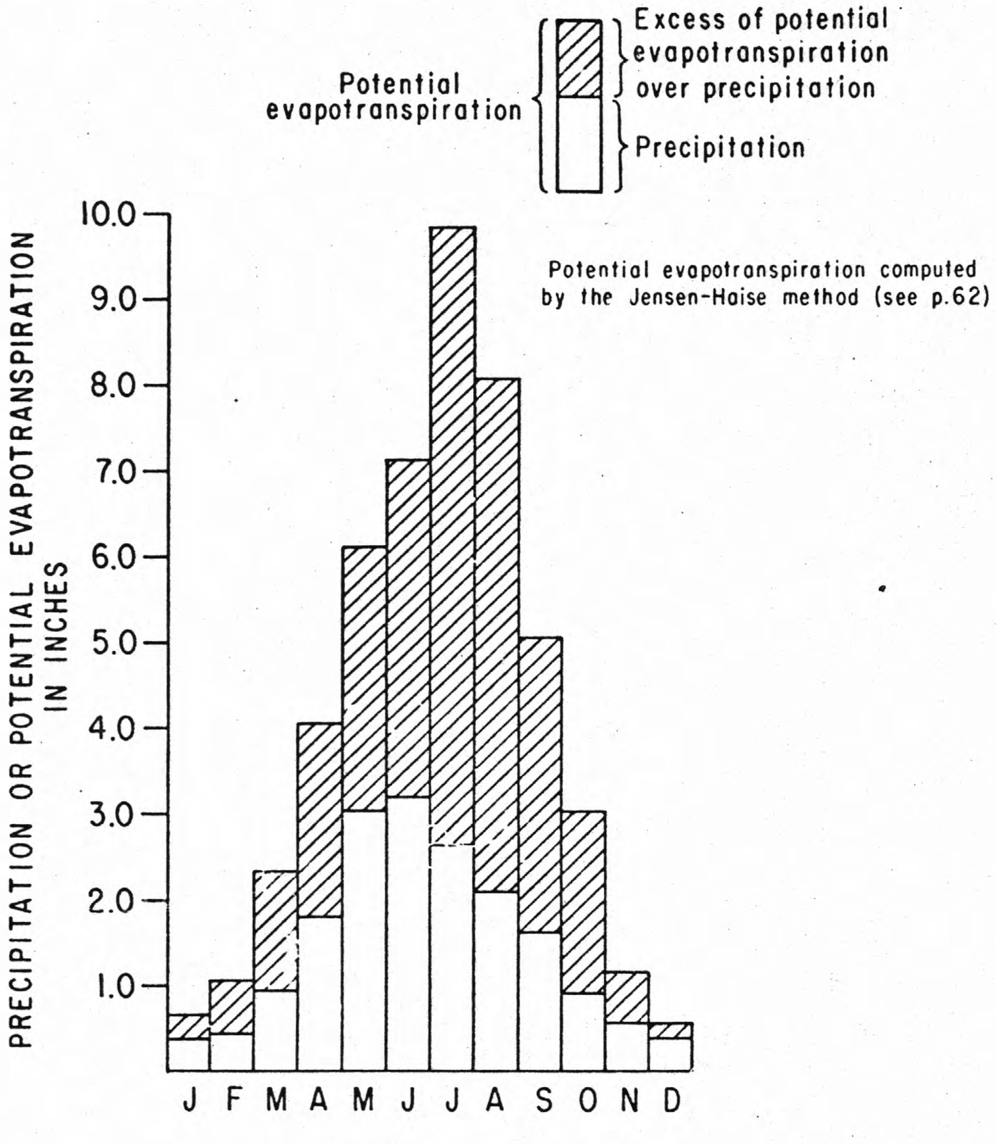


Figure 9.--Average monthly precipitation and potential evapotranspiration, 1925 through 1975, for the seven climatic stations shown in figure 6.

A paired t-test was also run on the monthly precipitation values. This test gives a measure of the significance of differences between stations when monthly readings at each station are paired with each other. Results of this test showed that while there may not be any significant difference between stations when the records as a whole are compared, variations on a month-by-month basis are extreme. Paired monthly precipitation data were found to be significantly different between all stations.

The last statistical test of spatial homogeneity was analysis of variance with Duncan's multiple range test. This test groups stations based on homogeneity of the total record. Only one homogeneous subset containing all seven stations resulted from this test.

All the statistical tests showed that monthly precipitation values from the seven stations may be pooled by averaging if long-term averages are desired. However, the paired t-test showed extreme variability of the records on a month-by-month basis. Consequently, for the remainder of the study, actual monthly values at the seven individual stations were used except where long-term average conditions were being analyzed. Thiessen polygons (fig. 6) were constructed around each station to areally distribute the point measurements of precipitation. All the area within each polygon is assumed to receive the same monthly precipitation as the representative station.

A complete monthly record from January 1925 to June 1975 was assembled for each station. Missing portions were synthesized using simple linear correlation with the station showing the best relationship. Correlation coefficients ranged from 0.72 to 0.85.

Although the actual monthly precipitation values were used for computations described in subsequent sections, two types of trend analyses were performed on the annual records to show dry and wet cycles. Five-year moving means and annual departures from the 1925-75 mean were computed for all stations and are presented on plate 6. This illustration clearly shows the dry periods of the 1930's and 1950's as well as the intervening wet periods.

Applied irrigation water.--The distribution of crops irrigated with both ground water and surface water is shown on plate 7. Most surface-water applications are in the valleys of the South Platte and Republican Rivers. Land in the South Platte valley is irrigated with water from the Western Canal and comprises approximately 12,600 acres (Nebraska Department of Water Resources, 1974). Total diversions in 1974 were 15,240 acre-ft. The water diverted into the Western Canal from the South Platte River is derived from outside the study area in the South Platte River basin (Hurr, Schneider, and Minges, 1975). Surface irrigation in the valleys of the Republican River and its tributaries uses water which was derived from ground-water discharge to the streams within and west of the study area and from surface runoff within and outside of the study area. In 1974, about 3,400 acres were irrigated in these areas.

Surface irrigation in the valleys of Spring, Stinking Water, and Frenchman Creeks (pl. 7) uses water derived principally from ground-water discharge to streams within the study area. Total area irrigated with surface water in these valleys was about 2,990 acres in 1974.

Use of ground water for irrigation occurs over most of the study area (pl. 7). Figure 4 shows that in 1974 an estimated 259,650 acres were irrigated with ground water with a total application of about 590,000 acre-ft. Methods used to compute annual applications are discussed in the section on withdrawals from the saturated zone.

Evapotranspiration.--The largest component of discharge from the soil zone is evapotranspiration (ET). No method exists at this time for measuring ET over large areas and for long periods of time. Consequently, ET is determined in such cases as the residual in a water-balance calculation or by empirically derived relationships among measured climatic variables. The latter approach was used herein because accurate water-balance methods require instrumentation and data analyses that are beyond the scope of this investigation.

The equation used in this study to compute potential ET was developed by Jensen and Haise and described by Jensen, Wright, and Pratt (1969). It is an adaption of the Penman equation with the wind velocity term neglected:

$$ET_p = R_s C(T-Tp) \qquad (2)$$

where ET_p = monthly potential evapotranspiration, inches,

R_s = total monthly solar radiation, expressed as inches of evaporation equivalent.

$$C = 1/[68 - 0.0036E + 650/(e_2 - e_1)] \text{ °F} \tag{3}$$

where E = altitude, in feet,

e_2 = saturation vapor pressure of water in mb at mean maximum air temperature for warmest month of year (July in this study),

e_1 = saturation vapor pressure of water in mb at mean minimum air temperature for warmest month of year (July),

T = mean monthly air temperature, °F.

$$T_p = 27.5 - 0.25(e_2 - e_1) - 0.001E, \text{ °F} \tag{4}$$

Solar radiation was computed for this study using the empirical relationship with percent possible sunshine developed by Rosenberg, (1964):

$$R_s = 0.000673d\, R_o(0.35 + 0.61S), \tag{5}$$

where d = number of days/month,

R_o = daily solar radiation on cloudless days, langleys/day,

S = mean monthly percent possible sunshine, expressed as a fraction,

0.000673 = a unit conversion from langleys to inches of water evaporation equivalent.

The nearest station to the study area measuring percent possible sunshine is the National Weather Service station at North Platte. Nebr. Monthly values of R_s were computed with equation 5 for the period January 1925 through June 1975 using monthly values of S (National Oceanic and Atmospheric Administration, annual) and the twelve monthly values of R_o for North Platte taken from Fritz (1949).

Using these values of solar radiation, air temperature, and elevation for the seven stations shown in figure 6, monthly potential evapotranspiration values were computed using equation 2 for the period January 1925 through June 1975. Missing temperatures were synthesized by linear correlation with other stations. Coefficients of determination were all greater than 0.99. The average annual ET_p values for all stations are shown in figure 10.

Evaporation from a large free water surface is considered to be a reliable index of potential ET. The reliability of potential evapotranspiration computed with equation 2 was evaluated by linear correlation with computed monthly lake evaporation at Enders Reservoir. Measured class-A pan evaporation rates for the period 1950 through 1975 (April-September) were corrected to lake evaporation by multiplying by a pan correction of 0.75. Using temperature and elevation data for Enders Reservoir, equation 2 was used to compute potential ET. The relationship between computed potential ET and lake evaporation is shown in figure 11. Although some scatter exists in the data and the slope of the regression line deviates slightly from unity, the coefficient determination of 0.73 shows that equation 2 provides a sufficiently accurate means of computing monthly potential evapotranspiration.

The set of computed monthly potential ET values for the 1925 through 1975 period was subjected to the same statistical tests for spatial homogeneity as the precipitation data (see p. 31). The conclusions from these tests were virtually the same as those for precipitation: long-term homogeneity exists among the stations, but month-by-month variations are significant enough to preclude areal grouping of the data on a short-term basis. Consequently, actual monthly values of computed potential ET were used in computations of water demand except where long-term conditions were being analyzed. Point values of potential ET at the seven weather stations were areally distributed using the same Thiessen polygons used for precipitation data (fig. 6).

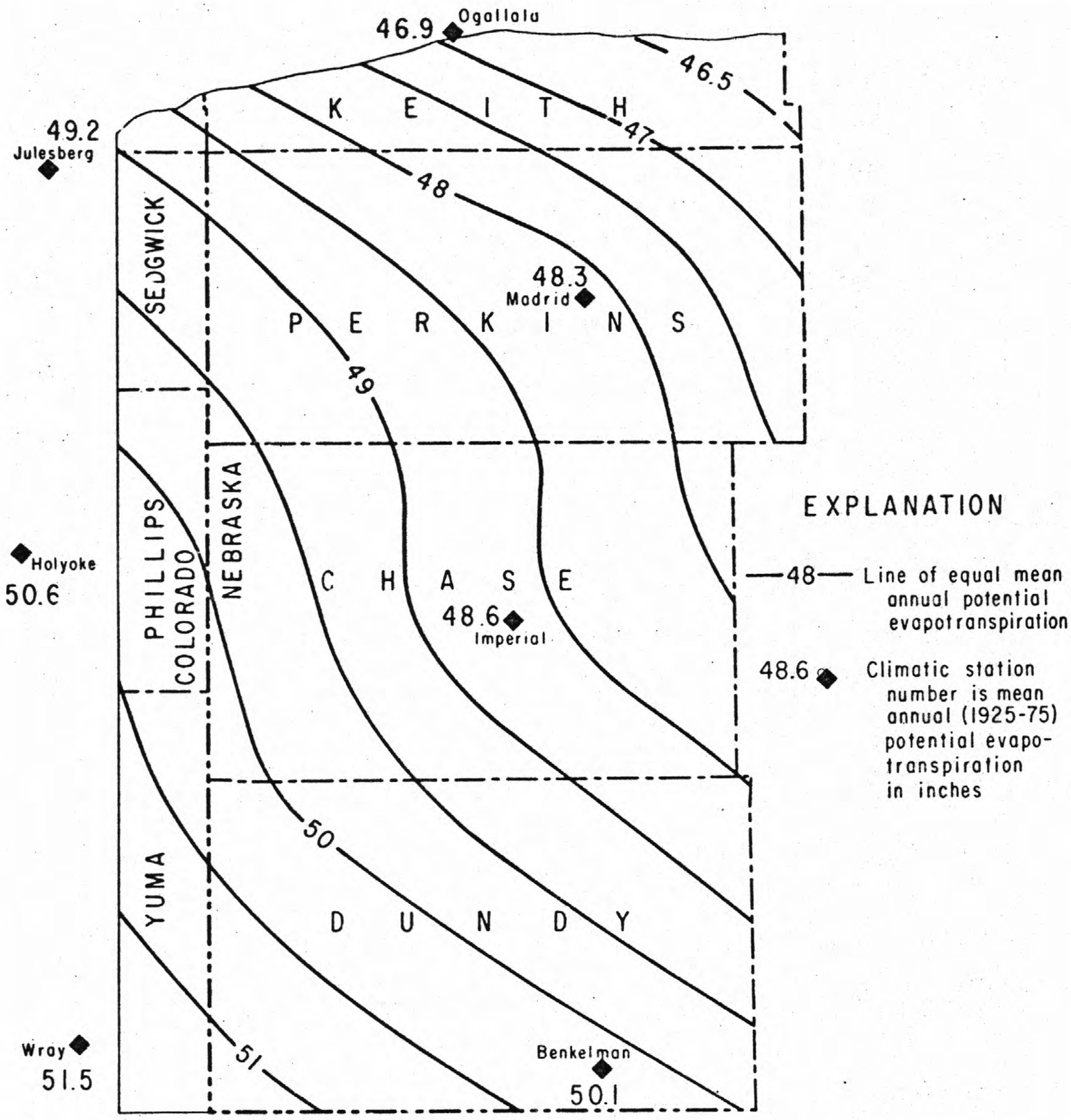


Figure 10.--Areal distribution of 1925 through 1975 of average annual potential evapotranspiration.

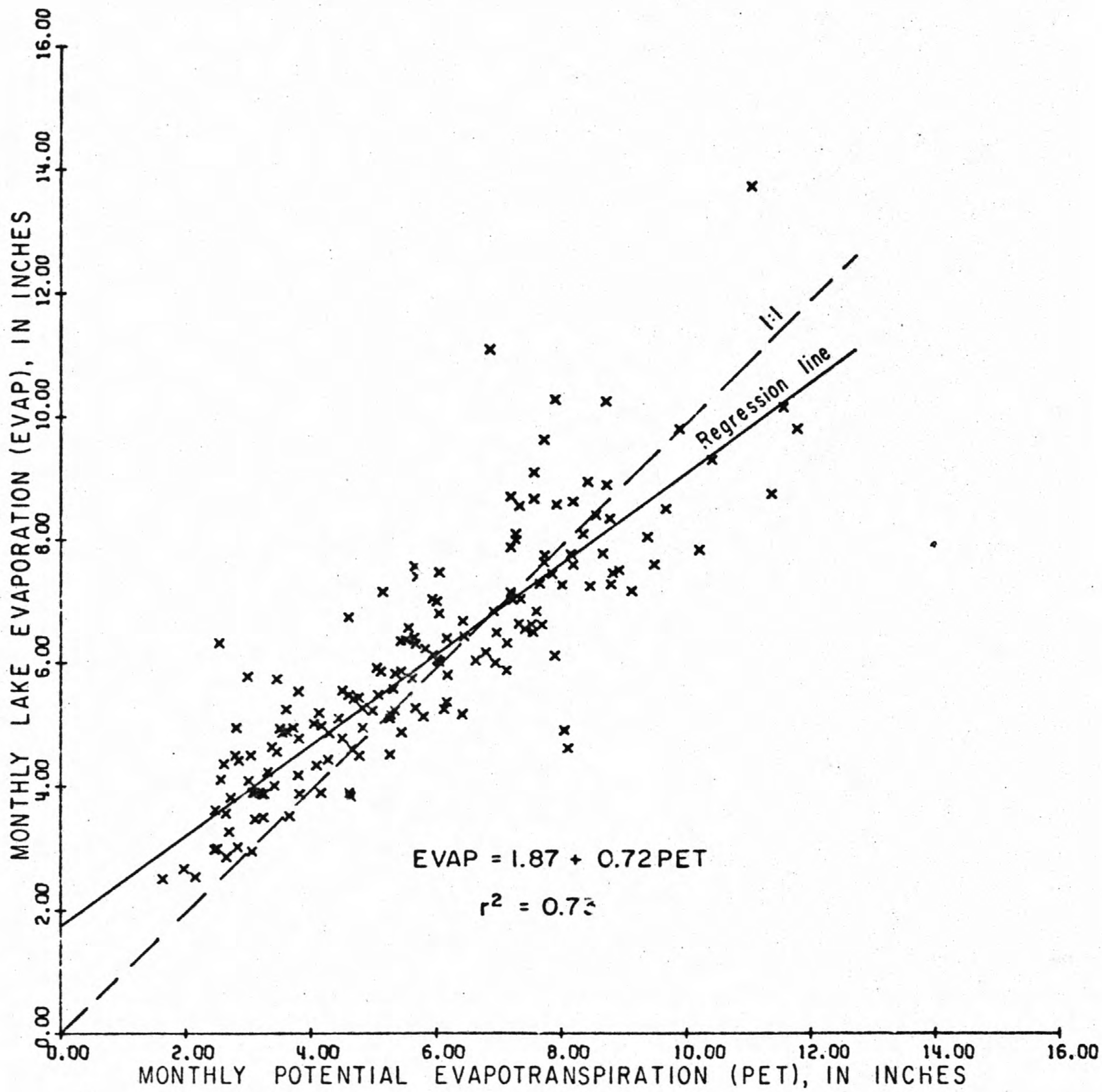


Figure 11.--Relationship between lake evapoiation computed from class-A pan evaporaticn and potential eva, transpitation computed with the Jensen-Haise method for Enders Reservoir, 1950 through 1975.

Actual evapotranspiration is a function of water availability and the moisture-stress adaptive mechanisms of the vegetative cover. The relationships at present are not well enough understood in terms of readily measured parameters to apply a rigorous approach to their determination in this study. Consequently, empirical methods were used for computing actual ET and are described in subsequent sections.

Deep percolation.--Data are not available for direct determination of amounts of deep percolation from the soil zone to the lower unsaturated zone over large areas. However, if the assumption of steady-state conditions (no addition to or depletion of storage in the lower unsaturated zone) is made, then long-term deep percolation must be equal to long-term recharge to the saturated zone. Further details of both temporal and spatial distribution of deep percolation under different soils, topography, crops, and climatic conditions are described in the subsequent section on recharge to the saturated zone.

Unsaturated Zone

The lower soil zone or unsaturated zone is hydrologically important because it transmits water from the soil zone to the saturated zone and provides potential storage space for excess water from natural or artificial sources. Both the storage capacity and the rate of water transmission to the base of this zone are determined by its thickness and lithologic character.

Hydrogeologic character

As used herein, the lower soil zone extends from the base of the soil zone to the top of the saturated zone, or to the capillary fringe immediately above the water table. The areal distribution of thickness of the unsaturated zone approximated by the depth to water measured in wells is shown in figure 12. The thickness of the soil zone and the saturated part of the capillary fringe should be subtracted from values shown on this map to arrive at the actual thickness of the zone as defined. However, over most of the study area their combined thickness is less than 5 percent of the values shown in figure 12. In this report the thickness of the soil zone is always considered to be less than 5 ft and the capillary fringe is generally less than 5 ft thick for the rock materials found immediately above the water table (Lohman, 1972).

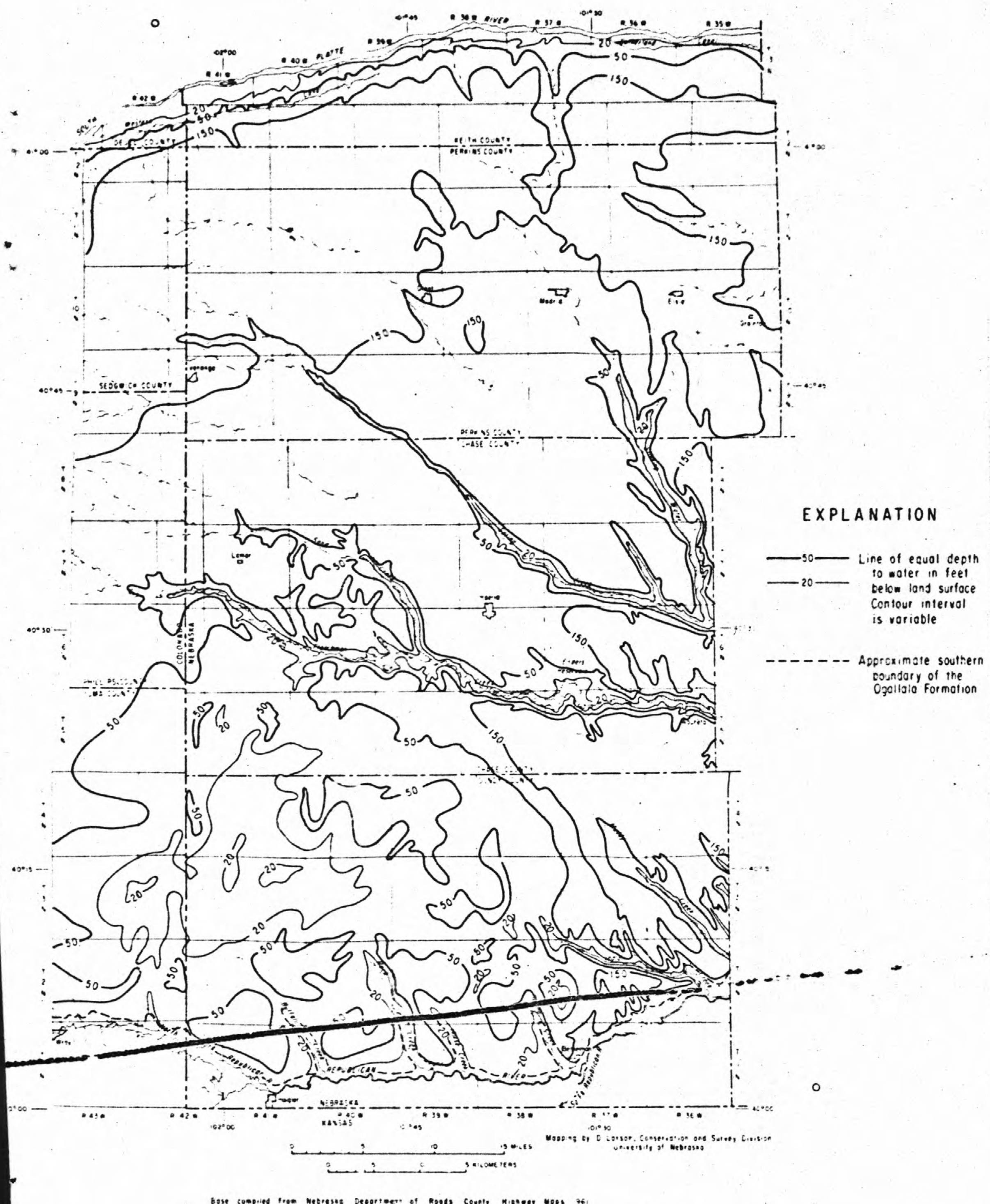


Figure 12.--Depth to water measured in wells in 1975.

The lower soil zone comprises materials ranging in size from clay through silt to sand and gravel and in age from Miocene through Holocene. The Miocene rocks in this zone consist of the upper part of the Ogallala Formation (Darton, 1905, p. 732-742) and may be unconsolidated or indurated clay, silt, sand and gravel, or caliche (locally referred to as magnesia rock). Detailed description of the origin and lithologic character of the Ogallala is given by Cardwell and Jenkins (1963, p. 40-49). The Ogallala Formation comprises most of the lower unsaturated zone throughout most of the study area. Much of this zone consists of fine-grained materials such as clay and silt, particularly in the western parts of Dundy and Perkins Counties. The hydraulic properties of the Ogallala Formation are discussed in the subsequent section on the saturated zone.

In about 14 percent of the study area the lower unsaturated zone consists of a combination of a relatively thick mantle of loess of the Quaternary Sanborn Formation as reported by Cardwell and Jenkins (1963, p. 49) and the Ogallala Formation. Soils developed on the Sanborn Formation are of the Keith Colby type, and the areal distribution of the Sanborn Formation is shown in figure 6 as areas underlain by these soils. Thickness of this loess deposit ranges from 0 to more than 190 ft. Texture is a uniform silt to silty clay, resulting in high storage capacity and specific retention and low hydraulic conductivity (table 2) These two properties reflect the ability of the loess to hold large amounts of deep percolation from the soil zone, thus delaying movement to the underlying Ogallala Formation and the Ogallala aquifer.

Much of the lower unsaturated zone in the study area consists of dune sand of Pleistocene to Holocene age overlying the Ogallala Formation and locally the Sanborn Formation. This dune sand consists of well rounded, well sorted quartzitic and feldspatic frosted grains (Cardwell and Jenkins, 1963, p. 51). The soils associated with the dune sand are principally the Valent and Valentine series (table 2). The areal distribution of the dune sand is shown by the occurrence of these soils (fig. 6). The dune sand is very important hydrologically as it has a low specific retention and relatively high permeability (table 2). These two properties result in rapid transmission of deep percolation from the soil zone to underlying materials.

Inflow and outflow

The lower unsaturated zone is considered to transmit all deep percolation from the soil zone to the saturated zone. Within the time frame of months to seasons used in this report, this assumption is not considered restrictive. Inflow to this zone is therefore considered equal to outflow. Deep percolation has been discussed previously, and recharge is discussed in the section on the saturated zone. Lateral movement into or out of the unsaturated zone is assumed to be negligible for this study.

Saturated Zone

The saturated zone comprises saturated water-bearing material of alluvial and aeolian deposits of Pleistocene to Holocene age, the Ogallala Formation of Miocene age, and the White River Group of Oligocene age. The Miocene (Ogallala), Pleistocene, and Holocene deposits containing the zone of saturation are combined and referred to as the Ogallala aquifer, or simply the aquifer. The sands of the underlying Chadron Formation of the White River Group of Oligocene age are considered a separate aquifer of limited areal extent and are called the Chadron sand aquifer.

The Ogallala aquifer

The Ogallala aquifer is characterized by large lateral and vertical variations in lithology within relatively small areas (Condra and Reed, 1959). This nonhomogeneous property is sufficiently complex to preclude separation of the aquifer into distinct water-yielding units when considering an area as large as the study area. The character of the aquifer is shown in three south-north hydrogeologic sections drawn through the study area (fig. 13). Correlation of sand and gravel units on these sections is generalized to show the occurrence of the principal water-yielding materials within the aquifer.

Lateral extent of the Ogallala aquifer.--The Ogallala aquifer underlies all but the extreme southern and extreme northwestern parts of the study area. The western boundary of the aquifer is in Colorado (fig. 14). The study was extended 5 miles into Colorado, and the western hydrologic boundary of the Ogallala aquifer discussed herein is taken as this line. In the extreme southern part of the study area the aquifer consists almost entirely of alluvium of Pleistocene and Holocene age along valleys of the Republican River and tributaries to its north side. The aquifer in the extreme northwestern corner of the study area consists almost entirely of alluvium of Pleistocene and Holocene age, where the Ogallala Formation has been removed (Condra and Reed, 1959).

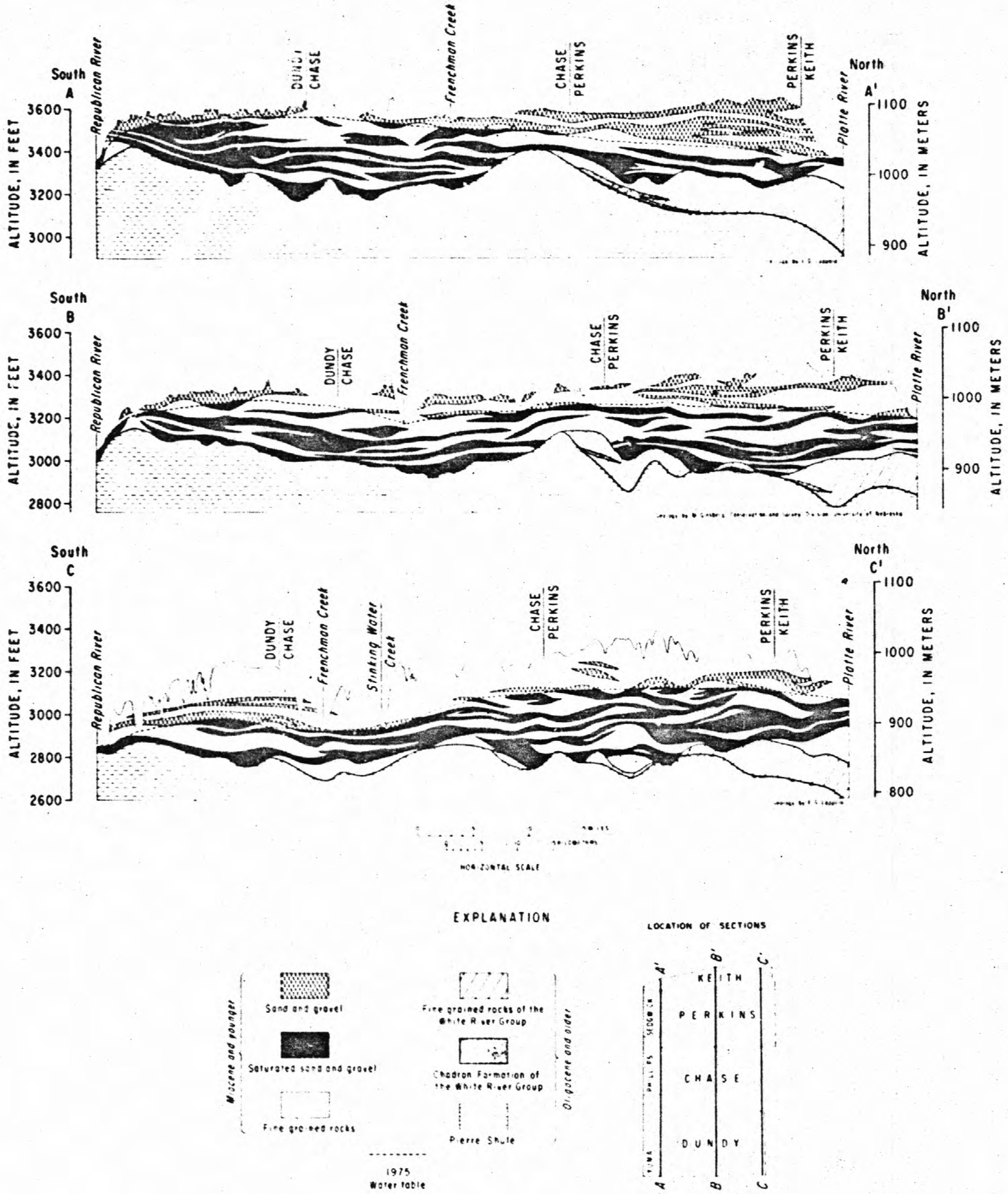


Figure 13.--Hydrogeologic sections A-A', B-B', and C-C'.

Both the Republican and South Platte Rivers act as hydrologic boundaries between the Ogallala aquifer and other aquifers lying south and north. The aquifer in the study area is continuous on both the west and east with the Ogallala aquifer lying outside of the study area (fig. 14).

Top of the Ogallala aquifer.--The top of the Ogallala aquifer is the top of the saturated zone or the top of the capillary fringe. This surface is difficult to measure for large areal studies. The top of the aquifer is therefore taken to be the surface defined at a given point in time by the level at which water stands in wells open to the aquifer. Most wells used for water-level measurements used in this study are irrigation wells that are screened through most of the saturated zone. Consequently, the surface defined by these measurements is a representation of the vertically averaged hydraulic head in the aquifer.

Water-level maps were prepared for two points in time for this study, one representing conditions prior to large-scale irrigation well development and one representing conditions in 1975. Preirrigation development conditions were considered to be represented by water levels measured over the period 1937 through 1952. Most of these were measured in 1949-53 as part of the study by Cardwell and Jenkins (1963). These measurements are supplemented by published water levels in Keith County measured in 1937 (Wenzel and Waite, 1941) and along the Republican and Frenchman Creek valleys measured prior to and during the 1950's (Bradley and Johnson, 1957; Condra, 1907; and Waite and others, 1946, 1948). The largest area having no measurements representing preirrigation development conditions was the part of Dundy County between the Frenchman Creek basin boundary and the Republican River valley. Preirrigation development water levels were estimated in this area using more recent measurements and extrapolating assumed changes backward in time. A map of the preirrigation development water-level configuration is discussed further in the section on model testing.

The configuration of the top of the aquifer was determined by using water-level measurements from about 450 wells measured in March 1975. The 1975 water levels provided the most consistently accurate widespread definition of the top of the aquifer used in the study (fig. 15). Water levels in wells in the Colorado part of the study area were taken from published records (Bjorklund and Brown, 1957; Boettecher, 1966; Boettecher and others, 1969; Cardwell, 1953; Hofstra and others, 1971; Hofstra and Luckey, 1973; Hofstra and others, 1972; Hurr and Luckey, 1973; Hurr and Schneider, 1972; and McLaughlin, 1948.)

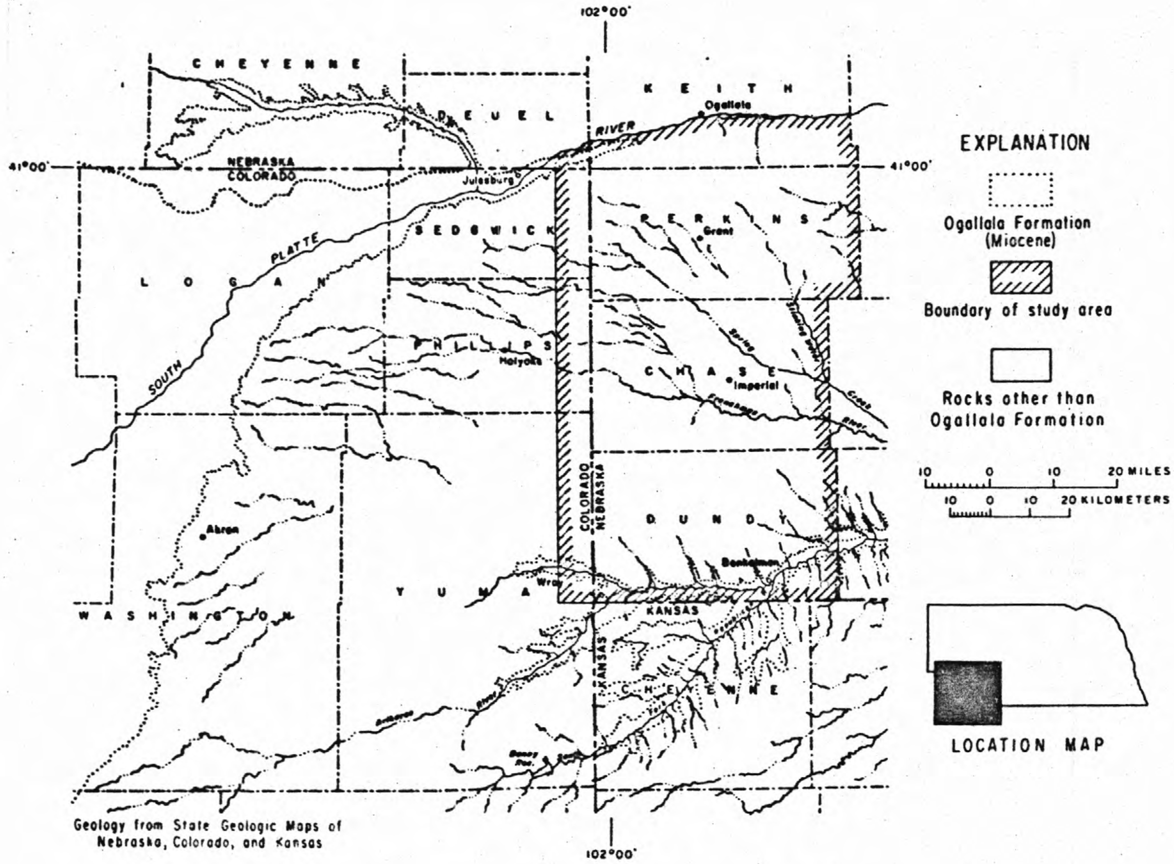


Figure 14.--Lateral extent of the Ogallala Formation in the region of the study area.

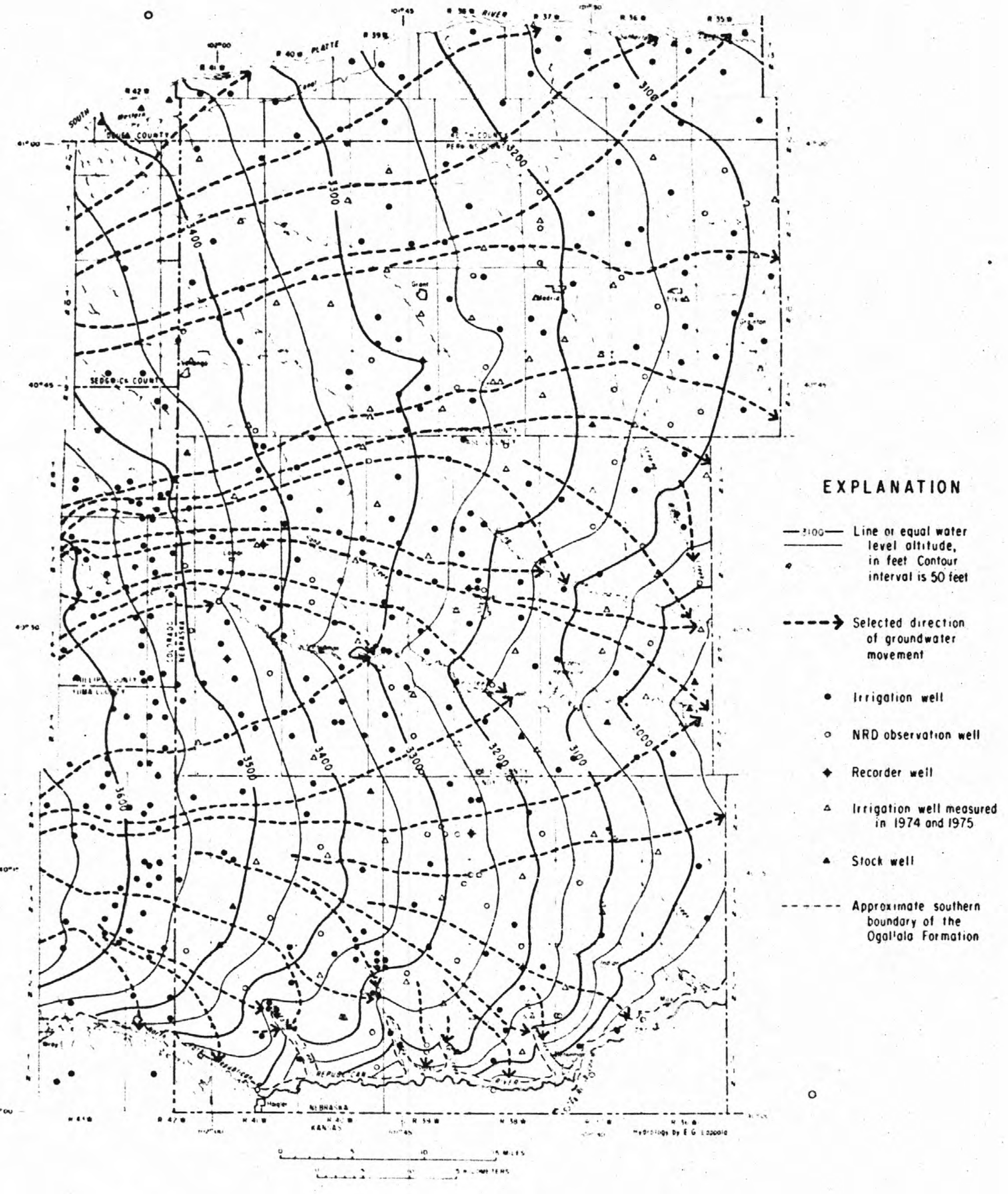


Figure 15.--Configuration of hydraulic head in the aquifer in the spring of 1975, showing general directions of ground-water movement.

The top of the aquifer slopes generally to the east at a rate ranging from more than 10 ft/mi in Dundy and Chase Counties to less than 5 ft/mi in eastern Perkins County (fig. 15). Gradients are steeper in the vicinity of hydraulically connected streams receiving ground-water discharge. These gradients are as high as 200 ft/mi along some of the tributaries to the north side of the Republican River. Areas of recharge can be identified on figure 15 as mounds or high areas in the water table. Areas of discharge are represented by low areas, principally along hydraulically connected streams such as Frenchman Creek. Amounts and areal distribution of recharge and discharge are discussed in a subsequent section.

Changes in water levels.--The top of the aquifer changes with time, depending upon the temporal variations in the balance between recharge and discharge. This balance is affected by both natural climatic variations and agricultural development. The response of water levels to local variations in climate is typified by the hydrograph for well 7N-38W-28CC prior to 1965 (pl. 8). The periods of water-level rise correspond to wet periods (pl. 6), and declines correspond to dry periods. The magnitude of these natural changes for this location is generally less than 2 ft, but may be as much as 5 ft.

Water-level changes due to agricultural practices have been superimposed on these natural fluctuations. Water-level declines are the direct result of increased discharge from the aquifer to supply the consumptive use of irrigated crops. These declines have been the most widespread over the southern two-thirds of the study area. The areas of largest declines correspond to areas of concentrated pumping. Although the declines shown in figure 16 are for the 1952-75 period, most of the declines have occurred since 1966. Water-level declines on the south side of Frenchman Creek in southeastern Chase and northwestern Dundy Counties may represent local declines in artesian head.

Natural seasonal patterns of water-level decline and recovery have been overshadowed by seasonal withdrawal of water for irrigation. Plate 8 shows the annual cycles as recorded by nine continuous recorder wells mantained in the District. The annual cycles of drawdown and recovery show the effect of an increasing draft on ground-water storage by irrigation wells. For example, the hydrograph of the Imperial recorder well (pl. 8) shows a pattern of increasing distance between the annual peaks and troughs as well as an overall decline in the recovery of water levels. This phenomenon in this area is caused by the increased density of irrigation wells in the vicinity of the recorder well.

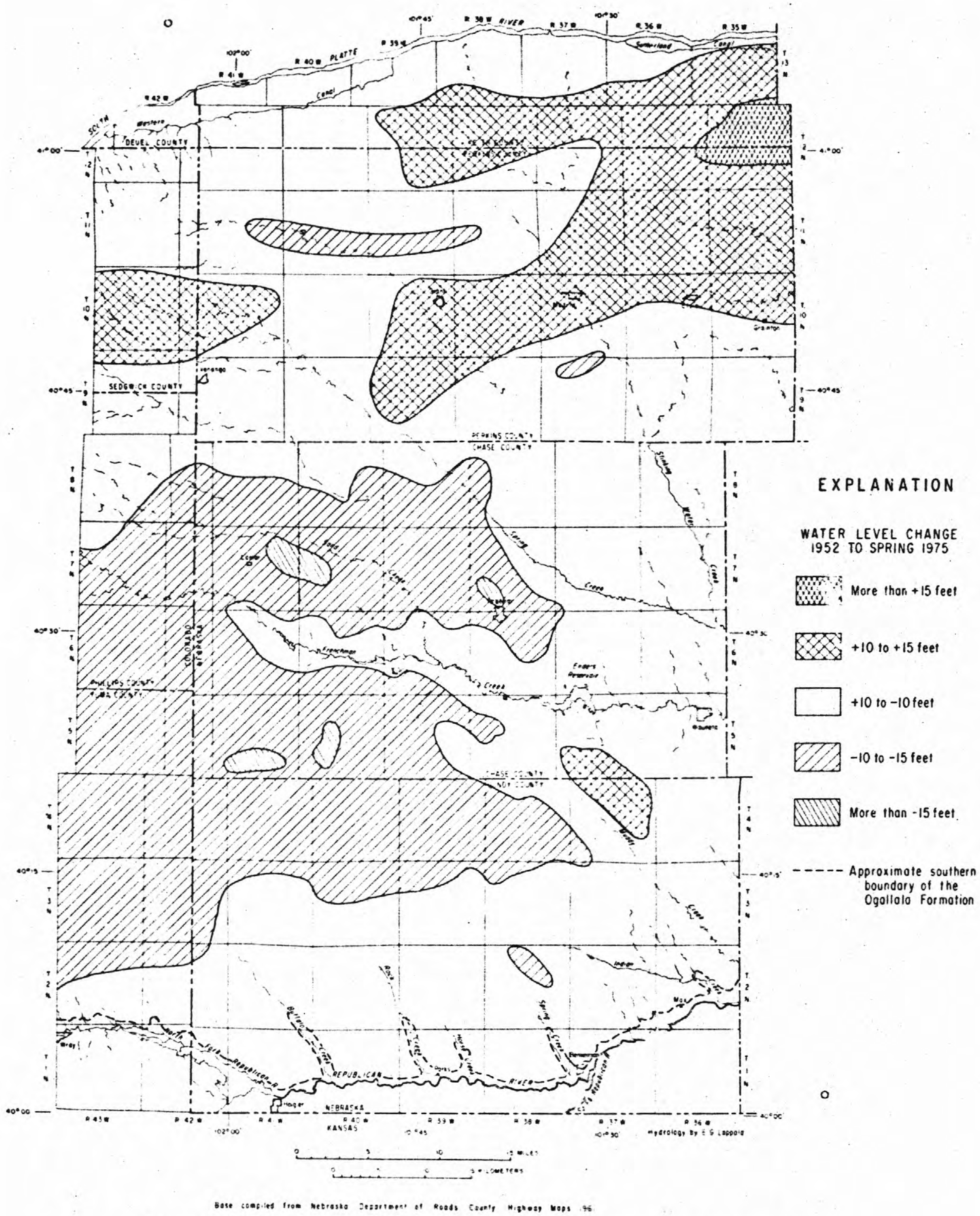


Figure 16.--Water-level changes observed between pre-irrigation development (1952) and spring 1975.

Water-level rises also have occurred in the District since at least 1937 as shown in figure 16 and plate 8. The rises are relatively widespread and, in general, underlie lands on which dryland wheat summer-fallow agriculture is practiced (pl. 5). Moisture available for recharge under these cropping conditions was computed for the 1925-52 period using methods discussed in the section on modeling of the soil zone (p. 94). In a given year over this period, about one-half of the land farmed under the wheat-fallow system was fallow (Nebraska Department of Agriculture) and was assumed to be subject to bare-soil evaporation only. Under these conditions, an average of 4.0 in/yr of water was available for recharge as deep percolation at the base of a 30-inch root zone. Similar computations for pastureland and rangeland for the same period resulted in only 0.20 in/yr of water available for recharge.

Records for five wells located in the area of water-level rise for which more than two measurements have been made over the 1935-75 period show rates of water-level rises during this period ranging from 4.2 to 14.5 in/yr and averaging 8.0 in/yr. Assuming an available moisture-storage deficit of 0.40 immediately above the capillary fringe, these rises are equivalent to a range of 1.7 to 5.8 inches and an average of 3.2 inches of water added to storage. This value is sufficiently close to the moisture available as deep percolation to explain the water-level rises. This analysis using water-level measurements assumes that rises are wide-spread enough to disregard the local short-term effects of mounding and lateral movement of water. The pattern of rise shown in figure 16 is considered large enough to make this assumption nonrestrictive.

The rising trend in water levels has been reversed (pl. 8) in areas where withdrawals for irrigation have increased. This reversal began in 1967 in some areas and has not yet occurred in others.

Base of the Ogallala aquifer.--The base of the Ogallala aquifer is defined as the lowermost sand or sand and gravel lying above:

(1) The weathered ochre clays and black shale of the Pierre Shale of Cretaceous age where the Oligocene White River Group is absent, or

(2) The olive-green and brown interbedded silts, siltstones, and clays of the White River Group.

As defined, the base of the Ogallala aquifer comprises a surface that is coincident with parts of early Tertiary through Pleistocene erosional surfaces cut into the bedrock. The character of this surface and the inferred drainage network are shown in figure 17. The base of the aquifer generally slopes to the east and northeast at about 20 ft/mi The bedrock ridge or high extending from south of Venango eastward roughly along the Chase-Perkins County line (fig. 17) is a hydrologically important feature.

This ridge is generally coincident with the southern extent of the underlying White River Group and its Chadron Formation. Sediments of the Ogallala aquifer overlying the western one-third of this ridge are generally fine grained, and high-yield wells are difficult to obtain. The bedrock ridge also corresponds reasonably well with a water-table divide (figs. 15 and 17). North of the divide, water moves to the north and east; south of the divide, water moves to the south and east.

Another hydrologically important bedrock ridge exists in the southern part of Dundy County (fig. 17). This ridge is parallel to the Republican River valley, and in places is sufficiently high to separate the Ogallala aquifer from the Republican River alluvium. The ridge has been breached by the southeast-flowing tributaries to the Republican River such as Rock Creek. This ridge is a predominant feature that is reported to parallel the north side of the Republican River as far east as Thayer County, Nebr. (Johnson, 1960).

The base of the Ogallala aquifer was mapped using the following data in order of reliability:

(1) Sample and electric logs of 100 test holes drilled by the Conservation and Survey Division of the University of Nebraska and the U.S. Geological Survey. Logs of wells within the Frenchman Creek Basin above Palisade, Nebr., were drilled during the 1950's (Cardwell and Jenkins, 1963). Logs in the Platte River basin were drilled as a part of separate studies (Wenzel and Waite, 1941). Test holes in Dundy and Perkins Counties were drilled during 1974 and 1975 under the Statewide cooperative test-drilling program between the Conservation and Survey Division and the U.S. Geological Survey.

(2) Drillers' logs of about 2,000 irrigation wells. These logs provided the detail shown in figure 17. Correlation of bedrock contacts between test-hole logs and drillers' logs was straightforward where the Pierre Shale underlies

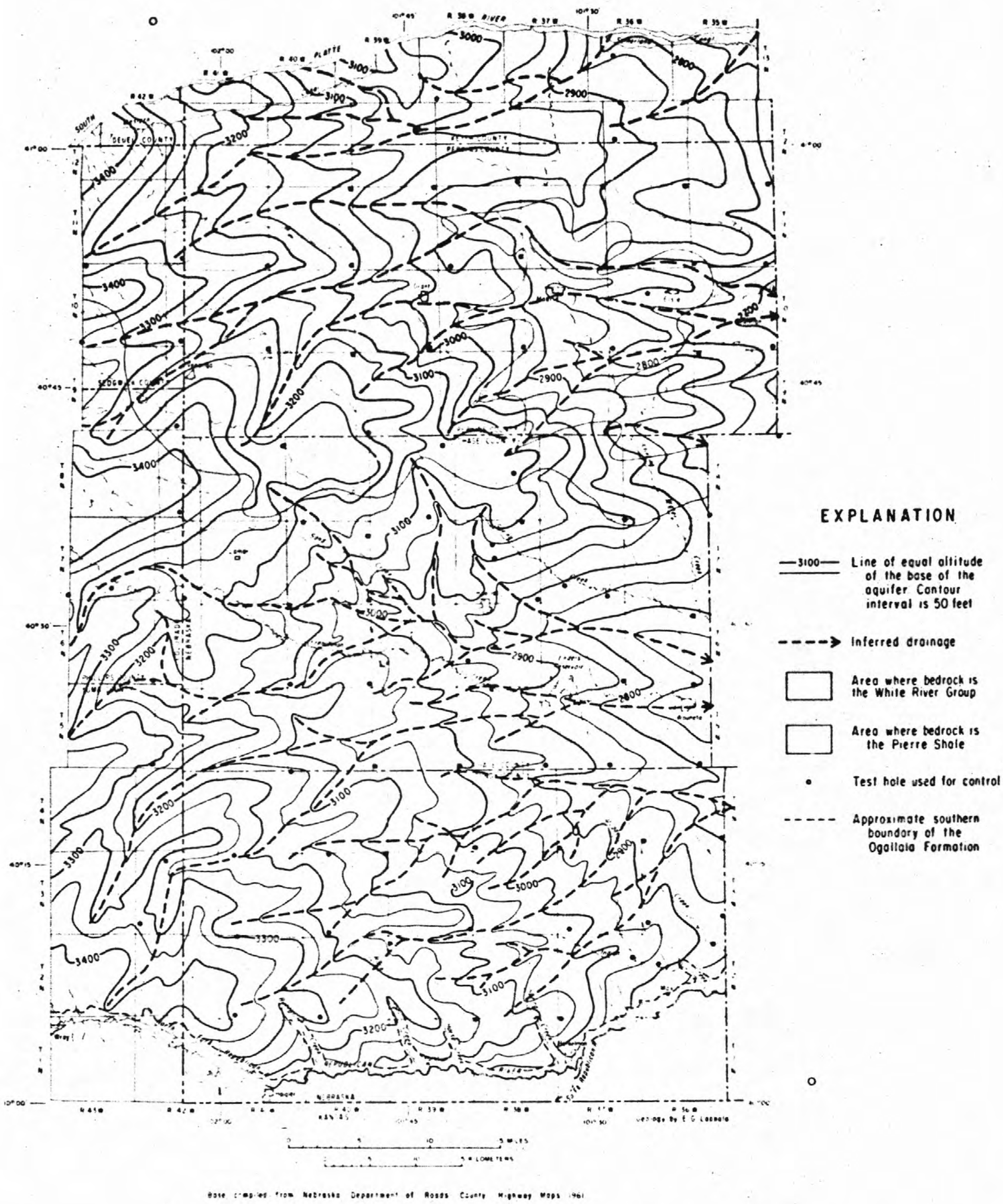


Figure 17.--Configuration of the base of the aquifer.

the aquifer. Correlating the contact between the Ogallala Formation and the White River Group was more difficult due to the often subtle nature of the differences in lithology.

(3) Geophysical and drillers' logs of oil and gas tests. Most of these are in Dundy and Keith Counties.

Water in storage.--Ground water is stored in the Ogallala aquifer n the pore spaces of rocks comprising the aquifer. The total volume of ater stored at a given point in the aquifer is determined by the hickness and total porosity of the aquifer. Recoverable water in torage is determined by aquifer thickness, the amount of water that can e derived by gravity drainage, the expansion of water, and the com-ression of the aquifer matrix.

The saturated thickness of the aquifer (fig. 18) ranges from less han 10 ft in the extreme northwest corner of the study area to over 450 t in western Dundy County. The porosity of the materials comprising he aquifer ranges from as low as 0.30 for sands and gravels to greater han 0.50 for some clays (Baver, 1956).

The lateral and vertical nonhomogeneity in the lithology of the gallala aquifer justifies spatial averaging of porosity. Variations in total porosity over the study area were not mapped. However, the general nature of the porosity of the aquifer is indicated by maps of the thickness of saturated sand and gravel (fig. 19) and the thickness of saturated clay and caliche (fig. 20). The remaining thickness of the aquifer comprises silt. These maps were prepared from sample description logs of the test holes.

Much of the Ogallala aquifer contains large thicknesses of silt and clay and, in many places, the sands and gravels are interbedded with silty zones (fig. 13) or are cemented with calcium carbonate. Using the total saturated thickness of the aquifer (fig. 18) and a range from 0.30 to 0.40 for total porosity, water stored in the Ogallala aquifer in 1975 was between 135 and 180 million acre-ft.

All of the water stored within the aquifer cannot be withdrawn by dewatering. The term drainable porosity ($\hat{\epsilon}_D$) is used to describe the fraction of stored water that is considered recoverable from ground-water storage by lowering the water table. Average values of drainable porosity for materials comprising the Ogallala aquifer are shown as follows (Johnson, 1967).

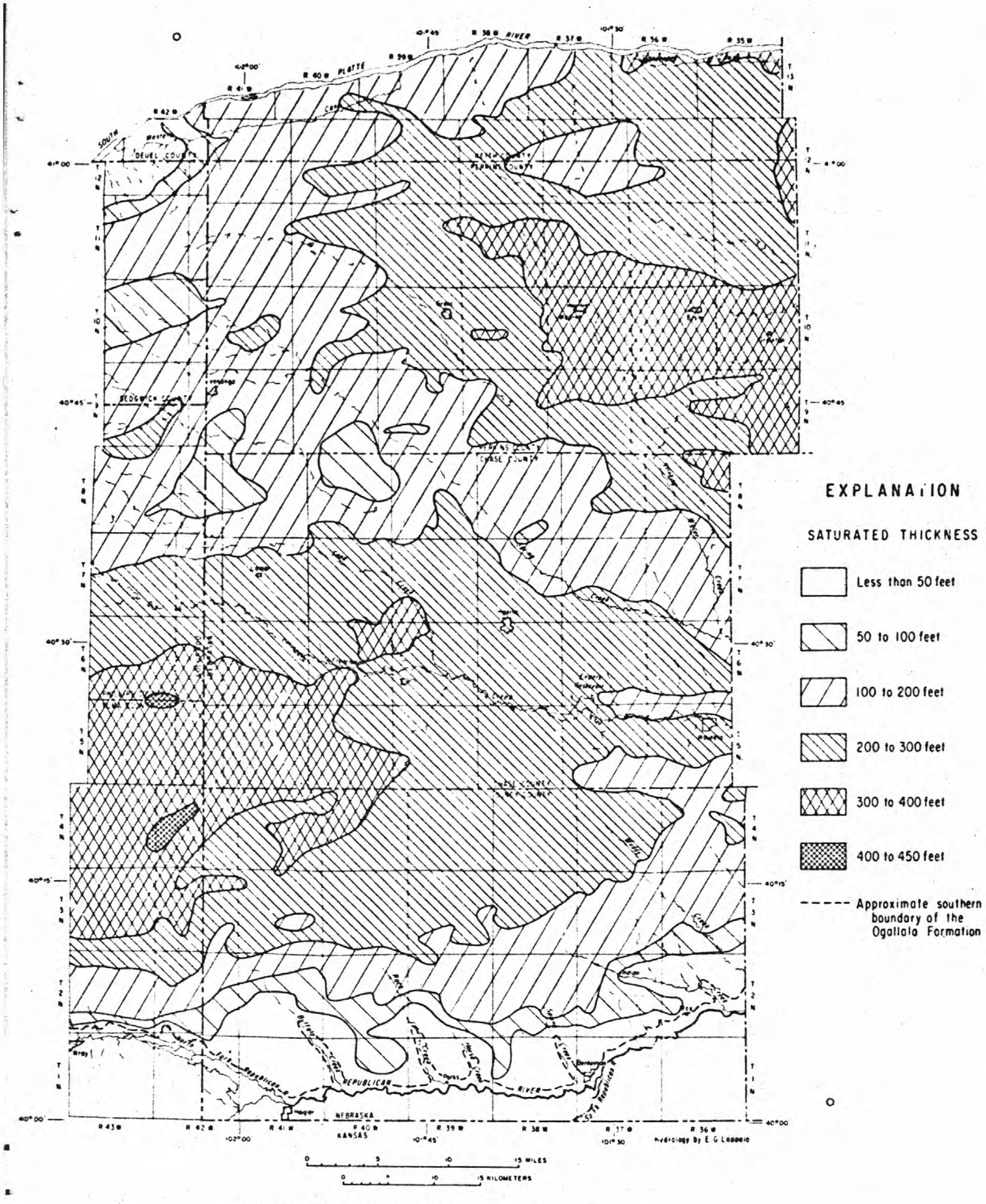


Figure 18.--Saturated thickness of the aquifer in 1975.

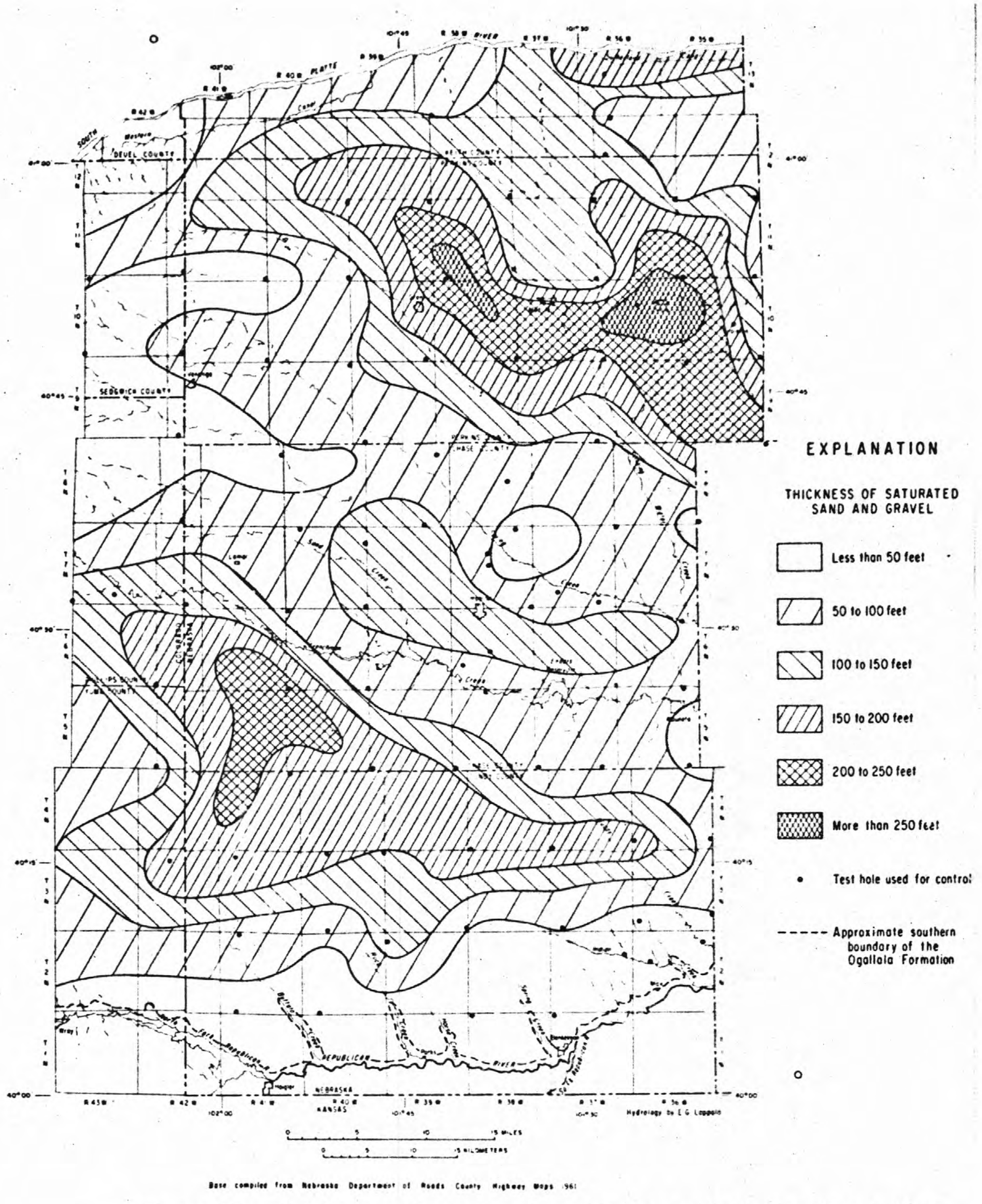


Figure 19.--Thickness of saturated sand and gravel within the aquifer.

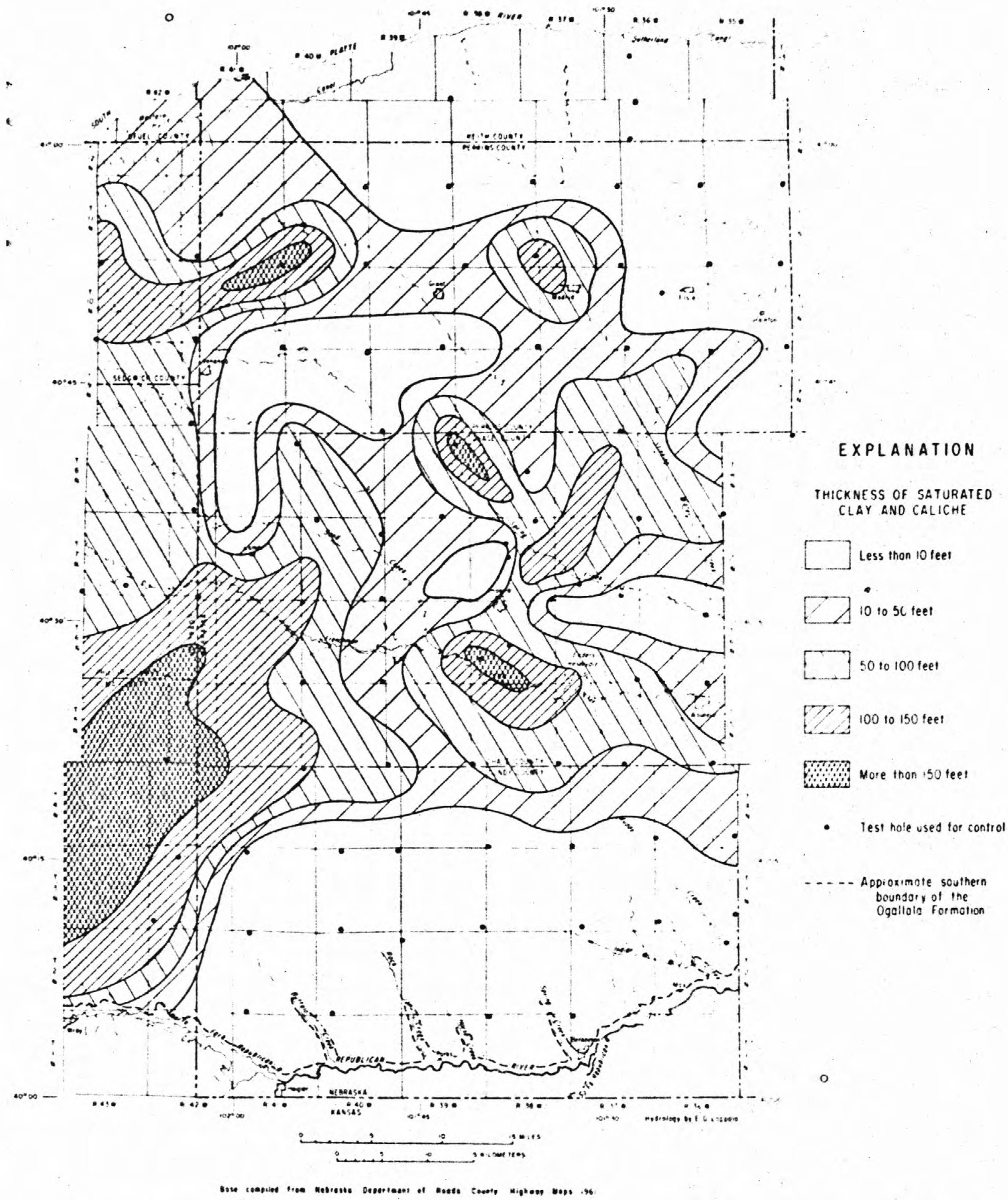


Figure 20.--Thickness of saturated fine-grained materials within the aquifer.

Drainable porosity for grain-size classes

Grain-size classes	Drainable porosity
Sand and gravel	0.25
Sandstone	.20
Silt	.10
Siltstone	.05

When unconfined parts of the Ogallala aquifer are dewatered by lowering water levels, the water from storage is drained from the top of the aquifer. Consequently, the drainable porosity used to describe changes in storage in the aquifer would be that of the dewatered materials. Rigorous evaluation of aquifer response at a given point would therefore use the value of drainable porosity for the materials being dewatered. This would require mapping vertical variations of this property over the study area. Although simulated aquifer response may be sensitive to these variations (DeVries and Kent, 1973), data availability precludes such a treatment for this study. Rather, vertically averaged values of drainable porosity were mapped using the values given in the previous table as applied to sample description logs from test holes (fig. 21).

The vertically averaged values of drainable porosity were computed as:

$$\hat{\varepsilon}_D = \frac{\sum_{i=1}^{n} \hat{\varepsilon}_i b_i}{\sum_{i=1}^{n} b_i} \qquad (6)$$

where $\hat{\varepsilon}_D$ = vertically averaged drainable porosity dimensionless,

$\hat{\varepsilon}_i$ = drainable porosity representative of the lithology in the sampled interval i,

i = an index on the number of sampled intervals,

n = number of sampled intervals,

b_i = thickness of the sampled interval, L.

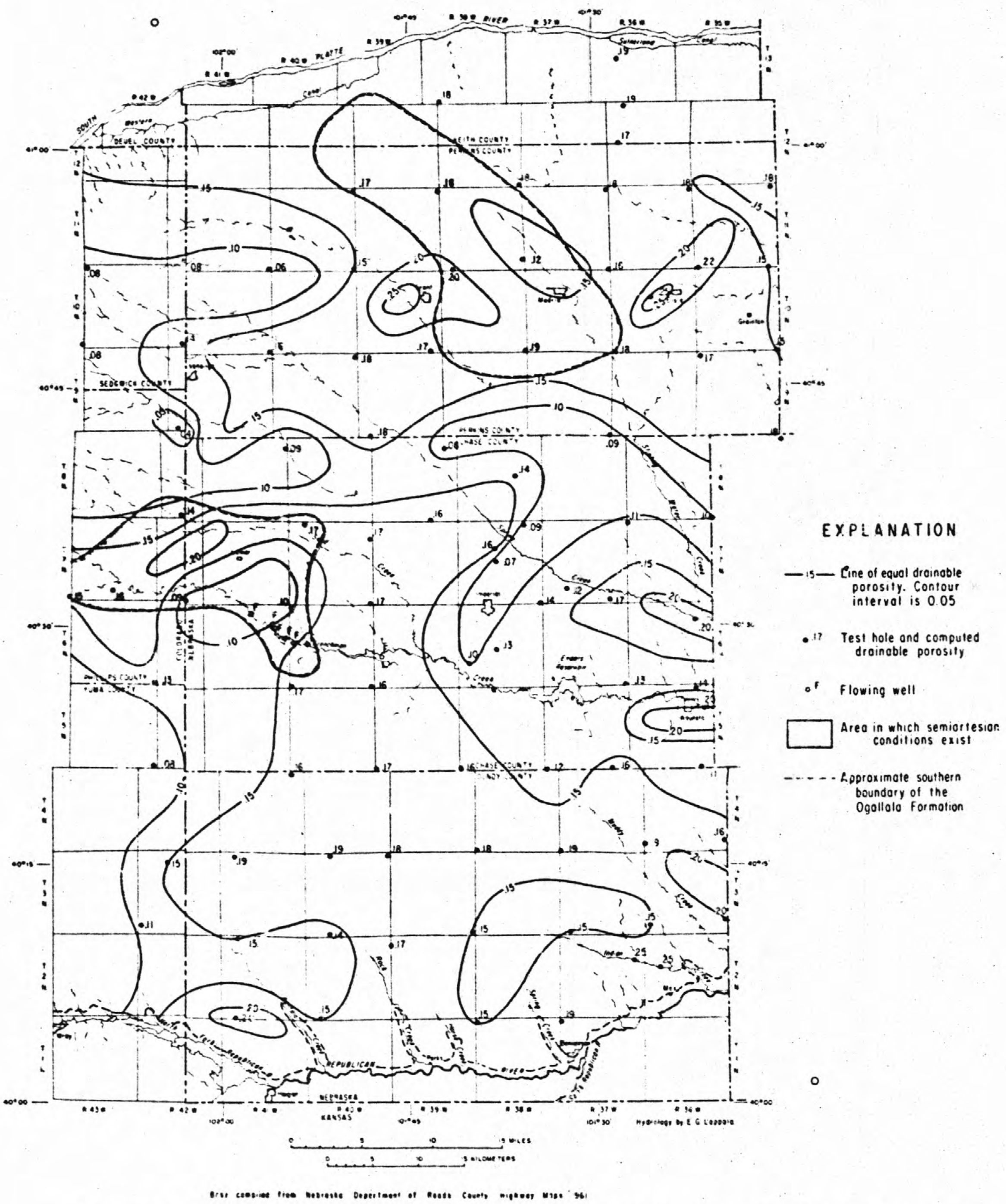


Figure 21.--Vertically averaged values of drainable porosity of the aquifer.

The vertically averaged values of drainable porosity determined with equation 6 are recognized as being possibly nonrepresentative of the correct value at a given point in time and space. However, the approach is considered justified for the following reasons:

(1) Data availability precludes a more detailed treatment.

(2) An areal distribution of $\hat{\varepsilon}_D$ is required that is valid for materials dewatered prior to 1975 and those which may be dewatered or have water added to them in the future.

Using the vertically averaged values of drainable porosity (fig. 21) and the saturated thickness of the aquifer (fig. 18), an estimated 68,355,000 acre-ft of recoverable ground water is stored within the Ogallala aquifer underlying the study area. The drainable porosity distribution is discussed further in a subsequent section on model testing.

Ground water in storage has been reduced about 420,000 acre-ft from 1952 to 1975. This is the net result of a reduction of 1.1 million acre-ft principally in the southern two-thirds of the study area and an increase of about 680,000 acre-ft in the northern one-third of the area. These storage change figures were computed using the northern one-third of the area. These storage change figures were computed using a mean value of 0.20 for drainable porosity (fig. 21) over the entire study area.

Local artesian or semiartesian conditions exist in parts of the study area. (See fig. 21.) Evaluation of historical and future water-level changes in these areas requires consideration of storage properties of the aquifer that are representative of these conditions in two areas.

The boundaries of these areas are only approximate and were delineated by examination of water levels and logs of test holes, locations of flowing wells, and measured and reported historic water-level fluctuations. The first of these covers about 130 mi^2 in western Chase County around Lamar (fig. 21). The materials comprising the semiconfining layer(s) are principally blocky, plastic clays (Cardwell and Jenkins, 1963) and thick deposits of fine sandy silt and caliche occur in the top one-third of the saturated zone (figs. 13 and 20). Evidence for semiconfined conditions in this area is the following:

1. Water levels in wells are above the base of the confining layer under nonpumping conditions. This semiartesian head averages about 20 ft. Over much of the area, the confined conditions become less important as pumping continues to locally draw water levels below the base of the confining layer. Typical reported drawdowns under pumping conditions in this area range from about 25 to 50 ft.

2. Flowing wells exist along the southern edge of the confined area. Measured yields of some of these wells are shown in table 4. All of these wells are in the valley of Frenchman Creek (fig. 21), suggesting the possibility of flowing water-table wells caused by increasing hydraulic head with depth in a ground-water discharge area (Lohman, 1972). However, available drillers' logs of these wells indicate that many of them are not drilled and screened to the base of the aquifer but are completed below the confining unit, thus minimizing the effect of increasing head with depth.

The unrestricted flow of well 6N-39W-19DBB has apparently not changed significantly since installation in 1956, as rates of flow are close to the original reported discharge. The delay in flow reduction of this well during the 1975 irrigation season until late summer indicates a response that is not typically artesian. Consequently, it is assumed that the hydraulic head causing this well to flow is derived at least partly from increasing head with depth in a ground-water discharge area. Unrestricted discharge rates of other flowing wells in this area are also shown in table 4.

Table 4.--Measured yields of flowing wells

Location	Date measured	Discharge (ft^3/s)
6N-39W-19DBB	05/20/75	2.10
	07/14/75	2.18
	10/07/75	1.74
6N-40W-17CDD	09/19/52	.67
	05/20/75	.96
6N-40W-20BAA	1/ 09/19/52	1/ .07
	07/14/75	1.00
	10/07/75	1.23
6N-41W-13ACC	09/20/52	.18
	05/20/75	.18
	07/14/75	.13
	10/07/75	.21

1/ May not be same well measured in 1975.

The flow of wells in the upper reaches of Frenchman Creek is thus thought to be caused by a combination of confined conditions and increased hydraulic head with depth in the ground-water discharge area along the valley. Unrestricted flows of these wells has reportedly declined since their installation. However, adequate documentation of the reduced flows does not exist to enable computation of reduced artesian head.

Local areas of artesian conditions exist throughout the study area. One small area supplies most of the flow of the artesian springs in the headwaters of Rock Creek in Dundy County (D. W. Thomssen, 1976). Semiconfined conditions also cover a large area, about 200 mi^2, principally in Perkins County (fig. 21). The producing zone is in the bottom one-third of the Ogallala aquifer in this area. The semiconfined conditions are evident from the response of observation wells to pumping. (See fig. 22.) Interference among wells during periods of heavy pumping in 1974-76 has caused pumping water levels to drop below the level of pumps in many wells. Pumps are reported to be set 30 to 35 ft below static

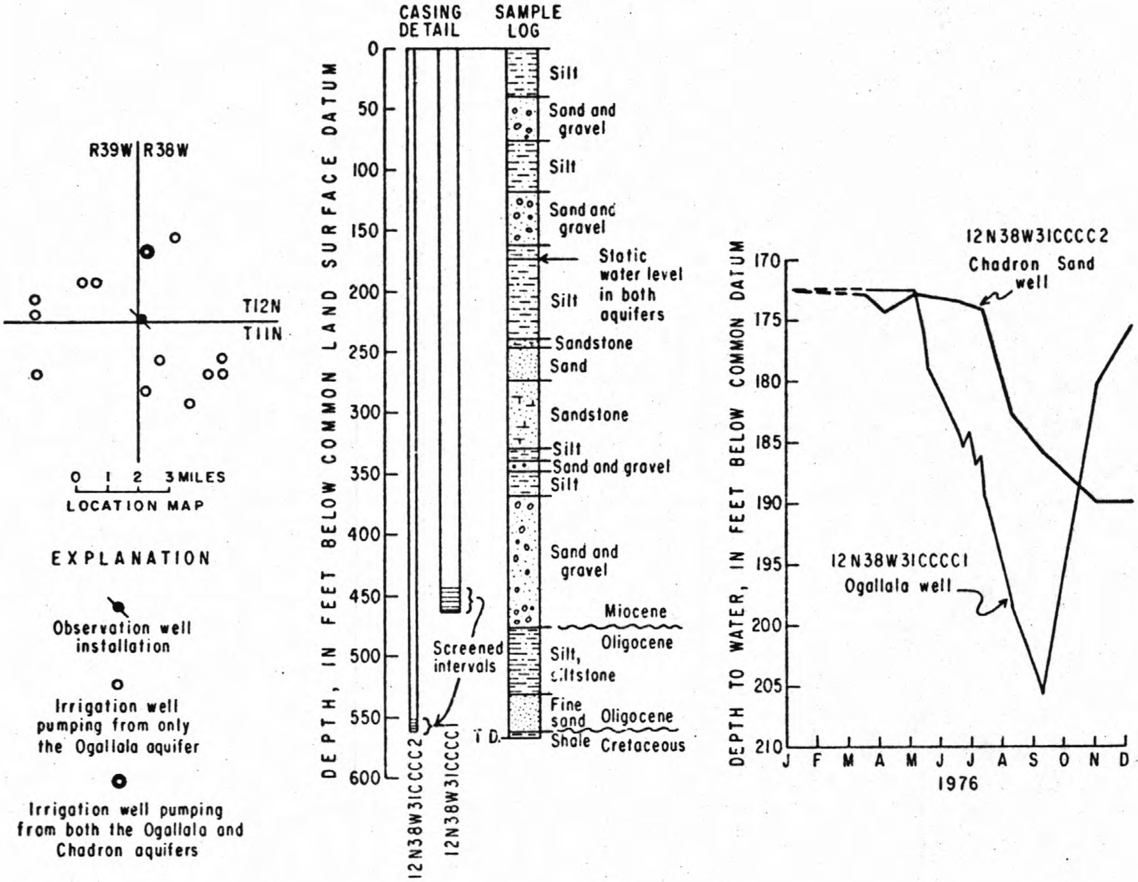


Figure 22.--Site details, stratigraphy, and 1976 hydrograph of observation wells 12N-38W-31CCCC1 and 12N-38W-31CCCC2.

water level in wells in the study area. This is generally sufficient under the unconfined conditions that occur over most of the study area. However, when the producing zone is in the lower part of the screened interval and the static water level in a well represents partly artesian head rather than the top of the producing zone, this pump setting often is not sufficient.

The artesian or semiartesian response of water levels to pumping in the central part of Perkins County is the result of interbedded clays, silts, and cemented sandstones in the upper two-thirds of the aquifer (fig 13). The semiconfined condition in this area is shown by the water-level response during the 1976 irrigation season of recorder well 12N-38W-31CCCC1 in Perkins County (fig. 22). This well is located 1.3 mi from the nearest irrigation well. To evaluate the approximate value of the artesian storage coefficient S, an analysis using the nonequilibrium well equations (Lohman, 1972) was applied to the area surrounding the recorder well for the May through September 1976 period. The nonequilibrium equation is:

$$W(u) = \frac{4\pi \hat{K} b s}{Q}, \qquad (7)$$

where $W(u)$ = the nonequilibrium well function

$$= \int_u^\infty \frac{e^{-u}}{u}\, du$$

s = drawdown in the observation well, L,

$\hat{K}$ = aquifer hydraulic conductivity, LT^{-1},

b = aquifer thickness, L,

Q = well discharge, L^3T^{-1},

$$u = \frac{r^2 S}{4\hat{K}bt}$$

where r = distance between the pumped well and the observation well, L,

S = the artesian storage coefficient, dimensionless,

t = time, T.

The drawdown in the observation well between June and August 1976 was about 30 ft. The aquifer thickness, considering only the sand and gravels (fig. 19), is about 195 ft. The value of hydraulic conductivity is estimated to be 50 ft/d based upon lithologic sample description methods described subsequently. The observation well is located about 14,000 ft from each of the three pumping centers. The assumption was made that the wells in these three groups are all pumped for the same 70-day period at a rate of about 900 gal/min. This is the average based on power records for 1976 for five of the wells shown on figure 22 (Wayne Sides, Midwest Electrical Cooperative, written commun., 1976). Applying the principle of superposition (Bear, 1972, p. 297), the drawdown (s) in the recorder well caused by each pumping center was assumed to be equal to the fraction of the total number of wells present in each group. Applying the above to equation 7 and solving for the artesian storage coefficient results in a value of about 0.002. The area over which this value is representative is difficult to estimate but was delineated in figure 21 by examining water-level fluctuations in wells and reported interference between wells. As water levels decline in this area, the semiconfined conditions may become less important if the fine-grained materials are permanently dewatered.

The second area in which semiartesian conditions exist is in the vicinity of Lamar, north of Frenchman Creek (fig. 21). Analysis of unpublished pumping-test data collected by Cardwell and Jenkins (1963) for well 7N-40W-5BBB using equation 7 resulted in a storage coefficient of about 0.04, which is low for unconfined conditions but very high for artesian conditions. An analysis of the same data accounting for the effects of vertical flow components during the early part of the test (Lohman, 1972) resulted in a storage coefficient of 0.01. Examination of the recorder charts for the Lamar recorder well (7N-41W-11DAA) indicates an apparent artesian response. However, there are 10 wells within 1 mi of this well, with two wells less than 1,500 ft away. By using an analysis similar to that used for well 12N-38W-31CCCC1 in Perkins County, the drawdown in the Lamar recorder well at the end of the irrigation season can be accounted for by the additive effect of pumping from these wells with a storage coefficient on the order of 0.01. It is concluded that this value is representative over the area around Lamar, However, as discussed subsequently in the section on model testing, a more reasonable value is 0.10 as shown in figure 21.

Ground-water movement.--Ground water moves through a porous medium in response to gradients in potential energy expressed as hydraulic head. Velocities are usually low and inertial forces may be ignored. The forces caused by the potential gradient are resisted by frictional forces. In analysis of areal water-supply problems, ground-water movement is defined as a statistically averaged unit volume flux of water as expressed by Darcy's law in vector notation:

$$\vec{q}_i = K \frac{\partial h}{\partial x_i} \qquad (8)$$

where $\vec{q}_i$ = the average unit rate of volume flux, LT^{-1},

K = the hydraulic conductivity, LT^{-1},

h = total hydraulic head, L,

x_i = a coordinate direction, L.

Interstitial rates of movement of individual water particles are approximated by dividing $\vec{q}_i$ by the total porosity. These velocities are important when considering transport of chemical species throughout an aquifer, the analysis of which is beyond the scope of this report.

Three separate methods of determining point values of hydraulic conductivity were used for this study. The results of each were analyzed and compared, but values derived using different methods were not intermixed in preparing the accepted hydraulic-conductivity distributions.

The first two methods to determine hydraulic conductivity used measured water-level responses in pumping and observation wells (Lohman, 1972). Cardwell and Jenkins (1963) made 18 aquifer tests, and the results are summarized in table 5 and shown in figure 23. The values of hydraulic conductivity determined using these tests are log-normally distributed (inset, fig. 23). This is in agreement with the findings of other investigators (Freeze, 1975). These tests were few in number and limited to only part of the study area, making areal extrapolation difficult. Cardwell and Jenkins (1963, pl. 7) attempted such an extrapolation of transmissivity, T (the product of aquifer thickness and hydraulic conductivity), over the Frenchman Creek basin part of the study area.

Table 5.--Hydraulic-conductivity values determined from aquifer tests (Qal=alluvium; To=Ogallala)

Well location	Aquifer	Duration of test (hours)	Hydraulic conductivity (ft/d)
5N-38W- 4DBA	To	0.6	174
5N-39W-35CBA	To	1.0	99
6N-36W-12CBB	Qal,To	5.8	130
6N-39W-30BBC	To	9.5	64
6N-41W- 1BBB	To	11.0	134
7N-38W- 5ADA	To	10.0	118
7N-40W- 5BBB	To	78.5	87
7N-40W-28BBB	To	436.0	70
7N-41W-21ABB	To	52.5	52
7N-43W-18ABB	To	13.0	160
7N-43W-27BBB	To	121.5	90
7N-43W-33ACD	To	359.5	106
7N-43W-35BCC	To	10.5	147
8N-40W-26BCD	To	13.0	41
8N-41W-34ABC	To	16.0	99
10N-41W-11DAB2	To	11.0	90
10N-43W- 3DCB	To	3.0	35
11N-39W-21BCD	To	10.0	24

(Source: Cardwell and Jenkins, 1963, p. 60.)

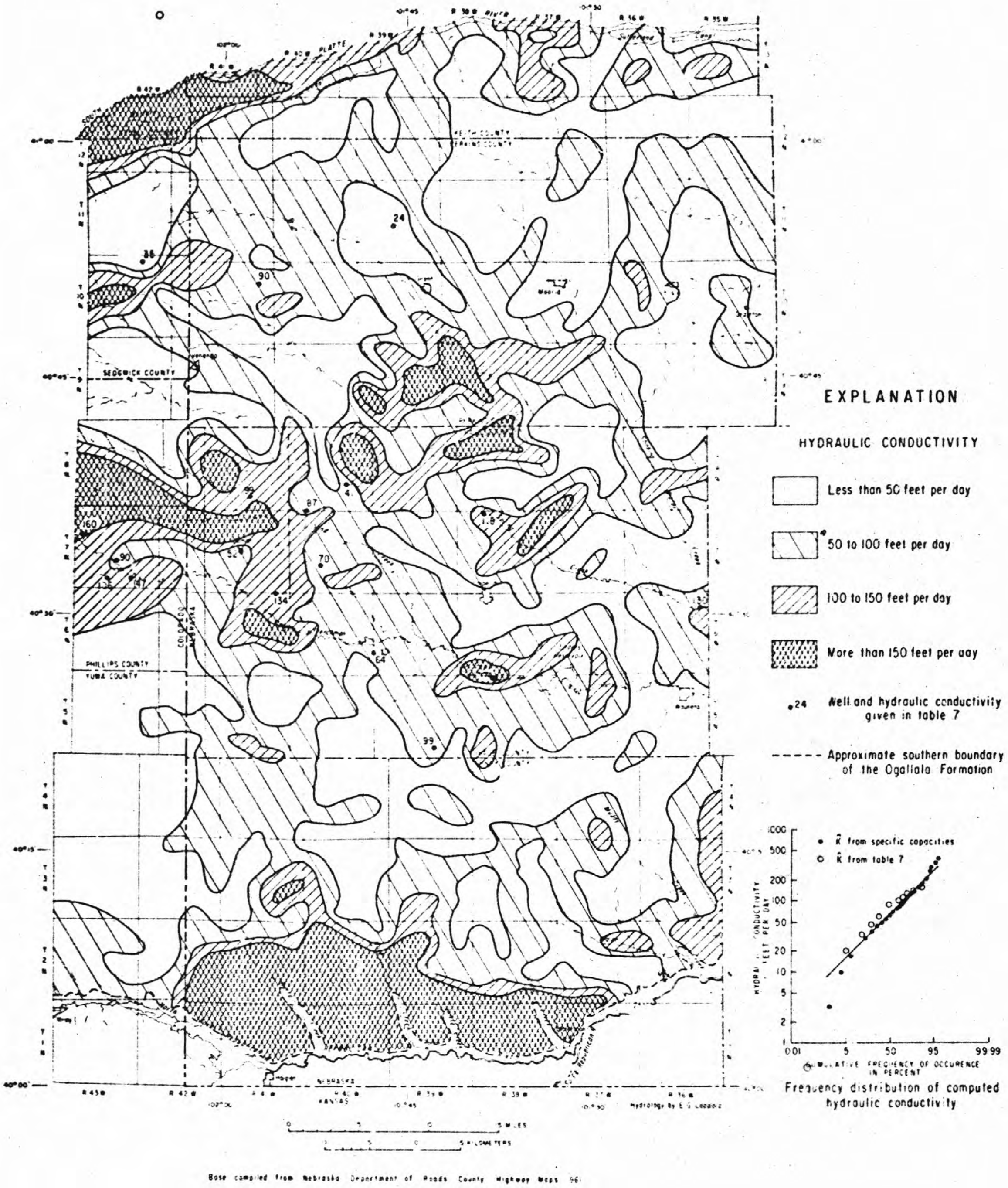


Figure 23.--Hydraulic-conductivity distribution determined from reported specific capacities of wells and from pumping tests made by Cardwell and Jenkins (1963).

The second method used to determine the hydraulic-conductivity distribution used drawdowns and yields of irrigation wells as reported on well registrations filed with the Nebraska Department of Water Resources. The quotient of the yield (Q) and drawdown (s) is termed the specific capacity (S_c) of a well:

$$S_c = Q/s, \ L^2T^{-1} \tag{9}$$

The relationship between specific capacity and transmissivity is dependent upon the storage coefficient, well diameter, and the duration of pumping (Hurr, 1966). Duration of pumping is not reported on the well registrations and is probably always less than a few hours. A simple relationship adapted from one derived by Cardwell and Jenkins (1963, p. 62) is:

$$K = \frac{0.22 \ S_c}{b} \tag{10}$$

where K = hydraulic conductivity, LT^{-1},

S_c = specific capacity, L^2T^{-1},

b = aquifer thickness, L.

Using this equation and values of aquifer thickness (fig. 18), values of hydraulic conductivity were determined for about 2,000 irrigation wells in the study area. The resultant hydraulic-conductivity distribution is shown in figure 23. These values were also log-normally distributed (inset, fig. 23) with a geometric mean of 61.5 ft/d. Hydraulic conductivities determined from specific capacities include all the errors in reported discharges and drawdowns. Pumping durations prior to drawdown measurement are not known but are reported to be on the order of a few hours. Until recently, pumping rates often were not reported accurately, tending to be high. These two factors combine to give a hydraulic conductivity computed using equation 10 that may be considerably too high. Preliminary analyses with models of the saturated zone (discussed in a subsequent section) indicated that the values of hydraulic conductivity using the second method augmented by values in table 5 were, in general, about twice those required to achieve reasonable values of ground-water discharge to streams under preirrigation development (1952) conditions.

Consequently, a third method was used to arrive at a hydraulic-conductivity distribution. This method determines a vertically integrated value of hydraulic conductivity using descriptions of samples from test holes and the following equation:

$$\hat{K} = \frac{\sum_{i=1}^{n} K_i b_i}{\sum_{i=1}^{n} b_i} \tag{11}$$

where $\hat{K}$ = vertically averaged hydraulic conductivity, LT^{-1},

K_i = hydraulic conductivity of the sampled interval, LT^{-1},

b_i = thickness of the sampled interval, L,

i = index on the sampled interval,

n = number of intervals logged.

The values of K_i in equation 11 were taken from table 6 developed by the Conservation and Survey Division of the University of Nebraska. The table was developed from laboratory permeability analyses of disturbed hydraulic rotary drill cuttings and well-performance tests (Reed and Piskin, University of Nebraska, no date). This method was applied to descriptions of samples from 100 test holes in and around the study area. The resulting areal distribution is shown in figure 24. The values are also log-normally distributed with a geometric mean of 30.8 ft/d (inset, fig. 24). The hydraulic-conductivity distribution resulting from interpretation of values from this third method was selected as the most representative, areally consistent distribution of this parameter. The basis for this selection was the agreement between computed and observed ground water discharge to streams as discussed further in the section on model testing.

Hydraulic conductivity can be a tensor quantity possessing directional properties (Bear, 1972). However, these directional properties or anisotropy in the aquifer being studied are not known. The aquifer is thus considered to be nonhomogeneous but isotropic.

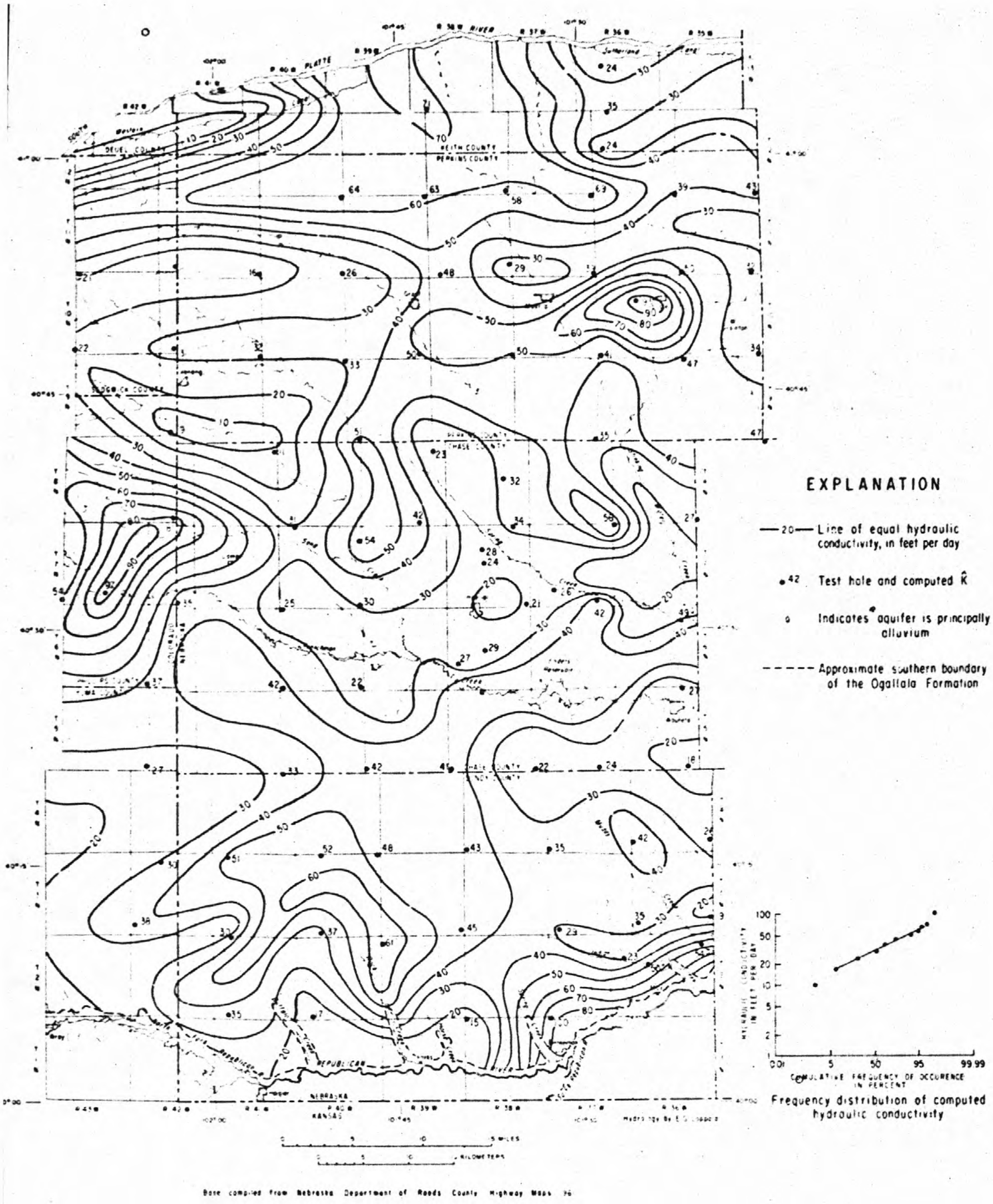


Figure 24.--Vertically averaged hydraulic conductivity determined from descriptions of samples from test holes.

Table 6.--Hydraulic conductivities estimated from grain-size descriptions
(In feet per day)

Grain-size class or range from sample description	Degree of sorting			Silt content		
	Poor	Moderate	Well	Slight	Moderate	High
Fine-grained materials						
Clay				0		
Silt, clayey				1 - 4		
Silt, slightly sandy				5		
Silt, moderately sandy				7 - 8		
Silt, very sandy				9 - 11		
Sandy silt				11		
Silty sand				13		
Sands and gravels[1/]						
Very fine sand	13	20	27	23	19	13
Very fine to fine sand	27	27	....	24	20	13
Very fine to medium sand	36 -	41 -	47	32	27	21
Very fine to coarse sand	48	....	....	40	31	24
Very fine to very coarse sand	59	....	....	51	40	29
Very fine sand to fine gravel	76	....	....	67	52	38
Very fine sand to medium gravel	99	....	....	80	66	49
Very fine sand to coarse gravel	128	....	....	107	86	64
Fine sand	27	40	53	33	27	20
Fine to medium sand	53	67	....	48	39	30
Fine to coarse sand	57 -	65 -	72	53	43	32
Fine to very coarse sand	70	....	....	60	47	35
Fine sand to fine gravel	88	....	....	74	59	44
Fine sand to medium gravel	114	....	....	94	75	57
Fine sand to coarse gravel	145	....	....	107	87	72
Medium sand	67	80	94	64	51	40
Medium to coarse sand	74	- 94	....	72	57	42
Medium to very coarse sand	84 -	98 -	111	71	61	49
Medium sand to fine gravel	103	....	....	84	68	52
Medium sand to medium gravel	131	....	....	114	82	66
Medium sand to coarse gravel	164	....	....	134	108	82
Coarse sand	80	10[illegible]	134	94	74	53
Coarse to very coarse sand	94	134	....	94	75	57
Coarse sand to fine gravel	116 -	136 -	156	107	88	68
Coarse sand to medium gravel	147	....	....	114	94	74
Coarse sand to coarse gravel	184	....	....	134	100	92

Table 6.--Hydraulic conductivities estimated from grain-size descriptions--Continued

Grain-size class or range from sample description	Degree of sorting			Silt content		
	Poor	Moderate	Well	Slight	Moderate	High
Sands and gravels[1]--Continued						
Very coarse sand	107	147	187	114	94	74
Very coarse sand to fine gravel	134	214	....	120	104	87
Very coarse sand to medium gravel	1,270 - 199-227		....	147	123	99
Very coarse sand to coarse gravel	207	....	....	160	132	104
Fine gravel	160	214	267	227	140	107
Fine to medium gravel	201	334	....	201	167	134
Fine to coarse gravel	245 - 289-334		....	234	189	144
Medium gravel	241	321	401	241	201	160
Medium to coarse gravel	294	468	....	294	243	191
Coarse gravel	334	468	602	334	284	234

[1] Reduce by 10 percent if grains are subangular.

Source: Reed and Piskin, no date.

Rates of ground-water movement computed with equation 8 for the study area range from less than 1 ft/yr to about 100 ft/yr under the prevailing hydraulic gradients (fig 15). The direction of ground-water movement in an isotropic, nonhomogeneous porous media is parallel to the hydraulic head gradient. Water generally moves from west to east through the Ogallala aquifer underlying the study area, as shown by selected arbitrary flow paths in figure 15. The regional ground-water flow pattern is altered in the vicinity of recharge and discharge areas. The flow paths diverge from areas of recharge and converge toward areas of discharge.

Inflow and outflow.--Inflow to the Ogallala aquifer in the study area is from the following: (1) ground-water underflow across the western project boundary, (2) recharge from precipitation and applied surface water that escapes the soil zone as deep percolation, and (3) possible local leakage from the underlying Chadron sand aquifer. Outflow from the Ogallala aquifer in the study area comprises: (1) ground-water underflow across the eastern project boundary; (2) discharge to the unsaturated and soil zones, including springs, streams, lakes, and phreatophyte transpiration; (3) discharge by wells; and (4) possible local leakage to the Chadron sand aquifer.

Temporary inflow to the Ogallala aquifer due to additions to bank storage from rising stream and reservoir stages and subsequent outflow due to declining stages is only significant in the reach of Frenchman Creek from Enders Reservoir downstream (U.S. Bureau of Reclamation, 1974)

Ground-water underflow was evaluated by subdividing the eastern and western project boundaries into equal increments and applying equation 8 over each increment for the entire aquifer thickness:

$$Q = \sum_{i=1}^{m} \hat{K}_i b_i w_i \left(\frac{\partial h}{\partial n}\right)_i \qquad (12)$$

where Q = underflow across the study area boundary L^3T^{-1},

i = an index on the interval used,

m = total number of increments,

b_i = average aquifer thickness over interval i, L,

w_i = width of the increment i, L,

$\hat{K}_i$ = average hydraulic conductivity over b_i and w_i, LT^{-1},

$\left(\frac{\partial h}{\partial n}\right)_i$ = hydraulic gradient normal to the boundary, dimensionless.

Using the hydraulic-conductivity distribution determined from test-hole data (fig. 23) and the water-table configurations in 1952 and 1975 (fig. 15), underflow was computed as follows:

	Acre-feet per year	
	1952	1975
Western project boundary	46,200	46,000
Eastern project boundary	38,300	38,000

These values differ from previously published values of about 80,000 acre-ft/yr (Cardwell and Jenkins, 1963, and Lappala, 1976) because the latter values were determined using the geometric mean hydraulic conductivity of 61.5 ft/d derived from specific-capacity measurements.

The small change in underflow from 1952 to 1975 is a result of the hydraulic-head gradient remaining fairly constant even with development along the Colorado-Nebraska State line. The aquifer cross-sectional area as indicated by water-level changes (fig. 16) has been reduced along the western project boundary by less than 0.5 percent.

Recharge to the Ogallala aquifer derived from precipitation and applied surface water is dependent upon factors discussed in the section on the soil and lower unsaturated zones (pp. 94-101). Quantification of recharge variations in time and space utilized modeling methods of both the soil and saturated zones as discussed in the section on model testing. These methods determined net recharge to the aquifer only. Applied water from both the aquifer and surface sources may be significantly different over small areas and periods of less than a few months.

The areal distribution of recharge under preirrigation development (1952) conditions was determined with steady-state modeling methods of the saturated zone. Recharge ranges from 0 to more than 6 in/yr and averages 2.6 in/yr. The techniques used are very sensitive to small variations in the water-table configuration which in turn are subject to measurement error and interpolation error between control points.

Consequently, the values at a given point may be in error. To eliminate some of the error in recharge values determined with this method, a smoothing process was applied to the computed values.

The pattern of recharge and discharge shown on figure 25 is consistent with known field conditions. Most ground-water discharge is concentrated along valleys of gaining streams, and most areas of high recharge are associated with sandy soils (fig. 6). The patterns of recharge and discharge are the result of local ground-water flow systems superimposed on the regional movement to the east.

Values of recharge in the study area have been determined by other investigators. Cardwell and Jenkins (1963, p. 97) computed a value of 1.03 inches over the Nebraska part of the Frenchman Creek basin above Palisade using a water-balance method. This value assumes recharge occurs equally over the entire study area. Redell (1967), using a steady-state analysis of the saturated zone similar to that discussed in a subsequent section of this report, determined recharge rates ranging from 0 to more than 6 inches per year for the Ogallala aquifer in Colorado including the western part of the study area. The highest rates occurred in areas underlain by dune sands. Longenbaugh (1975) in a recent study of recharge at Akron, Colo., 50 mi west of the study area, determined recharge rates to the Ogallala aquifer averaging more than 4 in/yr under three types of soil and farming practices similar to those in the study area.

An areal inflow-outflow method was also used to determine the recharge required to produce the average annual flow of Frenchman Creek near Imperial. The following equation was used:

$$R = \frac{Q_I - Q_u}{A} \tag{13}$$

where R = recharge equivalent over the entire ground-water contributing area, LT^{-1},

Q_I = annual baseflow before 1967 of Frenchman Creek above the gaging station near Imperial, L^3T^{-1},

Q_u = underflow across the north-south study area boundary line intersecting the ground-water basin divide, computed with equation 11, L^3T^{-1},

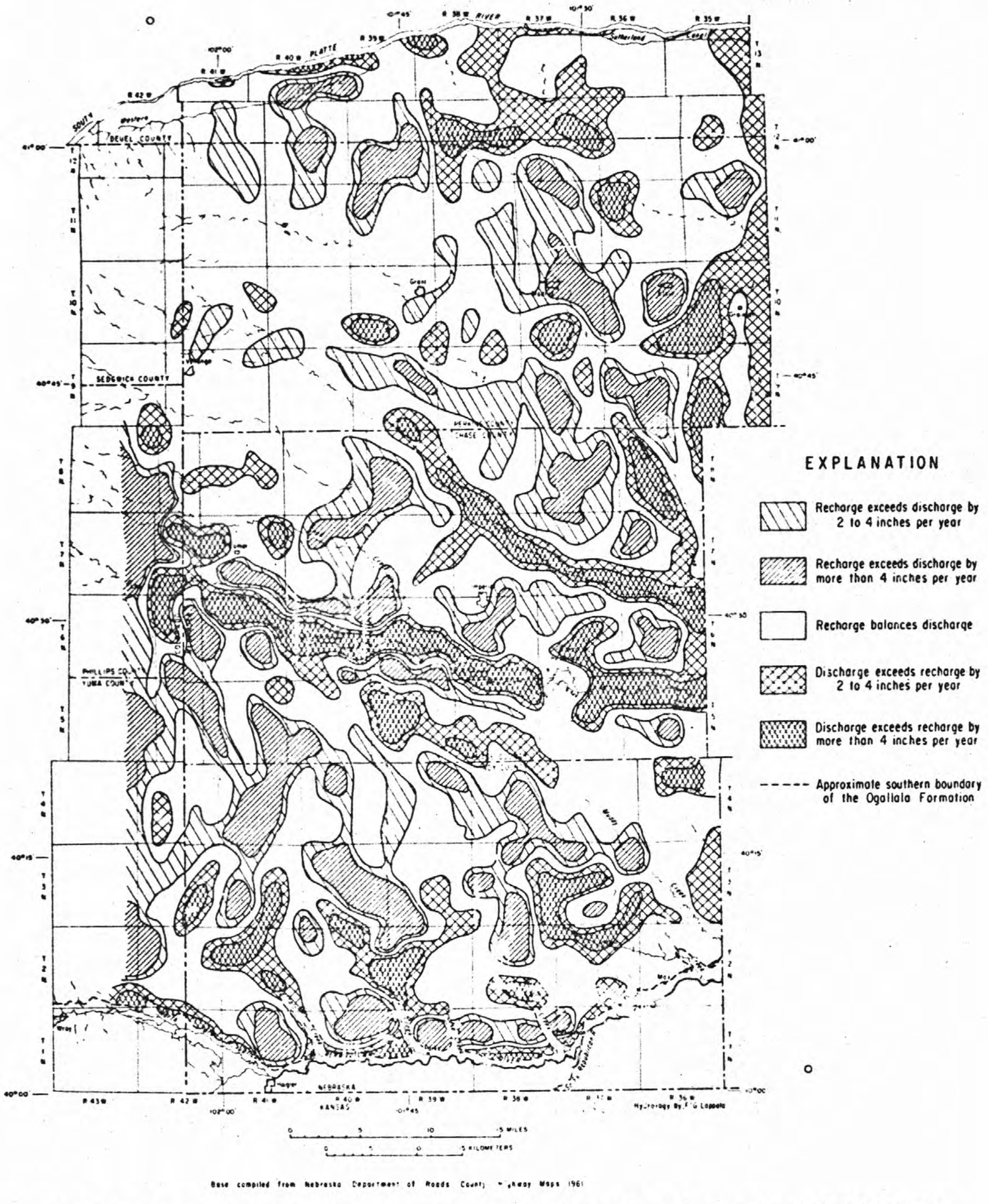


Figure 25.--Location and approximate magnitude of recharge to and discharge from the aquifer.

A = contributing area, equal to the ground-water basin east of the western edge of the study area (fig. 15) minus the areas of obvious ground-water discharge in the valley of Frenchman Creek, L^2.

Recharge computed with equation 13 is about 2.0 in/yr.

Recharge for the study area for preirrigation development conditions (prior to 1952) also was computed using modeling techniques of the soil zone discussed in a subsequent section. This method computed recharge as deep percolation based on average measured climatic conditions between 1925 and 1952 and gave an independent check on recharge computed with steady-state methods for the saturated zone. The principal drawback of the soil zone method was that no definition of where ground-water recharge or discharge occurs was possible. Potential recharge was computed for a given soil, crop, and topography, irrespective of conditions in the saturated zone such as a shallow water table that may control the recharge or discharge rate. Consequently, direct comparison of recharge computed with both methods in a given area was difficult, particularly as cropping patterns used in the soil model may not have been representative of steady-state conditions.

Recharge computed with the soil-zone model for different soil-topographic groups and crops is given in table 7. The values shown in table 7 were determined by averaging recharge rates computed from monthly climatic data for 1925-52 at each of the seven weather stations used in the study (fig. 6).

Table 7.--Average 1925-52 dryland recharge rates using soil depths determined by steady-state modeling methods

Soil-topographic complex	1925-52 average recharge, inches per year				
			Small grains		
	Row crops	Alfalfa	0 per-cent fallow	50 per-cent fallow	Pasture-land
Silt-loam, flat, upland	0	0	0.02	3.34	0
Silt-loam, rolling to hilly	0	0	0	2.52	0
Sandy loam, flat	0	0	.08	3.75	0
Silt-loam, flat, bottom land	0	0	.05	3.72	0
Sandy loam, rolling	.71	.12	1.08	4.70	.23
Sand and dune sand, hilly	3.15	1.32	3.72	6.31	1.41
Loamy sand, flat, interdune valleys	2.93	1.13	3.49	6.14	1.23
Rough broken land	0	0	0	2.56	0
Loam, rolling	.02	0	.12	3.45	0

The Ogallala aquifer is also recharged by seepage from streams, lakes, and reservoirs. Seepage through the beds of ephemeral streams provides much of the recharge in the western part of the study area. The recharge distribution shown in figure 25 reflects these seepage losses. The only stream that is perennial in the study area and is not directly connected to the aquifer is Muddy Creek in southeastern Dundy County. This stream apparently is perched above the water table throughout most of its length. Results of seepage measurements (pl. 2) indicate that there is little loss to the aquifer from this stream.

Seepage from numerous farm ponds and small runoff-detention structures also occurs in the study area, but the amounts are unknown. Seepage from Enders Reservoir and in the channel of Frenchman Creek during reservoir released have been previously discussed (fig. 5).

Discharge from the Ogallala aquifer to perennial streams in the study area has been previously discussed. Ground-water discharge to all perennial streams in the study area is a result of the valleys of these streams being incised into the aquifer and intercepting the predominantly eastward to northeastward direction of regional ground-water movement (fig. 15). This interception explains the asymmetrical water-table configuration on either side of these streams. Further, this asymmetry results in higher hydraulic head gradients and consequently larger ground-water discharge along the south and west sides of these streams than along the north and east sides.

One of the most obvious examples of the foregoing is Frenchman Creek above the gaging station near Imperial (fig. 15). Discharge to Frenchman Creek in this basin is predominantly from the south side. The ground-water basin providing this discharge is asymmetrical and does not coincide with the topographic basin. In fact, the water-table divide defining the north side of the ground-water basin actually crosses Frenchman Creek 8 valley miles above the point of effluence. Based on steady-state analysis of the saturated zone, this reach of Frenchman Creek receives about 65 percent of its base flow from the south side and 35 percent from the north side.

Discharge from the aquifer also occurs in areas where the water table is near enough to the land surface to permit evapotranspiration. Ground-water discharge by ET is highest in areas where phreatophytes such as willows (Salix) and sedges (Carex) grow. Most consumptive use of ground water occurs along stream valleys in the study area where depth to water is less than 10 ft. (See fig. 12.)

Timing of the seepage measurements made in 1975 (May, July, and November) was selected to determine the seasonal effects of ground-water pumping and evapotranspiration in the stream valleys. Withdrawal by evapotranspiration and wells should be at a minimum in May and November and near maximum in July. The measurements (pl. 2) generally support this contention. Stream depletion continues long after discharge by ET and wells has ceased. The November measurements were made at least 10 weeks after most wells stopped pumping and at least 6 weeks after the end of the growing season.

The magnitude of ET discharge of ground water during 1975 was computed using the May and July seepage measurements for Indian Creek which is not subject to significant depletion by wells. Other perennial streams (Horse, Buffalo, and Spring Creeks) were either measured at only one point or ground-water discharge by wells was considered too large to be compared to ET discharge (Frenchman, Spring, and Stinking Water Creeks). The apparent magnitude of ET discharge along Indian Creek was 0.25 $(ft^3/s)/mi$ during July of 1975. This value represents a maximum ET discharge; the average over the growing season is probably about one-half of the maximum or 0.12 $(ft^3/s)/mi$.

The annual ET discharge along perennial stream valleys was also computed using streamflow records at selected gaging stations. This method is based on the following assumptions:

1. Total ground-water discharge at a given point in time in these valleys is the sum of discharge to streams and ET.

2. Over long periods of time the total ground-water discharge remains constant under natural conditions.

3. Measured streamflow is essentially all base flow.

Using the above, average monthly flows were determined for the periods of record not affected by ground-water withdrawals for Frenchman Creek near Imperial, Stinking Water Creek near Wauneta, and Buffalo Creek near Haigler.

Assuming that ET ceases by November, average ET discharge above these streamflow stations was determined by subtracting average monthly flow during the growing season from the November flow (fig. 26). The annual ET rates determined by this method compare reasonably well with the value determined for Indian Creek. From the foregoing, it is estimated that under historic climatic conditions average ET discharge of ground water from valleys of perennial streams in the study area is 0.10 $(ft^3/s)/mi$. This represents a discharge of 14,500 acre-ft/yr for the approximate 200 mi of stream valley assumed to be ground-water discharge areas.

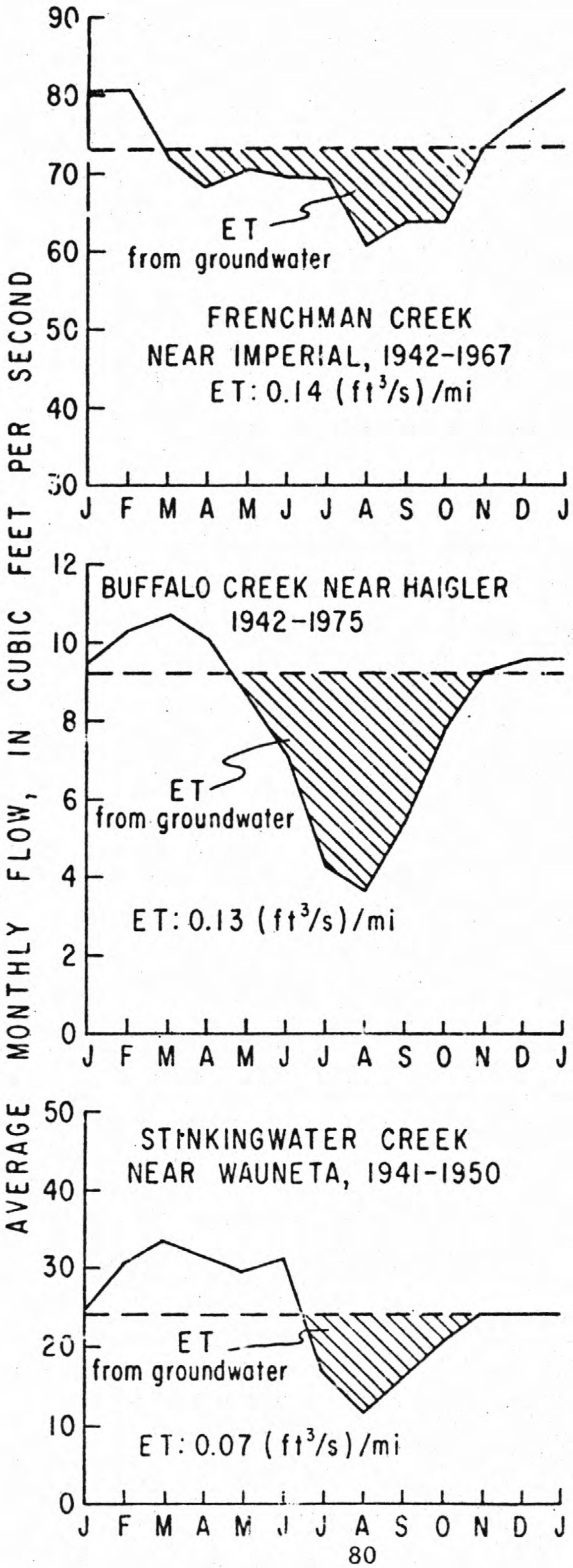


Figure 26.--Evapotranspiration discharge from stream valleys, computed using average monthly streamflow hydrographs for selected stations.

Leakage.--Leakage to or from the underlying Chadron sand aquifer may also account for inflow to and outflow from the Ogallala aquifer. Little information is available on hydraulic connection and head gradient between the two aquifers. In 1976 two observation wells were drilled at one point in northern Perkins County to determine the direction and magnitude of head gradient between the two aquifers (fig. 22). One well was screened in the bottom 20 ft of the Ogallala aquifer and equipped with a digital recorder. The other well, screened in the bottom 10 ft of the 30 ft of Chadron sand present at the site, was measured monthly. Chemical analyses of water from the two wells indicated they are open to the two different aquifers (R. A. Engberg, U.S. Geological Survey, oral commun.). At this site, no significant head difference existed during the period from January to May 1976 (fig. 22). However, as the irrigation season progressed, a maximum head difference of 20 ft was produced. This difference is thought to have been caused by pumping of a well about 2.2 mi north of the observation wells (inset, fig 22), which is the only well in the immediate vicinity which is completed in both the Ogallala and Chadron sand aquifers. The magnitude of leakage at this site cannot be calculated as no values are available for the vertical permeability of the silts, siltstones, and claystones between the two aquifers. At this site, then, leakage between the two aquifers is seasonal, occurring from the Chadron sand to the Ogallala as pumping progresses. This situation may or may not exist over the remaining area possibly underlain by the Chadron sand aquifer (fig. 17).

Abstraction by wells.--Withdrawal of water through low-yield stock and domestic wells and high-yield irrigation and municipal wells represents the largest widespread alteration of balance between natural inflow to and outflow from the Ogallala aquifer. Withdrawals from low-yield stock and domestic wells are not measured, but may be estimated. Using one-half the combined withdrawal estimate for Chase County, Nebr., and Phillips County, Colo. (Cardwell and Jenkins (1963, p. 86) as being representative of each of the counties in the study area and assuming that this use has not changed significantly since 1952, total withdrawal from low-yield wells of 2,000 acre-ft/yr can be considered to be consumptively used.

Although the annual volume of water withdrawn from them is small, these wells are important sources of water. Many of the wells are completed only in the top few feet of the Ogallala aquifer. Water-level declines due to withdrawal by irrigation wells have resulted in the failure of many of the low-yield wells and is a cause for concern in the study area.

Withdrawal from high-yield irrigation and municipal wells accounts for the largest outflow from the Ogallala aquifer. In 1952 there were less than 220 irrigation wells in the study area, most of which were along the South Platte River in Keith County. The number of irrigation wells increased to 2,617 by May of 1974. Figure 27 shows locations of irrigation and municipal wells and the annual cumulative number of wells installed south of the Platte River in Chase, Dundy, Perkins, and Keith Counties.

Until 1973, no records were kept of pumpage from irrigation wells. Since then, about 40 wells have been metered by irrigators in cooperation with the Southwest Nebraska Irrigation District, the Upper Republican Natural Resources District, and the University of Nebraska. Pumpage from these wells is summarized in table 8.

Construction and calibration of a model of the saturated zone requires knowledge of both temporal and spatial variations in net withdrawal from the aquifer over the entire 1952-75 period. The values in table 8 are too few and cover too short a time period for this purpose. Consequently, two methods were used to compute withdrawals from the aquifer. The first of these uses records of power consumed by irrigation pumping plants; the second computes consumptive use of irrigated crops.

Withdrawal from irrigation wells was computed for the 1960-74 period using electric power and natural gas fuel consumption records. These records were supplied by the Highline, Midwest, and Southwest Nebraska Electrical Cooperatives and Kansas-Nebraska Natural Gas Company. Table 9 summarizes the number of wells using each power type. The remaining pumping plants are assumed to be powered by diesel, propane, and gasoline.

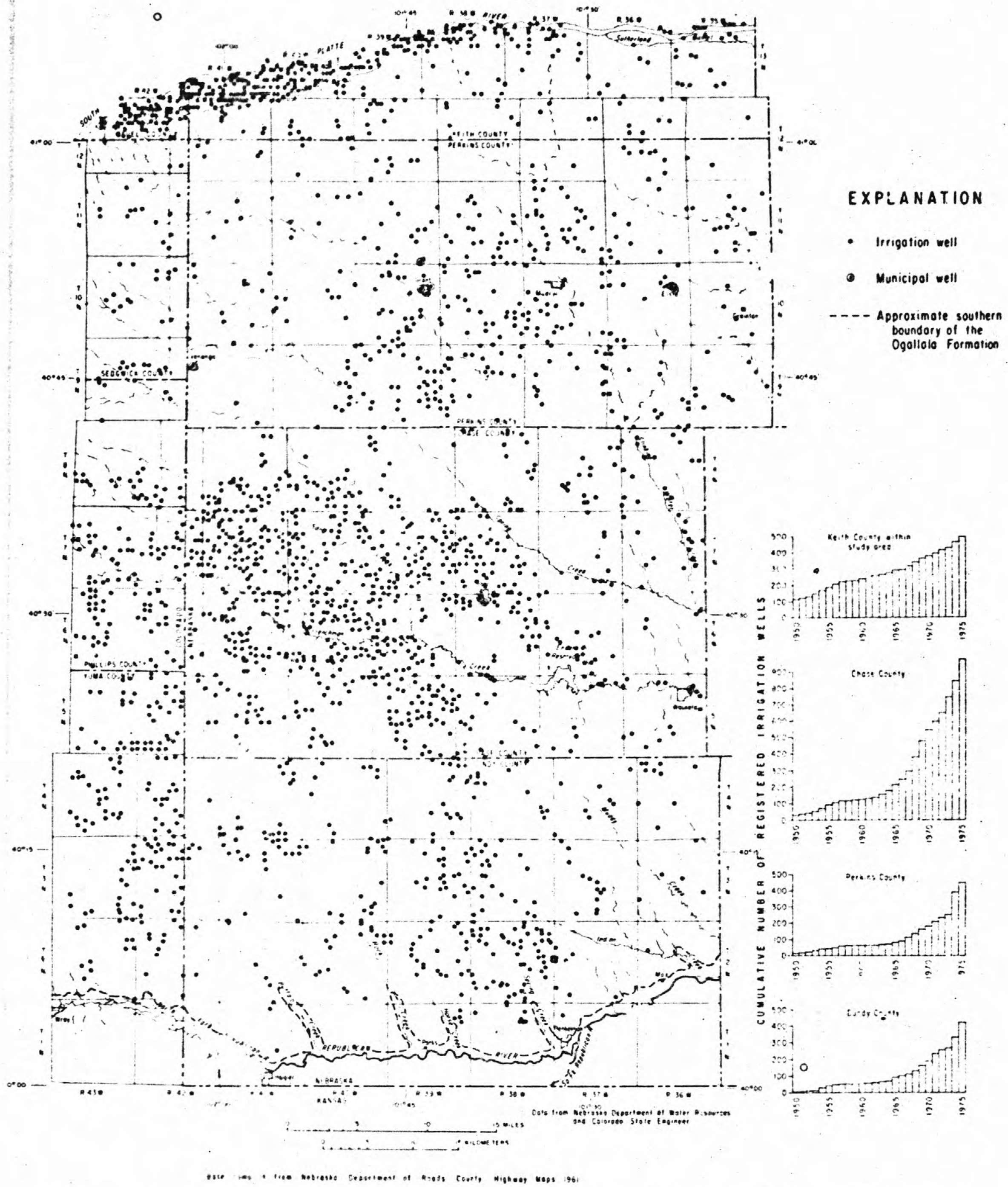


Figure 27.--Locations of irrigation and municipal wells and the annual cumulative number of wells installed south of the South Platte River in Chase, Dundy, Perkins, and Keith Counties.

Table 8.--Pumpage from metered wells, 1973-75

Well location	1973 Volume pumped (acre-ft)	1973 Unit pumpage (ft)	1974 Volume pumped (acre-ft)	1974 Unit pumpage (ft)	1975 Volume pumped (acre-ft)	1975 Unit pumpage (ft)
2N-38W- 3A	98.9	1.24	130.0	1.63	112.4	1.41
2N-38W- 3D	149.6	1.11	187.6	1.39	166.0	1.23
2N-38W- 6C	146.7	1.17	161.8	1.29	243.4	1.95
2N-38W-12A	156.6	1.84	173.4	2.04	179.6	2.11
2N-39W-17	114.0	.89	125.1	.98	148.8	1.16
2N-39W-26B	147.6	1.14	156.5	1.20	163.6	1.26
2N-40W-35C	79.7	.80	118.5	1.19	139.3	1.39
3N-36W-22B	369.0	1.35	416.7	2.08	344.0	1.72
3N-37W-16C	71.5	1.02	105.6	1.51	163.9	1.17
3N-37W-22C	63.1	.51	81.2	.66	113.7	.92
3N-37W-27B	136.0	1.09	156.0	1.25	152.3	1.22
3N-38W-10A	238.2	1.12	206.5	.97	....	......
3N-39W- 6B	105.5	.78	96.5	.72	49.7	.37
3N-39W-12C	129.1	1.01	142.8	1.12	128.0	1.00
3N-39W-14A	134.7	1.00	150.6	1.12	164.7	1.22
4N-37W-28A	272.4	1.41	364.0	1.89	348.7	1.81
4N-37W-28D	....	......	294.0	1.84	240.4	1.50
4N-38W- 7D	....	......	230.4	1.84	219.9	1.76
4N-38W-25B	134.0	1.34	158.4	1.27	185.6	1.49
4N-39W-27B	148.7	1.21	166.0	1.35	139.5	1.13
4N-40W-32A	188.8	.94	182.6	.91	200.2	1.00
4N-41W-32D	294.3	2.26	184.3	1.37	309.8	2.30
5N-37W-28D	387.4	1.05	417.4	1.28	390.5	1.20
5N-38W-19C	132.8	1.01	138.3	1.05	180.0	1.36
5N-38W-19D	379.4	1.72	386.3	1.76	315.7	1.44
5N-38W-25A	249.7	1.14	294.6	1.34	333.3	1.52
5N-39W- 2C	296.4	1.88	....	......	316.5	1.60
5N-40W- 9A	....	......	338.0	1.30	....	....
6N-37W-28B	....	......	117.7	.98	145.8	1.22
6N-38W-13A	....	......	120.0	.86	165.9	1.19
6N-38W-23D	126.5	1.10	137.9	1.20	181.7	1.58
6N-41W-17B	178.9	1.38	207.8	1.60	244.6	1.88
7N-38W-25B	160.0	1.19	155.8	1.16	175.1	1.31
7N-38W-30D	191.2	.96	219.0	1.10	244.5	1.22
7N-38W-31D	131.7	1.22	106.6	.89	122.9	1.02

Table 8.--Pumpage from metered wells, 1973-75--Continued

Well location	Volume pumped (acre-ft)	Unit pumpage (ft)	Volume pumped (acre-ft)	Unit pumpage (ft)	Volume pumped (acre-ft)	Unit pumpage (ft)
7N-39W-14B	256.4	1.35	249.8	1.28	278.2	1.43
7N-39W-21B	....	......	258.1	1.02	....	......
7N-39W-21C	....	......	298.5	1.18	....	......
7N-40W-29D	127.7	1.07	156.9	1.32	198.1	1.67
8N-36W-16B	235.1	2.03	233.2	2.01	245.2	2.11
8N-37W-34A	379.4	3.45	....	......	....	......
8N-39W-32B	261.7	1.27	253.9	1.23	307.6	1.49
8N-40W-22D	174.1	1.58	289.5	2.63	281.5	2.56
9N-41W- 6D	....	......	159.0	1.22	235.4	1.81
10N-37W-33C	....	......	164.8	1.24	189.8	.84

(Source: D. Watts, University of Nebraska, written commun., 1975)

Table 9.--Power types used for irrigation wells

County	Total registered wells (1974)	Electric		Natural gas		Remainder	
		Number	Percent of total	Number	Percent of total	Number	Percent of total
Chase	890	388	44	178	20	324	36
Dundy	435	141	32	59	14	235	54
Perkins	393	279	71	4	18	110	28
Keith	511	108	21	0	0	403	79
Phillips, Colo.[1]	139	116	84	3	2	20	14
Sedgwick, Colo.[1]	53	33	62	0	0	20	38
Yuma, Colo.[1]	196	94	48	0	0	102	52

[1] Within study area only.

To convert power or fuel consumption to water pumped, two approaches are possible. The first uses a direct conversion factor and an equation of the form:

$$Q_w = CP \tag{14}$$

where Q_w = water pumped in a given period, L^3,

P = power or fuel consumed in the same time period,

C = a conversion factor.

This method was applied to the power records for 21 of the wells listed in table 8. The sample size was small, and the standard deviation of C was large. This result is consistent with the findings of Kastner (1974) and Longenbaugh (1970). Variations in pumping-plant design and repair as well as operating conditions are highly variable and cause large variations in the conversion factors. Consequently, this approach was abandoned.

The second method of computing pumpage from energy used in pumping uses the following two equations which account for some of the variability described above:

for electric plants,

$$Q_w = \frac{0.977\ E_{kw}\ E}{TDH} \tag{15a}$$

where Q_w = water pumped in an irrigation season in acre-ft,

E_{kw} = kilowatt hours of electricity used over the irrigation season,

E = overall pumping plant efficiency, dimensionless,

TDH = total dynamic head against which the pumping plant is operating, in feet;

for gas plants,

$$Q_w = \frac{0.272\ E_{ng}\ E}{TDH} \tag{15b}$$

where Q_w = acre-ft of water pumped during irrigation season,

E_{ng} = cubic feet of natural gas consumed during the irrigation season,

E = overall pumping plant efficiency, dimensionless,

TDH = total dynamic head, feet.

Both of these equations were adapted from those used by Longenbaugh (1970). The constants in both equations account for unit conversions. The constant in equation 15b assumes a fuel rating of 950 Btu/ft^3 for natural gas, which is the standard for interstate transportation of this fuel (Longenbaugh, 1970).

Pumping-plant efficiencies (E) were determined for both gas and electric plants by applying equations 15 to metered wells in Chase and Dundy Counties, Nebr. (table 8). The efficiencies thus computed are shown in table 10 along with efficiencies computed for wells in north-eastern Colorado (Longenbaugh, 1970) and in Perkins County, Nebr. (Wayne Sides, Midwest Electrical Cooperative, written commun., 1975).

Table 10.--Overall pumping-plant efficiencies

Locations	Natural gas			Electric		
	Number tested	Mean	Standard deviation	Number tested	Mean	Standard deviation
Chase and Dundy Counties, Nebr.	9	0.133	0.046	12	0.629	0.074
Phillips, Yuma, Sedgwick, and Washington Counties, Colo.	17	.141	.022	55	.550	.129
Perkins County Nebr.	------	-----	---------	13	.645	.074

Based on the results in table 10, an assumed average value of 0.60 for electric plants and 0.14 for natural gas plants was chosen for overall pumping-plant efficiencies to apply equations 15 to power records.

Total dynamic head in equations 15 was computed using the sum of the pumping lift (pumping water level) reported on well registrations and the head loss due to the irrigation delivery systems. For gravity systems, the delivery system head loss was assumed to be equivalent to 5 ft of water. For center-pivot systems, an assumed normal operating pressure of 70 lb/in^2 or 162 ft of water was used for the delivery system head loss. Wells which deliver water to center-pivot systems were identified by locating all center pivots on color infrared imagery and comparing these locations with registered well locations (fig. 27). This assumes that the type of system present in May 1975 was the same type of system present since the installation of the well and pumping plant.

Using power records to compute pumpage from the Ogallala aquifer resulted in more complete temporal and spatial documentation than that provided by the metered wells. However, the percentage of wells for which records were available did not give detail considered sufficient for model-testing purposes. Also, pumpage data computed using power and fuel consumption data went back only to 1960, and the completeness of data prior to 1965 was questionable in some instances.

The most limiting problem in using pumpage computed from power requirements for model testing is the requirement to use net withdrawal from the aquifer for a given time period. The relationship between total pumpage and net withdrawal or consumptive use supplied with ground water must thus be determined to use total pumpage data for model testing. As described below and reported by D. Watts, University of Nebraska (oral commun., 1976), no relationship can be established between these two variables from records of metered wells and power records in the study area.

However, it was felt that pumpage computed from power consumption should give a reasonable estimate of total yearly pumpage for the study area from 1960 to 1974. The annual pumpage per well was computed for those wells for which data were available. These annual values were then multiplied by the total number of registered wells in the study area in each year. The results are given in table 11.

Table 11.--Computed pumpage and consumptive-irrigation requirements of crops from 1960 to 1974

Year	Total withdrawal from power records (acre-ft)	Number of records	Average withdrawal per well (acre-ft)	Total number of wells in study area	Total withdrawals (acre-ft)	Consumptive irrigation requirement (acre-ft)
1960	9,062	37	245	509	124,705	32,942
1961	7,269	41	177	533	94,341	20,752
1962	5,605	47	119	545	64,855	5,195
1963	15,698	51	308	585	180,180	58,029
1964	16,839	56	301	637	191,737	58,341
1965	12,053	73	165	723	119,295	22,150
1966	15,672	93	168	785	131,880	33,562
1967	28,515	144	198	902	178,596	32,486
1968	78,074	271	288	1,092	314,496	121,732
1969	110,796	398	278	1,295	360,010	125,362
1970	143,680	469	306	1,462	447,372	188,827
1971	133,428	593	225	1,624	365,400	120,114
1972	142,329	657	217	1,746	378,882	119,064
1973	166,523	775	215	1,949	419,035	167,079
1974	242,646	920	264	2,243	592,152	246,602
			Total		3,962,936	1,627,399

Net withdrawal from the aquifer over the 1952-75 period was computed by determining monthly values of consumptive use for the vegetative cover (pl. 5), soils, and topography (fig. 6) present at each modeled point in the study area. The methodology used is described in the subsequent section on the model of the soil zone. Using consumptive use as the net withdrawal from the aquifer introduces the assumption that water pumped in excess of consumptive use returns to the aquifer within the time periods of a few months used in the modeling of the saturated zone. This assumption is considered nonrestrictive although data availability precludes determining transit times for deep percolation to reach the water table. The volume of consumptive use at a given well location or irrigated area was determined by multiplying average annual irrigated area per well determined for each county (fig. 28) by the unit consumptive demand. These areas were determined annually by

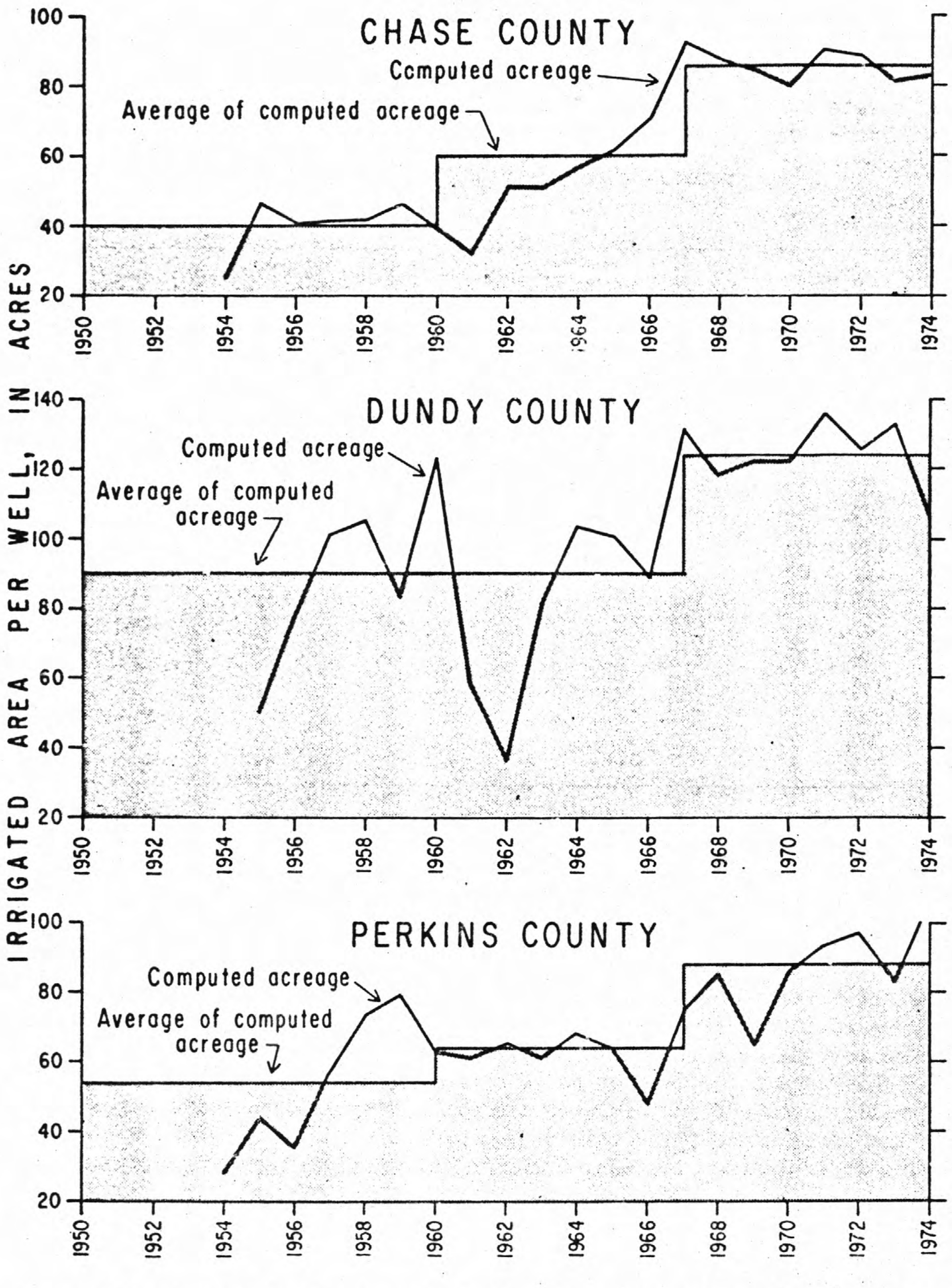


Figure 28.--Computed and averaged irrigated acreage per well, 1952 through 1975, for Chase, Dundy, and Perkins Counties.

dividing the total area of harvested crops irrigated with ground water in each county by the number of registered wells in each county. The area irrigated with ground water was computed by subtracting the assumed annual area irrigated with surface water in each county (Nebraska Department of Water Resources, annual) from the harvested irrigated crops (Nebraska Department of Agriculture, annual). The total annual volume of consumptive use of ground water for the study area is shown in table 11 for the 1960-74 period.

The relationship between annual pumpage computed using fuel consumption and annual consumptive use for individual wells was examined by linear regression analysis on a yearly basis and grouping the values over the entire 1960-74 period. No significant relationship was found, and coefficients of determination were generally less than 0.10.

Using the total volume pumped and the total consumptive use for the 1960-74 period (table 11), excess pumpage or assumed return flow amounts to about 50 percent of total pumpage. This percentage is high compared to results of other studies (Huntoon, 1974, and Lappala, in prep.). A commonly accepted value for return flow of ground water is 30 to 40 percent. To obtain this percentage from table 11, the total pumpage computed from power records would have to be reduced by about 20 percent. Reductions of this magnitude could be attained if the pumping-plant efficiency were lowered and(or) the total dynamic head (pumping water level) were raised. The latter is considered the most plausible because reported pumping water levels are probably conservative and do not account for permanent water-level changes occuring since well installation.

The Chadron sand aquifer

The Chadron sand aquifer comprises very fine to very coarse sands and gravels in the lower part of the Chadron Formation of the White River Group of Oligocene age. These sands are channel deposits (Cardwell and Jenkins, 1963, p. 35) and as such are discontinuous. The occurrence of this aquifer is limited to the northern one-third of the study area (fig. 17). Aquifer thickness locally may be as great as 50 ft, but generally it is less.

Many irrigation wells in southern Perkins and northern Chase Counties are screened in both the Chadron sand and Ogallala aquifers.

The contribution of water from the Chadron sand to total well yield is unknown but it is probably relatively small where the aquifer consists principally of very fine sand. Examination of test-hole and drillers' logs shows this fine-grained characteristic to be common. Particle-size analyses of samples from test hole 11N-39W-1AAAA in Perkins County show that at this location 70 percent of the sand is finer than 0.1 mm.

The Chadron sand aquifer is connected hydraulically with the overlying Ogallala aquifer. The two may be in direct contact or may be separated by as much as 100 to 200 ft of silt, siltstones, and claystones. The nature of the interchange of water between the two aquifers is not known over all areas where both exist.

Based upon sample logs of test holes, the hydraulic conductivity of the Chadron sand is about 20 to 25 ft/d. Total porosity is probably about 0.35 (Johnson, 1967).

Owing to the lack of areal and stratigraphic definition and the apparently minor importance of the Chadron sand in yielding water to irrigation wells, this aquifer is not included implicitly in model analyses of the saturated zone.

MATHEMATICAL FORMULATION AND TESTING OF MODELS OF THE HYDROGEOLOGIC SYSTEM

Movement and storage of water within each of the segments of the hydrogeologic cycle is described by the mathematical statement of the conservation of mass: Over a closed region in a n-dimensional space, the water input must equal the water output after accounting for changes in water stored within the region and the strengths of sources or sinks that produce or withdraw water from the region.

In differential equation form, over a three-dimensional region (Eskinazi, 1975):

$$\frac{\partial \rho}{\partial t} + W = \frac{\partial}{\partial x_i} (\rho \vec{V}) \qquad i = 1,2,3 \tag{16}$$

where ρ = mass density of water, ML^{-3},

x_i = Cartesian coordinate directions, L, $i = 1,2,3$

$\vec{V}$ = fluid velocity, LT^{-1},

t = time, T,

W = source or sink term, $ML^{-3}T^{-1}$.

This general equation is augmented by one or more supplementary relationships, such as Darcy's law, which relate ρ and $(\rho\vec{V})$ to the physical dimensions and measurable properties of the segment of the hydrogeologic cycle being considered.

The following sections describe these relationships, summarize the mathematically formulated conceptual models, and describe methods used for their solution.

Model of the Soil Zone

This segment of the hydrogeologic system was modeled to compute (1) crop water requirements used in stressing the saturated zone model and (2) recharge to be routed through the unsaturated zone to the saturated zone. Movement of water through the soil zone is considered to be one-dimensional (vertical) at a given point in the study area. The mathematical model used to describe this segment is the following differential form of the continuity equation derived for a constant density fluid:

$$\frac{dq}{dz} = \frac{dS}{dt} \tag{17}$$

A simple difference form of this equation is:

$$\frac{\Delta q}{\Delta z} = \frac{\Delta S}{\Delta t} \tag{18}$$

where Δz = thickness of the zone affected by evapotranspiration during the time step t, L,

Δt = increment of time, T,

ΔS = dimensionless storage = $\frac{\text{volume stored}}{\text{total storage capacity}}$,

Δq = unit rate of water flux, LT^{-1}.

Recognizing that Δq can be written as $\Delta L/\Delta t$ where L is a scalar measure of input with dimension of length (e.g., inches of precipitation or evapotranspiration) then equation 18 can be written

$$\Delta L = (\Delta z)(\Delta S) \tag{19}$$

Defining ΔL as the algebraic sum of infiltration (I), evapotranspiration (ET), and the change in dimensionless storage (ΔS) as $S_i - S_{i-1}$, where i is an index on the time periods being considered, equation 4 written for S_i becomes

$$S_i = S_{i-1} - \frac{1}{\Delta z}(I - ET) \tag{20}$$

The definition sketch of the soil zone in figure 29 shows the relationship of the variables in equation 20. This equation holds except when the storage capacity of the soil zone is exceeded.

Rigorous treatments implicitly consider the storage capacity as a continuous function of the tension with which water is held to the soil matrix. This tension is also referred to as the capillary pressure (Pc), which is a negative pressure head when using atmospheric pressure as the datum.

Field capacity or the moisture content corresponding to Pc = 0.33 atmospheres (Baver, 1956) is commonly assumed to be the upper limit to soil-water storage held against gravity drainage. This assumption is valid when considering soil-zone depths of less than 5 ft and time steps of 1 month. When S_i computed with equation 20 exceeds field capacity, deep percolation equal to the excess is simulated and S_i is set to field capacity.

Equation 20 was solved on a monthly basis using a FORTRAN IV computer program modified by the author from one used by the U.S. Bureau of Reclamation. Use of this program requires determination of monthly infiltration computed from measured precipitation. This was accomplished by the use of curves (fig. 30) which were derived from empirical monthly rainfall-runoff curves for conditions of varying soils, topography, and crop type by the Agricultural Research Service (ARS) at Rosemont, Nebr., (F. J. Otradovsky, U.S. Bureau of Reclamation, written commun., 1975). Conditions used in the derivation of the curves were assumed to be sufficiently similar to those in the study area to permit their use. The difference between monthly precipitation and infiltration is assumed to be surface runoff.

The program used to solve equation 20 also incorporates a provision for delay of infiltration and runoff computation when air temperatures are below 27°F by placing all precipitation in snow storage. Snow storage is maintained until either temperatures exceed 27°F or a maximum number of days have elapsed. This maximum-allowed time of snow storage is determined from the published weather service records of average number of days with snow on the ground for a given year. A value of 45 days was used for all stations over all years for this study based on examination of records at Imperial. This feature affects the monthly distribution of moisture available for infiltration but has little effect on seasonal recharge or consumptive-irrigation requirements. Sublimation of snow cover is ignored.

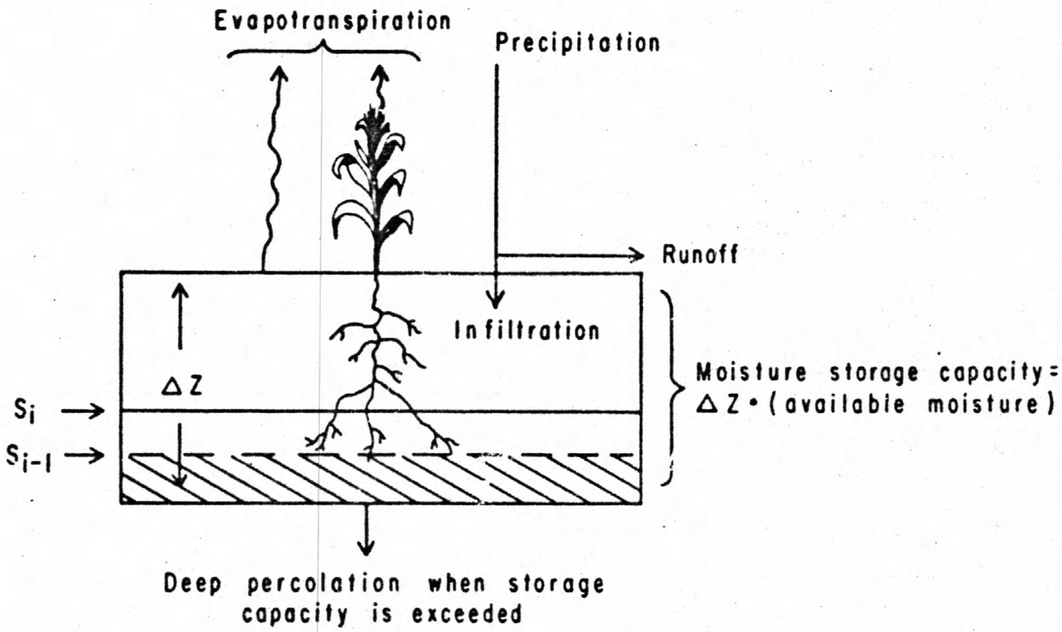


Figure 29.--Definition of the soil zone.

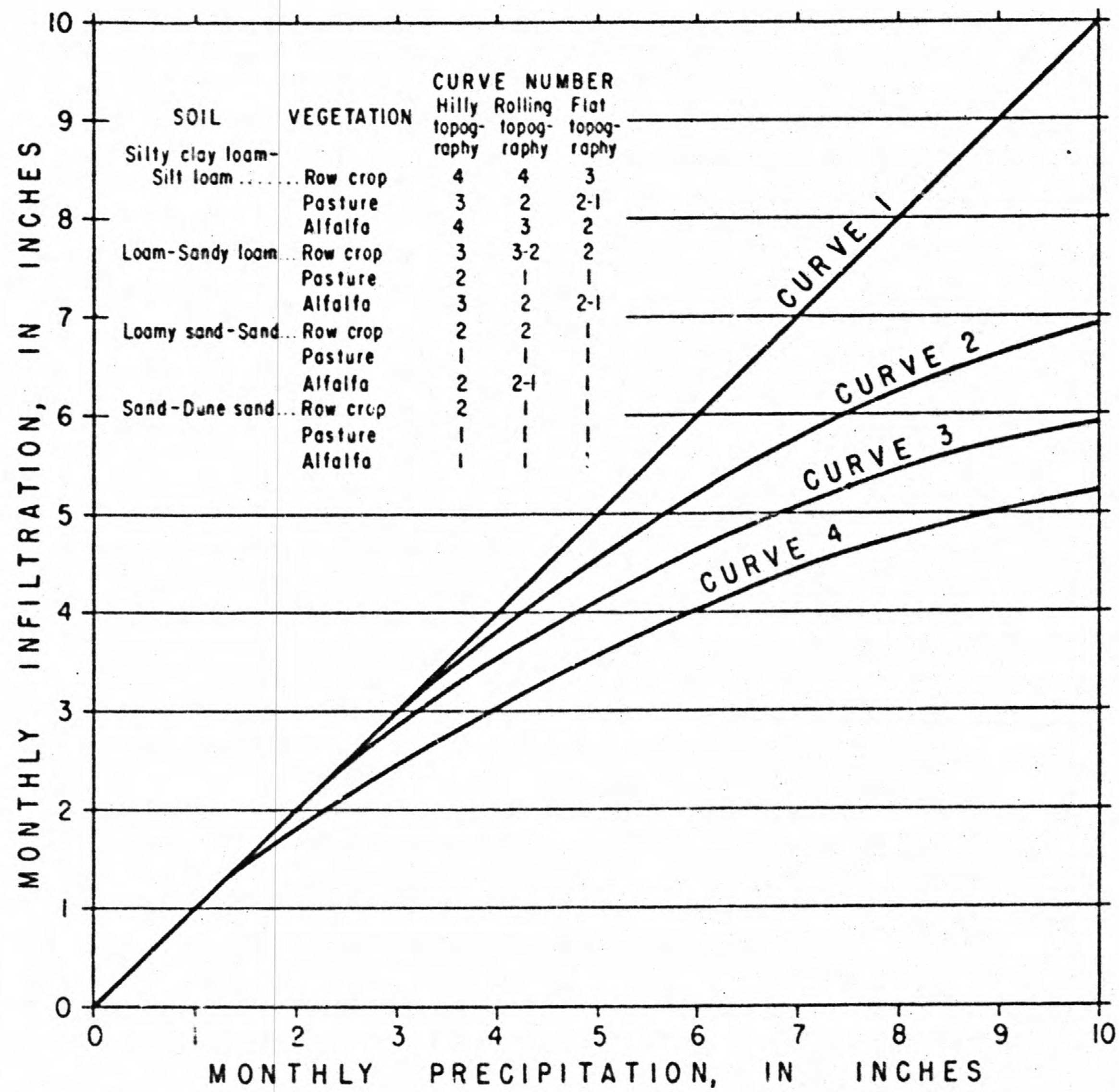


Figure 30.--Monthly rainfall infiltration for different crops, soils and topography.

The model of the soil zone was used to compute the consumptive requirement of different crops under both dryland and irrigated conditions. Empirical ratios of actual ET to potential ET shown in figure 31 for various crops at different growth stages were used to compute actual ET in the model. These crop-coefficient curves were those used in the Platte River Level B Study (Missouri River Basin Commission, 1975) and were adapted from published curves for five crops by F. J. Otradovsky, U.S. Bureau of Reclamation (written commun., 1974). Fallow land is considered to be subject to bare-soil evaporation only and is represented by a horizontal line in figure 31.

Under dryland conditions when ET demand exceeded stored soil moisture plus infiltration in a given month, the excess was accounted for by reducing ET so that soil moisture did not fall below 25 percent of the available moisture capacity. This lower limit is arbitrary but keeps the soil moisture above the permanent wilting point and thus crudely simulates moisture-stress protection mechanisms of the vegetative cover. Under irrigated conditions, a consumptive-irrigation requirement was computed as the amount of water required to maintain soil moisture at 50 percent of the moisture-holding capacity. Again, the 50 percent limit is arbitrary, but it is considered reasonable based on other studies (D. Watts, University of Nebraska, oral commun., 1975).

Figure 32 is a flow chart of the soil-zone model showing the essential steps used in solving equation 20 and the required inputs and outputs. The model was operated monthly in this study, with the monthly values being summed over an irrigation (June-August) and a nonirrigation (September-March) season for use as input to the saturated-zone model. For this study the model was run for all crop types, soil categories, and climatic stations simultaneously, and values of dryland and irrigated recharge and consumptive-irrigation requirements were stored for subsequent use by the saturated-zone model.

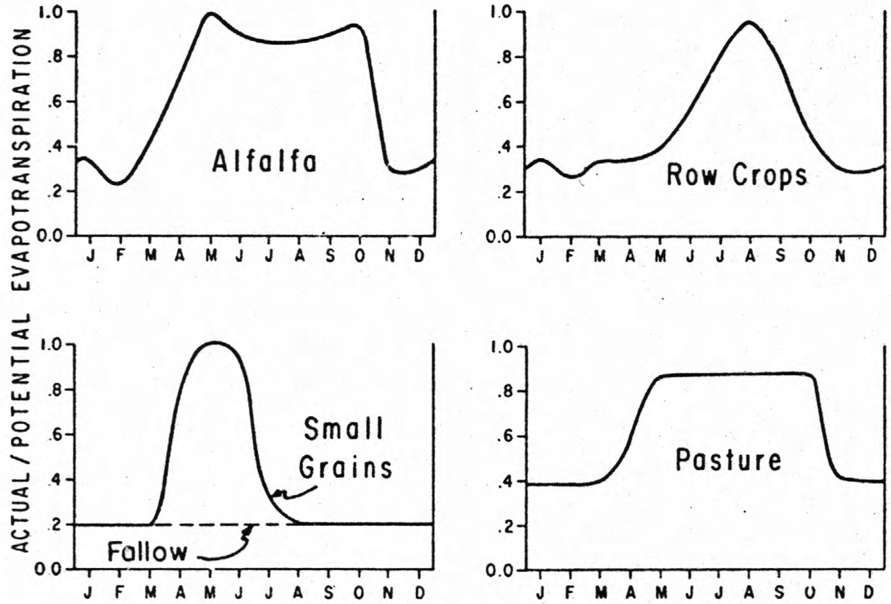


Figure 31.--Monthly ratio of actual to potential evapotranspiration for crop types in the study area.

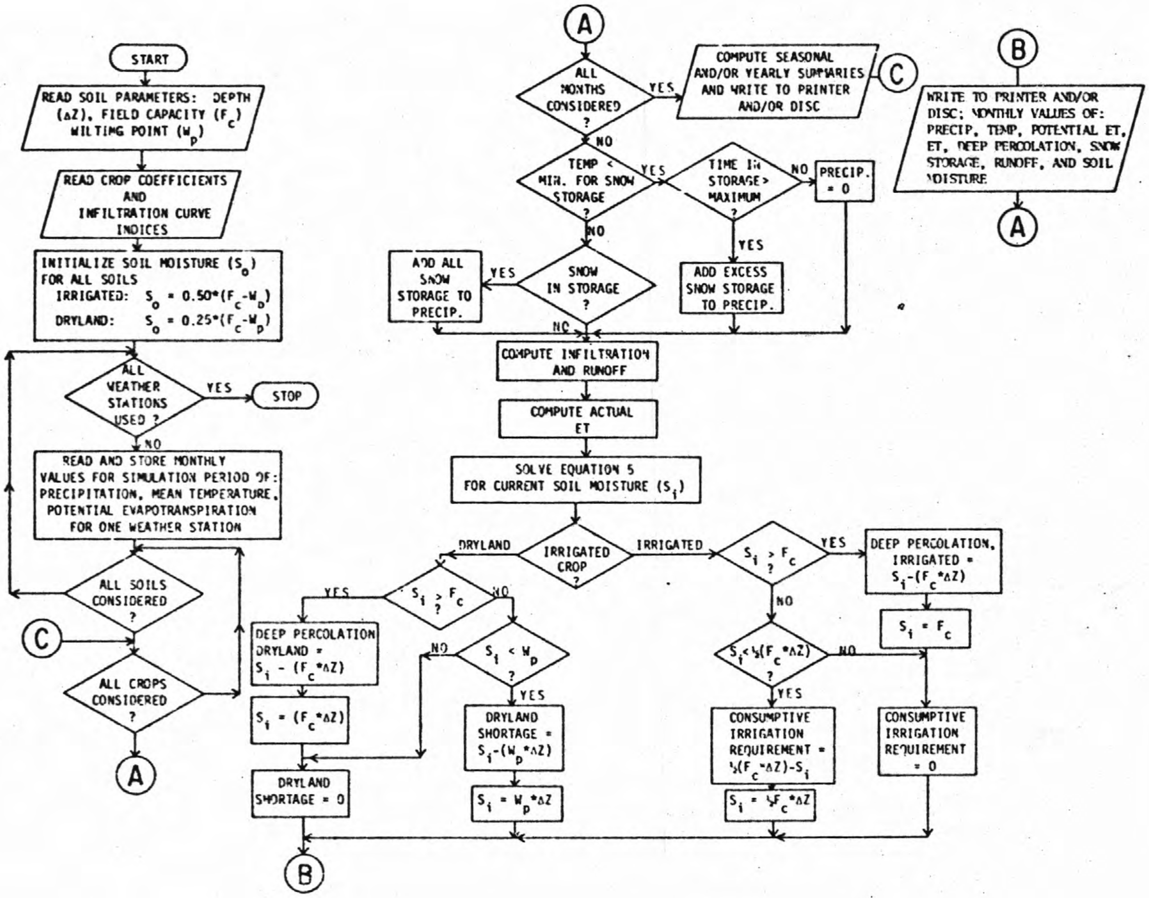


Figure 32.--Flow chart of the soil-zone model.

Model of the Unsaturated Zone

Formal mathematical modeling of the unsaturated zone was not included in this study. Determining the water requirements of the soil zone and observing the effects of supplying these requirements from the saturated zone are the principal objectives of the study. The time delay of deep percolation through the unsaturated zone is so short relative to the time period of interest (seasons to years) that it can be ignored. However, the possibility exists that there may be areas within the study area where this assumption is not valid. The principal restriction on modeling delays caused by redistribution of deep percolation is the lack of enough data on the functional relationships between (1) moisture content and pressure head, (2) moisture content and hydraulic conductivity, and (3) vertical distribution of moisture content over time to enable testing of a rigorous model of the unsaturated zone.

The mathematical model of the unsaturated zone used in this study is the steady-state continuity of mass flux:

$$Q_{DP} = Q_R \tag{21}$$

where Q_{DP} = inflow to the unsaturated zone as deep percolation, $M^{-3}LT^{-1}$,

Q_R = recharge to the aquifer at the base of the unsaturated zone, $M^{-3}LT^{-1}$.

Model of the Saturated Zone and Surface-Water System

The saturated-zone model is used in this study to evaluate the effects of given pumpage and recharge distributions in time and space. The effects are determined as changes in water levels or measures of water in storage in the aquifer and as changes to aquifer discharge or flow of springs and hydraulically connected streams and lakes.

The mathematical model used to describe the saturated zone is adapted from that given by Hantush (1966, p. 301) and Cooley (1974, p. 3-10). Developments given in these references show that the actual assumptions made in applying the final equation to measured field variables for an unconfined aquifer are not necessarily as restrictive as those embodied in the Dupuit-Forcheimer or horizontal flow assumptions (Cooley, 1974).

The flow equation results from combining the vertically integrated equation of continuity for mass flux of water with Darcy's law and considering compressive storage to be negligible compared to gravity drainage:

$$\frac{\partial}{\partial x_i}\left[\hat{K}\,(\hat{h} - z_1)\,\frac{\partial \hat{h}}{\partial x_i}\right] = \hat{\varepsilon}_D\,\frac{\partial \hat{h}}{\partial t} + \hat{q}_1 - \hat{q}_u + \frac{k'}{b'}\,(\hat{h} - H_r) + W \qquad (22)$$

$$i = 1,2$$

where

$$\hat{K} = \frac{\int_{z_1}^{z_u} K(x_i)\,dz}{\int_{z_1}^{z_u} dz} \qquad i = 1,2$$

and is equivalent to the method described by equation 10

$$h = \frac{\int_{z_1}^{z_u} h(x_i)\,dz}{\int_{z_u}^{z_1} dz} \qquad i = 1,2$$

which is the average hydraulic head produced when contouring field measurements of water levels screened in most or all of the interval from z_1 to z_u. Most wells used in determining the 1975 water-table configuration (fig. 15) and many used for the 1952 configuration in the study area satisfy this criterion.

x_i = a horizontal coordinate direction, $i = 1,2$

Remaining terms are defined in figure 33.

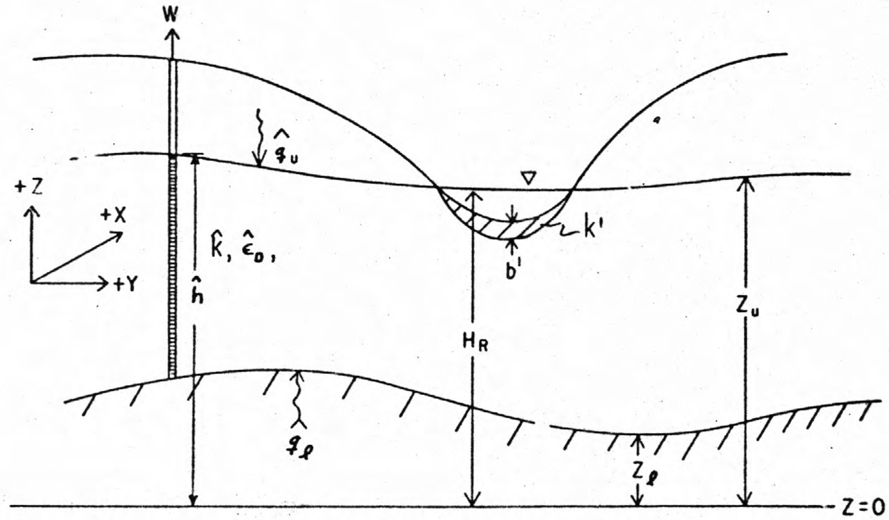


Figure 33.--Definition of variables for the saturated zone.

Boundary and initial conditions that are specified in conjunction with equation 22 for this study are the following:

(1) Specified head gradient boundary condition:

$$\frac{\partial \hat{h}}{\partial x_i} = C \qquad i = 1,2 \tag{22a}$$

(2) Initial head condition: $\hat{h}\,(x,y) = \hat{h}_o(x,y), t=0$ (22b)

or

(3) Initial flux condition:

$$\hat{q}_{l_o} - \hat{q}_{u_o} = \frac{\partial}{\partial x_i}\left[\hat{K}\,(\hat{h}_o - z_l)\,\frac{\partial \hat{h}_o}{\partial x_i}\right], t=0 \qquad i = 1,2 \tag{22c}$$

True specified head boundary conditions are not used in this study. They are replaced by considering the hydraulically connected surface-water systems (streams, lakes, or reservoirs) to be represented as an overlying aquifer with the free water surface elevation being H_r. This overlying aquifer is simulated as being separated from the main aquifer by a semiconfining bed (streambed) having thickness b' and vertical hydraulic conductivity k'. The flux between the surface-water system and the aquifer is then computed by writing the following form of Darcy's law across the streambed:

$$\hat{q}_s = \frac{k'}{b'}\,(\hat{h} - H_r) \tag{23}$$

The initial flux condition is required when it is desired to evaluate modeled changes in $\hat{h}$ due to changes in stresses imposed on the system rather than absolute value of $\hat{h}$ at given points in time (Konikow, 1976, and Trescott and others, 1976). This boundary condition is satisfied by either one of two approaches, both using equation 22c. The first commonly used approach is to determine values of $(\hat{q}_{l_o} - \hat{q}_{u_o})$ by examining the hydrologic operation of the soil and unsaturated zones prior to the time corresponding to the start of transient simulation of the saturated zone. These values are inserted into equation 22c, which is solved for $\hat{h}_o(x,y)$ using an iterative finite difference method (Pinder, 1969). This computed hydraulic head distribution is then compared with the assumed steady-state observed head distribution. Since the computed head distribution is independent of the heads used to initiate the solution to equation 22c (Konikow, 1976), differences between computed $\hat{h}$ and observed $\hat{h}$ are a result of:

1. Incorrect values of $(\hat{q}_{1_o} - \hat{q}_{u_o})$,

2. Incorrect values of $\hat{K}(\hat{h}_o - z_1)$,

3. Incorrect values of observed head values,

4. Combinations of 1 and 2,

5. Violations of one of the assumptions made in developing equation 22,

6. Errors due to iterative numerical solution of equation 22c.

It is implicitly assumed that errors due to violation of the assumptions and the numerical method are not restrictive for the remainder of this report unless specifically stated.

The advantage of this method is the independence of the computed heads on the initial head used to initiate the solution. The disadvantages are (1) the requirement for an accurate spatial distribution of $(\hat{q}_{u_o} - \hat{q}_{1_o})$ if adequacy of the hydraulic conductivity ($\hat{K}$) distribution is to be evaluated or an accurate distribution $\hat{K}$ if $(\hat{q}_{u_o} - \hat{q}_{1_o})$ is to be evaluated and (2) the iterative solution may require significant computer time and core storage space. If this approach is used exclusively to satisfy the initial flux condition, the computed values of $\hat{h}$ are used as the initial heads ($\hat{h}_o$) at the start of transient simulations. The stresses used in transient simulations are the difference between $(\hat{q}_{u_o} - \hat{q}_{1_o})$ used to compute $\hat{h}_o$ and the transient stresses.

The second approach used to satisfy the initial flux condition is to solve the finite difference analog to equation 22c for $(\hat{q}_{1_o} - \hat{q}_{u_o})$ directly using initial values of $\hat{h}_o$, $\hat{K}$, and z_1. The resultant net flux values represent recharge or discharge required at each point in the solution domain to maintain the $\hat{h}_o$ distribution with the given values of $\hat{K}$ and z_1. These net flux values can then be compared to observed values of aquifer discharge (base flow and spring flow) and observed or computed values of recharge using independent methods such as those described in the previous section. Differences between observed and computed values of net flux $(\hat{q}_{u_o} - \hat{q}_{1_o})$ are attributable to:

1. Incorrect values of $\hat{h}_o$,
2. Incorrect values of $\hat{K}\ (\hat{h}_o - z_1)$,
3. Combinations of 1 and 2,
4. Violation of assumptions used in deriving equation 22,
5. Errors due to direct numerical solution of equation 22c which are not removed by iteration.

The advantages of this solution method are: (1) when the $\hat{h}$ and z_1 distributions are well known, conclusions regarding the $\hat{K}$ distribution can be drawn by comparing known and computed net flux values; and (2) the direct solution is very fast and cheap, even for large areas. The disadvantages of the direct solution method are that (1) observed values of net flux are usually available only for obvious ground-water discharge areas such as springs and predominantly base flow-fed streams and (2) the method is relatively sensitive to the initial head $(\hat{h}_o)$ distribution, so undetected errors in measurement, interpolation, and spatial discretization may cause significant errors in computed flux.

When this method is used to satisfy the initial flux condition, the initial values of $\hat{h}_o$ are used to start transient simulations and the stresses imposed on the system should be the difference between assumed flux conditions at the start of transient analyses and the conditions at a given point in time during the transient analysis.

Neither of these methods of satisfying equation 22c was used exclusively for this study. Rather, both methods were used in an alternating fashion as discussed further in the section on model testing.

Iterative solutions to equation 23 were achieved for both steady-state and transient analyses by writing a finite-difference analog which was solved by the IADI (Iterative Alternating Direction Implicit) Method. The basic FORTRAN IV program to solve equation 22 was written by G. F. Pinder of the U.S. Geological Survey (1969) and modified by the author to permit application to field problems in the Platte River Level B Study (Missouri River Basin Commission, 1975) and this study.

In using the program to solve equation 22, the study area was divided into a network of 3,340 square finite-difference cells, 1 mi on a side (fig. 34). This network is arbitrary and does not correspond to the township-range-section system of land subdivision. Figure 34 also shows locations of finite-difference cells representing specified gradient boundaries and cells representing streams and reservoirs modeled as hydraulically connected to the aquifer.

In addition to the locations of these boundaries, input data to the model consisted of the following for each finite-difference cell:

(1) Initial average hydraulic head ($\hat{h}$) in 1952 expressed as elevation above mean sea level.

(2) Vertically averaged hydraulic conductivity ($\hat{K}$) (figs. 23 or 24).

(3) Altitude of the base of the aquifer (z_1) (fig. 17).

(4) Vertically averaged drainable porosity (ϵ_D) (fig. 21).

(5) Predominant vegetative type in 1975 (pl. 5).

(6) Soil-topographic type (fig. 6).

(7) Thiessen polygon index (fig. 6).

(8) Number of registered irrigation wells present in each year from 1952 to 1974.

A_I was constrained to be always less than 88 percent of the cell area or 560 acres. This limit was imposed to account for roads and areas not irrigated with a 140-acre center-pivot irrigation system.

To simulate the decline in well yield as aquifer thickness declines, the following scheme was used (E. P. Weeks, U.S. Geological Survey, written commun., 1975) for a given finite difference cell

$$W \leq 0.8\, N_w \;\; K(b^{n-1})^2 / \ln\,(\Delta x/\sqrt{r_{ew}\, \pi\, N_w}) \tag{26}$$

where W = nodal discharge, $L^{-3}T^{-1}$,

N_w = number of wells in the cell,

K = hydraulic conductivity of the cell, LT^{-1},

b^{n-1} = aquifer thickness for the previous time step, L,

x = dimension of finite-difference cell, L,

r_{ew} = effective radius of each of N_w wells in the cell, L, (1.0 ft was used in this study).

W was kept below the limit imposed by equation 26 until a minimum discharge equivalent to 0.10 ft^3/s was reached. Below this yield, wells were assumed to fail and were shut off.

Simulation of the surface-water ground-water interaction was incorporated by

1. Computing baseflow of streams as ground-water discharge using equation 22c solved for $(\hat{q}_u - \hat{q}_1)$ or equation 23 using computed steady-state heads. A network traversal algorithm was used to sum incremental flows along the stream system.

2. Allowing the addition of any surface-water inflow (e.g. South Platte River at Julesburg) or subtracting diversions from streamflow at any point along the stream network to determine total streamflow.

3. Allowing a stream reach to go dry or become disconnected from the aquifer when the leakage out of the stream exceeded the available flow along the reach. This decoupling was accomplished by setting the value of k' in equation 23 to zero for the next time step.

4. Allowing a stream to become live after drying up if $\hat{h}$ becomes greater than H_r. This accomplished by setting k' back to its original value.

Values of $\hat{h}$, or changes between $\hat{h}_0$ and $\hat{h}$ computed with the model at the end of specified time periods represent average values over the entire modeled cell. To more accurately simulate the water levels in irrigation wells within the cells, the following equation was adapted from Hantush (1966).

$$\Delta S = \frac{Q/2 \cdot \pi \cdot \hat{k}\, n(\hat{h} - z_1)\ \ln \Delta X}{\sqrt{r_w n \pi}} \tag{27}$$

where Q = flux rate for the finite-difference cell, L^3T^{-1},

ΔS = difference between the average change in $\hat{h}$ over the cell and the value at a distance r_w from the well,

r_w = effective radius of the well (taken as 1 foot in this study),

n = number of wells in the cell.

Output from the model of the saturated zone consists of the following at the end of each pumping period and(or) time step:

1. Hydraulic head in each finite difference cell.

2. Changes in hydraulic head from the start of simulation in each cell.

3. Changes in hydraulic head computed by adding ΔS from equation 27 to the average change in each cell.

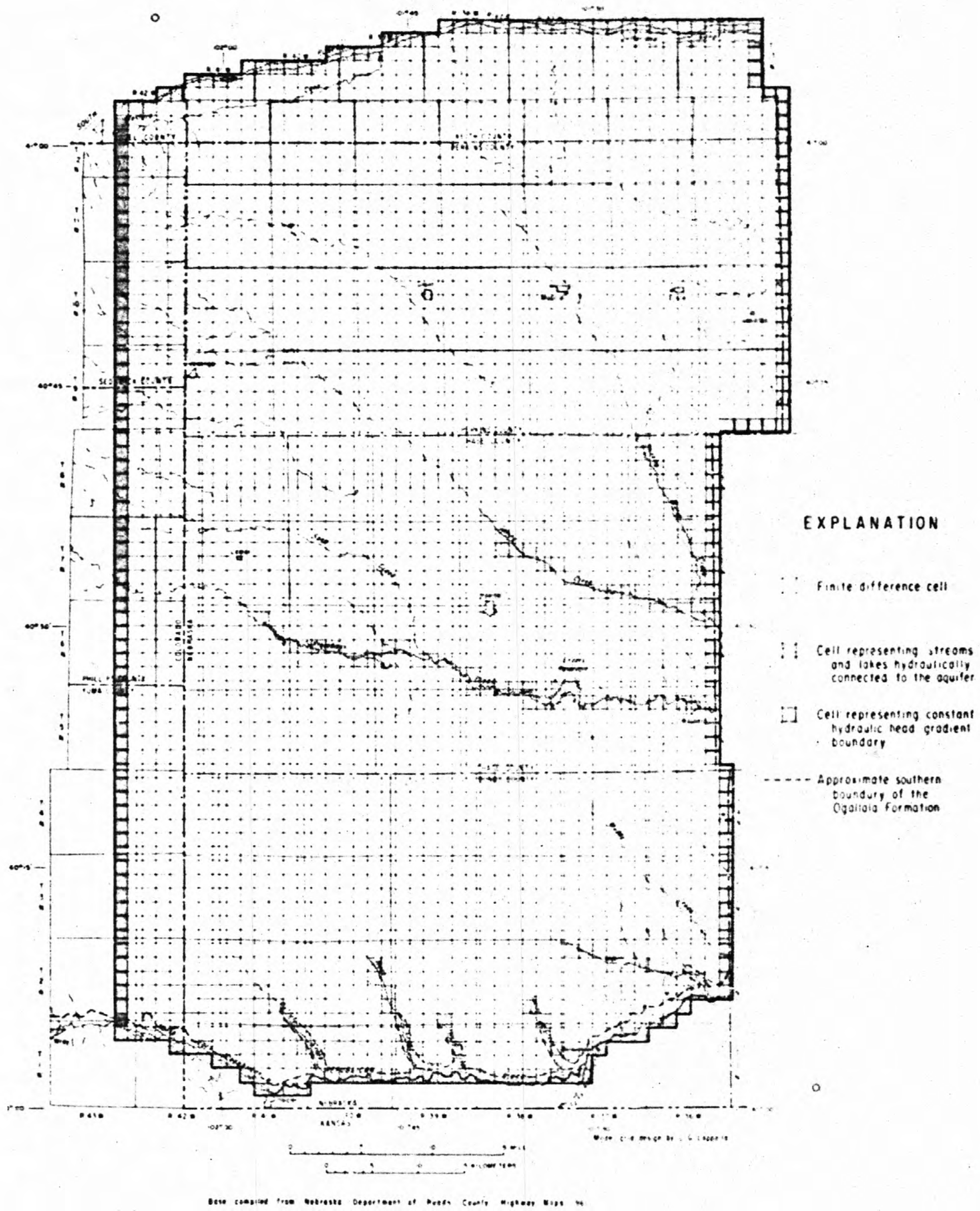


Figure 34.--Grid system used for the model of the saturated zone showing locations of hydraulically connected surface-water systems and specified hydraulic head gradient boundaries.

The above were discretized by overlaying the finite-difference network shown in figure 34 on maps of each of the above and picking an average or representative value for each cell. The program solves equation 22 by progressing forward in time by discrete time steps. These ranged from 1 to 3 months for this study. Before advancing to a new time step, iteration was used until the difference in computed values of $\hat{h}$ between two subsequent iterations was less than 0.01 ft for all finite-difference cells.

The time domain was also divided into pumping periods to adequately simulate the irrigation and nonirrigation seasons. The 3-month period from June through August was used for the irrigation season. Examination of hydrographs on plate 8 in the vicinity of irrigation wells showed this to be a reasonable approximation. For each irrigation (June-August) and nonirrigation (September-May) season, the net input or output to the aquifer (W) in each cell was computed as

$$W = Q_I \cdot A_I + Q_D \cdot A_D \quad . \tag{25}$$

where

Q_I = net flux (withdrawal-recharge under irrigated conditions determined with the soil model, LT^{-1},

A_I = area irrigated in the cell during the current season, L^2,

Q_D = net flux (withdrawal-recharge) under dryland conditions, LT^{-1},

A_D = area under dryland in the cell during the current season, L^2 (equal to cell area minus A_I).

Values of A_I were determined for each cell in each year by

$$A_I = N_w \cdot a \tag{25a}$$

where

N_w = number of registered irrigation wells in the cell in the given year,

a = area irrigated per well for the county in which the well is located in a given year (fig. 28).

4. Streamflow values at each point along the stream network(s).

5. Changes to streamflow at selected points of interest along the stream network(s).

6. Summary of total pumping and recharge applied during the current pumping period.

7. A mass balance summary of the cumulative values of total stress on the aquifer, amount of water derived from streams and lakes, and the amount of water derived from ground-water storage.

8. Total flow across specified gradient boundaries.

The relationship between models of the soil and saturated zones and the method of modeling the ground water-surface water interaction are shown schematically in figure 35.

Testing the Mathematical Models

The mathematical formulations of the conceptual models of the soil and saturated zones should be tested, and it is accepted practice to test them against the observed historical operation of the hydrologic system in order to minimize or evaluate the effect of the following:

1. Errors in the conceptual model.

2. Errors in measurement and interpretation of parameters describing the system.

3. Errors in mathematical formulation of the conceptual model.

The errors are measured by the difference between variables describing the observed and computed state of the modeled system such as hydraulic head and mass flux of water. The state of the hydrologic system is determined by functional relationships among a given set of parameters describing the system. Examples of these parameters are hydraulic conductivity and drainable porosity. The functional relationships are the mathematical formulations of the conceptual models (equations 20 and 22).

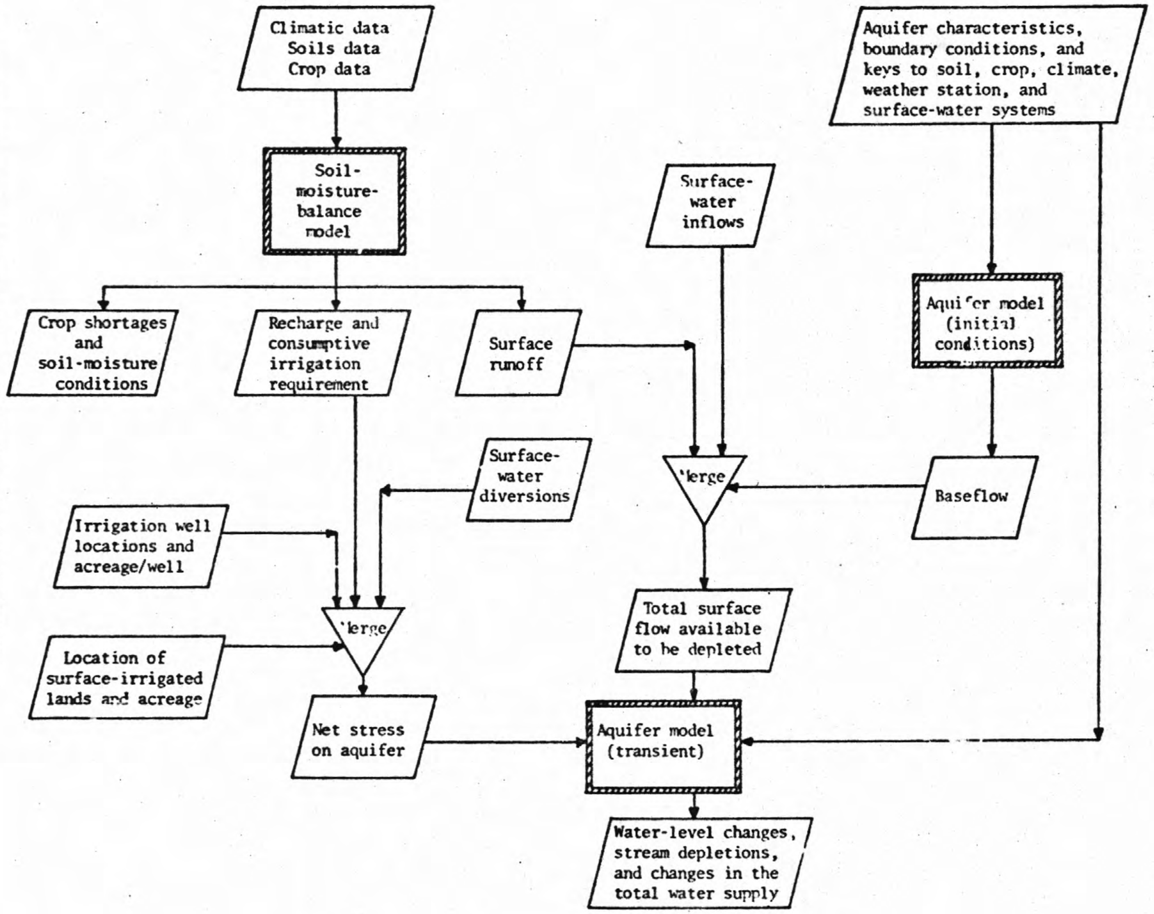


Figure 35.--Relationship between models of the soil and saturated zones.

The parameters examined in testing of the soil-zone model were:

1. Depth of the soil zone (Δz in equation 20).

2. Available moisture-holding capacity.

3. Infiltration-curve number used to separate runoff from precipitation.

4. Initial soil-moisture levels.

The response used for the soil-zone model was deep percolation which is considered equal to net recharge to the saturated zone.

The parameters used for testing the model of the saturated zone were:

1. Hydraulic conductivity ($\hat{K}$).

2. Drainable porosity $\hat{\varepsilon}_D$.

3. Net flux ($\hat{q}_u - \hat{q}_1 + W$).

The response variables for the saturated-zone model were hydraulic head and net flux.

Sensitivity analyses

The first step in testing both models was to determine the sensitivity of the state of the system to variations in the parameters describing the system as given above. The sensitivity analyses were accomplished by computing sensitivity functions of either an absolute value of a given state of the system or the difference between an observed and a computed state.

For sensitivity analyses of the soil-zone model, the sensitivity function used was the average annual dryland deep percolation between 1925 and 1952 over the entire study area, under the assumed vegetative cover in 1952 (see pl. 5). This value was weighted by the area covered by each soil type.

Values of net recharge were computed with the program used to solve equation 20 by varying each of the four parameters about the assumed correct value while holding the remaining three constant. Figure 36 shows that long-term average recharge is most sensitive to soil depth and least sensitive to initial conditions. The limits on soil depth were the least well known, but those used were bracket values available in the literature (Baver, 1956). The limits used for the sensitivity to the infiltration curve number were arbitrary, since the scatter in the data used to derive the curves in figure 30 was not known. The limits used for available moisture were the ranges given in table 2.

For sensitivity analyses of the saturated-zone model, the parameters used were input or output from the aquifer, hydraulic conductivity, and drainable porosity when solving equation 22 iteratively for hydraulic head.

The results of the sensitivity analyses are shown in figure 37. The hydraulic-conductivity distribution was varied over the entire model using simple multipliers. The 1925-52 average recharge was used for aquifer input. Recharge values were varied by using simple multipliers.

From figure 37, it is concluded that computed hydraulic head is more sensitive to recharge or discharge than to hydraulic conductivity. This is consistent with the findings of other studies in unconfined aquifers (Missouri River Basin Commission, 1975 and Luckey, 1973).

The sensitivity of computed hydraulic head and computed aquifer discharge to variations in drainable porosity ($\hat{\varepsilon}_D$) was evaluated by solving the transient form of the flow equation while varying $\hat{\varepsilon}_D$ (fig. 38).

Drainable porosity was varied about its original assumed correct value as shown in figure 21 by using simple multipliers. Direct comparison between sensitivity of computed head to $\hat{\varepsilon}_D$ and to hydraulic conductivity and recharge or discharge is not straightforward due to different sensitivity functions used. However, the relative sensitivity of computed hydraulic head to specific yield is of the same order of magnitude as the sensitivity to recharge. The high sensitivity of computed hydraulic head to specific yield is important as this property is the least well known at a given point in space and time. Also the sensitivity of hydraulic head to hydraulic conductivity becomes relatively more important as the drainable porosity becomes very small.

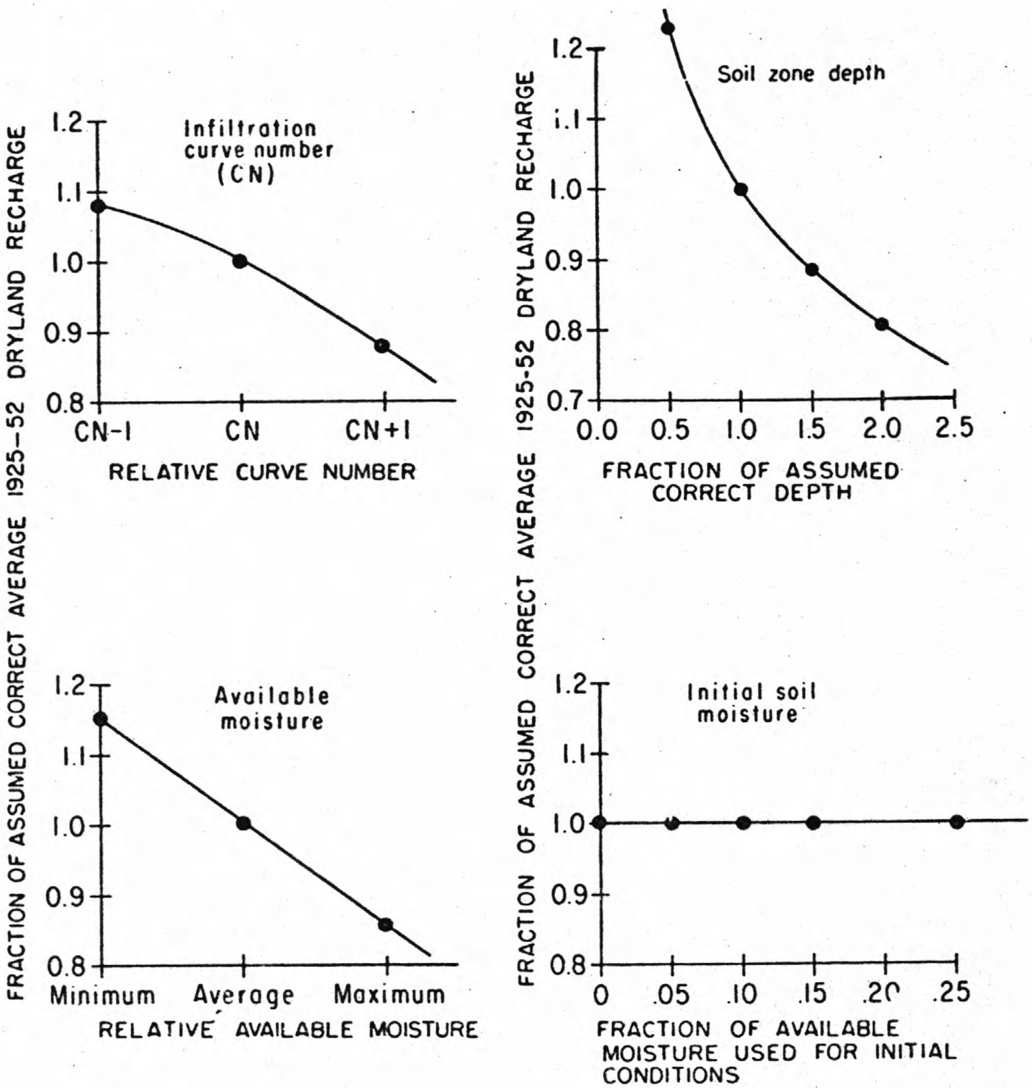


Figure 36.--Sensitivity of computed net recharge to soil-zone depth, infiltration curve number, available moisture capacity, and initial soil-moisture conditions.

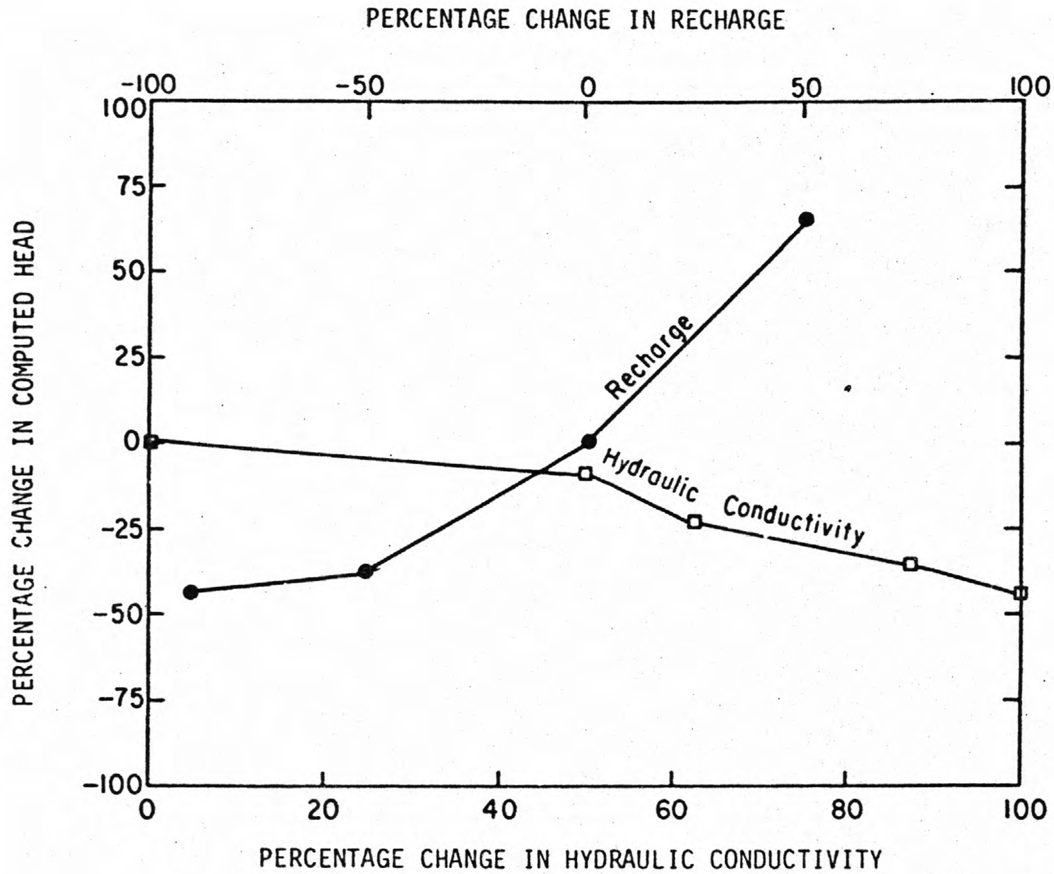


Figure 37.--Sensitivity of computed hydraulic head to recharge and to hydraulic conductivity.

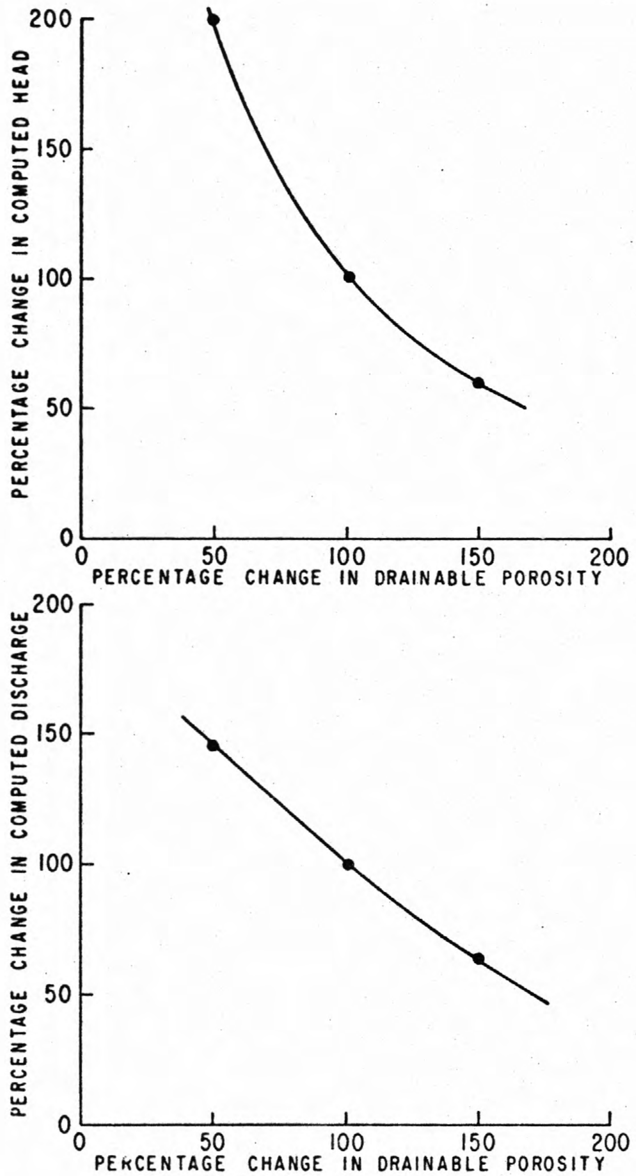


Figure 38.--Sensitivity of computed hydraulic head and computed aquifer discharge to drainable porosity.

Sensitivity analyses of aquifer discharge to hydraulic conductivity were made by using a direct solution to equation 22c. The sensitivity function used was the sum of the squared differences between observed and computed steady-state base flow of streams at ten gaging stations in the study area. The observed values were taken to be the average October, November, and December flows over the period of record prior to 1952. Flows during these months are least affected by irrigation, evapotranspiration, and ice. Hydraulic conductivity was varied using the same method given in equation 28.

In addition to testing the sensitivity to the distributed values of $\hat{K}$, sensitivity of aquifer discharge to a single value of hydraulic conductivity over the entire study area was analyzed. Figure 39 shows the results of both of these sensitivity analyses.

Sensitivity of aquifer discharge to drainable porosity was determined by solving the transient form of the flow equation and computing changes in net flux to streams and lakes for different values of $\hat{\varepsilon}_D$. The results of varying $\hat{\varepsilon}_D$ about the same values used to compute sensitivity of hydraulic head are shown in figure 38.

Table 12 summarizes the results of the sensitivity analyses made for this study.

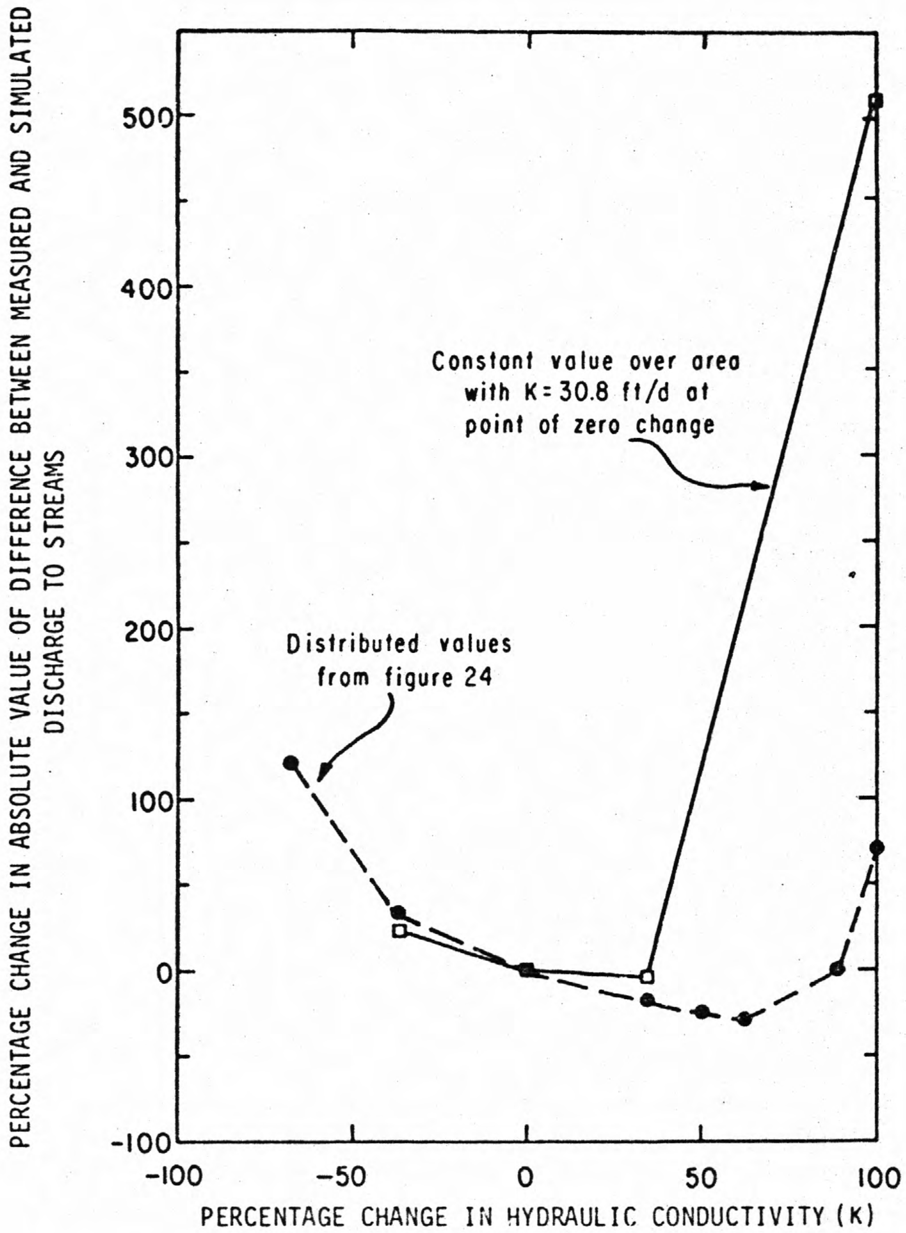


Figure 39.--Sensitivity of aquifer discharge to streams to hydraulic conductivity.

Table 12.--Sensitivity of computed hydrologic response to expected values of model parameters

Model	Measure of hydrologic response	Model parameters ranked in order of sensitivity: Most			Least
Soil zone	Weighted average net recharge, 1925-52	Soil depth	Available moisture	Infiltration curve number	Initial soil moisture
Saturated zone	Hydraulic head	Recharge or discharge	Drainable[1/] porosity		Hydraulic conductivity
Saturated zone	Discharge to streams	Hydraulic conductivity			Drainable[1/] porosity

[1/] Within the range of values representative of unconfined conditions ($\hat{\varepsilon}_D \geq 0.01$).

Steady-state testing of model parameters

The second step in model testing used for this study consisted of steady-state and transient simulation using the assumed correct values of the model parameters in equations 20 and 22. The computed and observed values of hydraulic head and net flux were then compared. Differences between observed and computed values were minimized in some sense by trial and error adjustment of values of the model parameters. The results of the sensitivity analyses were used extensively in this process.

Models of both the soil and saturated zones were tested together. This was necessary because no measured values of soil moisture and deep percolation were available to test the soil-zone model. The net flux computed with the soil-zone model was used as input to the saturated-zone model and resultant computed hydraulic head in and aquifer discharge to streams from the saturated zone were used to test the soil-zone parameters.

Hydraulic conductivity.--Hydraulic-conductivity values were tested by comparing computed steady-state base-flow values with seepage runs made in 1952. Steady-state base flows are computed using a direct solution to equation 22c using 1952 head values. The following lists the hydraulic-conductivity distributions tested in order of producing the best fit to aquifer discharge.

1. Hydraulic conductivity determined from the sample log description method (fig. 24). Some recontouring of the field data was done to prepare figure 24. This process was guided by the distribution of high and low $\hat{k}$ values determined from specific capacity measurements (fig. 23). Hydraulic-conductivity values near perennial streams were changed more than others. However, recontouring necessarily considered much larger areas. Most reinterpretation was done in the eastern parts of Chase County and southern Dundy County.

2. Constant value of $\hat{k}$ equal to the geometric mean of values determined from test holes (30.8 ft/d).

3. Constant value of $\hat{k}$ equal to the geometric mean of values determined from specific capacity (61.5 ft/d).

4. Distributed values equal to one-half of values determined from the specific capacity method.

5. Distributed values determined from the specific capacity method (fig. 23).

The poor ranking of the $\hat{k}$ distribution determined from reported specific capacities probably is due to insufficient and inconsistent durations of pumping prior to measurement of pumping water levels as well as the relatively simplistic relation used between specific capacity and transmissivity (equation 9).

Base flows computed using the best $\hat{k}$ distribution are compared with observed values in table 13. The comparison using seepage measurements in 1952 is shown in plate 2. Although no seepage measurements were made on Rock and Indian Creeks during 1952, conditions measured in 1975 are considered fairly representative of 1952 (preirrigation development) conditions.

The resultant "best" hydraulic-conductivity distribution is not unique and may be considerably in error at some places in the study areas. Errors should be at a minimum along perennial streams.

Table 13.--Measured and modeled base flow in 1952 and 1975 at selected sites

Station	Observed flow, fall 1952 (ft^3/s)	Average Oct.-Dec. flows through 1967 (ft^3/s)	Computed flow using eq. 22c (ft^3/s)	Computed flow using computed steady-state heads (ft^3/s)	Observed flow spring 1975 (ft^3/s)	Average Oct.-Dec. flows, 1973-74 (ft^3/s)	Computed flow, spring 1975 (ft^3/s)
South Platte River between Julesburg, Colo., and Paxton, Nebr. (gain only)	....	10	11	14	....	....	4
Frenchman Creek below Champion	a/ 45	42	44	30	36	....	21
Frenchman Creek nr. Imperial (Enders Reservoir inflow)	a/ 66	73	68	46	53	50	51
Stinking Water Creek near Wauneta (includes Spring Creek in Chase County)	a/ 21	21	27	21	19	....	15
Indian Creek near Max	....	....	4	5	4	....	4
Spring Creek nr. Benkelman	....	1	2	1	.6	....	2
Horse Creek near Parks	....	2	4	2	.1	....	.3
Rock Creek near Parks	b/ 15	13	16	12	14	14	10
Buffalo Creek near Parks	b/ 11	9	5	7	9	8	2

a/ Seepage runs.

b/ Average daily flow, October.

Recharge from the soil zone.--Steady-state recharge to the Ogallala aquifer is the sum of recharge from deep percolation and possible inflow from the Chadron sand aquifer. Flux from the Chadron sand aquifer is unknown over the subcrop area shown in figure 17. Based upon measured head differences at one point in Perkins County (fig. 22), the flux is negligible under nonpumping conditions at this location. Consequently, the flux to the Ogallala from the Chadron sand under steady-state conditions is considered negligible. Average values of recharge for the 1925-52 period were computed using the soil-zone model. Since steady-state recharge is the most sensitive to soil-zone depth, recharge was varied by changing the soil-zone depth. This has the effect of changing the storage capacity of the soil zone.

Crop distributions used in computing steady-state recharge were those shown on plate 5. Dryland conditions were considered to prevail over the entire study area prior to 1952. Irrigation wells present prior to 1952 were scattered enough that any additional recharge from return flow of pumped ground water was considered negligible.

Net recharge was computed using monthly values of precipitation and evapotranspiration between 1925 and 1952, combining these into seasonal values (June-August and September-May), and averaging the seasonal values over the 1925-52 period and on the seven weather stations used (table 7).

Hydraulic heads computed with these net recharge values are compared to the 1952 water-table in figure 40. Conditions in 1952 are assumed to represent at least quasi-steady-state conditions. This assumption is valid over the southern two-thirds of the study area where no long-term measured change in water levels occurred prior to 1952. Hydrologic conditions in the northern one-third of the study area may not be representative of steady-state conditions as significant long-term changes were occurring in 1952 as indicated by the hydrograph of well 11N-39W-35DD shown on plate 8.

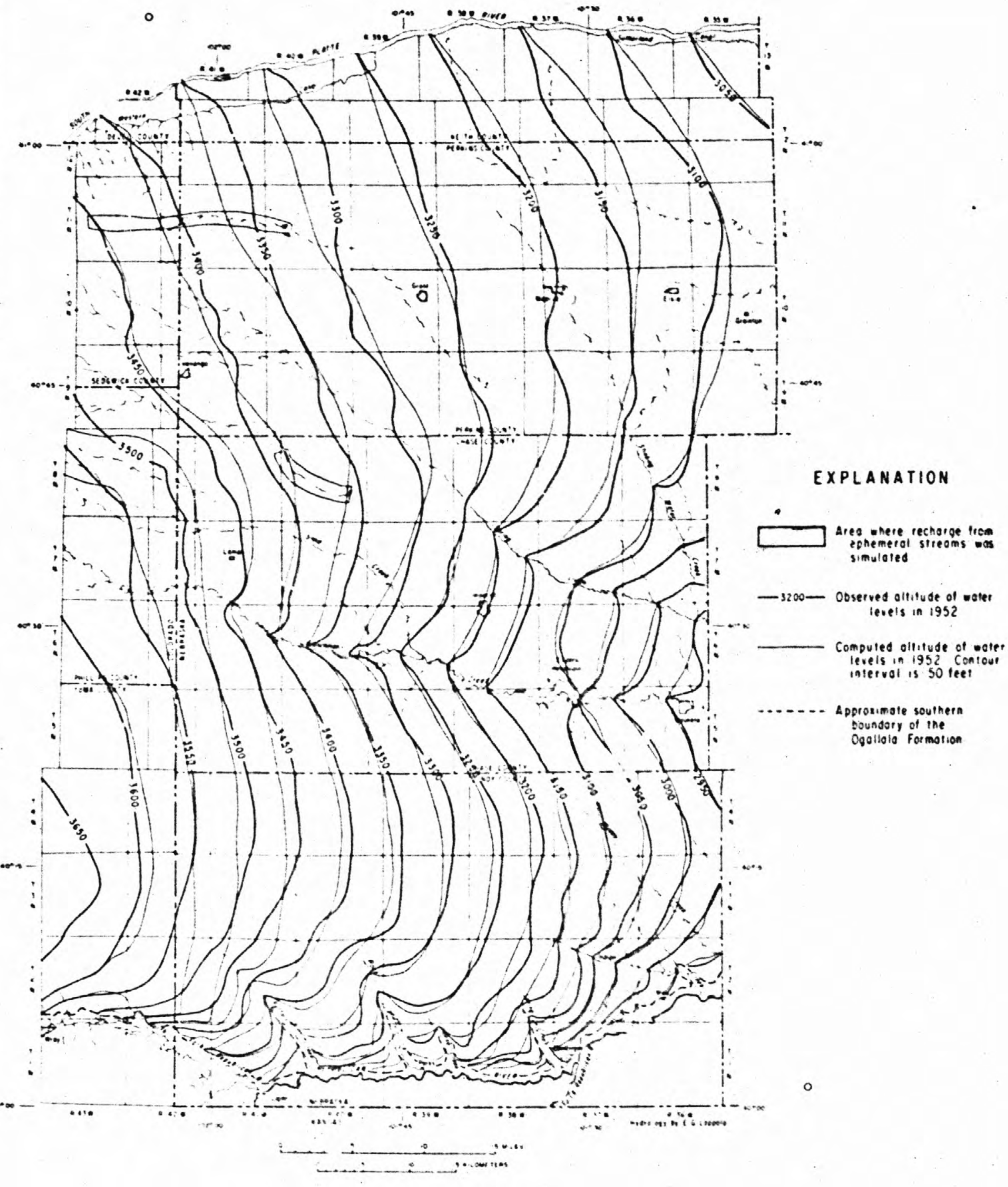


Figure 40.--Comparison between the computed steady-state hydraulic-head configuration assumed to represent preirrigation (1952) conditions.

Table 14 ranks soil depths used in order of producing the best fit to the 1952 water table. The depths producing the best fit may or may not be representative of actual field conditions during a period of time. They represent the depth required to produce the average recharge shown in table 7 when monthly time increments are used to solve equation 20. The solution to equation 20 used in this study assumes that ET demands are satisfied before deep percolation. The situation may be reversed in sandy soils because of their high hydraulic conductivities even at relatively low moisture contents. Use of a small Δz for sandy soils in the solution to equation 20 crudely simulates this effect.

The shallow modeled soil depths required to produce adequate recharge may also represent relatively low plant populations found under pasture in sandy soils. The crop coefficients used (fig. 31) assume an effectively continuous vegetative cover. Field observations show this condition to be approached in the interdune valleys, but not in the dune areas. This effect has been observed elsewhere in sand-dune areas of Nebraska (Keech and Bentall, 1971).

Table 14.--Soil depths (ΔZ) used to compute average recharge rates, 1925-52

Soil-topographic complex	Effective soil depth (ΔZ) (inches)				
	Best fit				Worst fit
Silt-loam, flat, upland	60	..	..	30	15
Silt-loam, rolling to hilly	60	..	..	30	15
Sandy loam, flat	60	..	..	30	15
Silt-loam, flat, bottom land	60	..	..	30	15
Sandy loam, rolling	30	..	..	30	60
Sand and dune sand, hilly	12	..	..	30	60
Loamy sand, flat, interdune valleys	12	15	20	30	60
Rough broken land	60	..		30	15
Loam, rolling	48	60		30	15

Areal distribution of recharge rates shown in table 7 was also evaluated during this step. Recharge was applied: (1) only where recharge was indicated by solutions for net flux obtained by direct solution to equation 22c and (2) over the entire model. The former case resulted in computed steady-state heads that were much lower than 1952 conditions. Applying recharge over the entire area resulted in the best fit between computed steady-state and measured 1952 water levels. The recharge rate over the entire model resulting from this analysis was 117,000 acre-ft/yr. This value is reasonably close to the 126,800 acre-ft/yr total estimated average discharge of ground water to streams prior to 1952 for the entire study area.

The value of total recharge used to compute the steady-state water table includes simulated seepage from the ephemeral reaches of three streams in the western part of Chase and Perkins Counties. These reaches are shown in figure 40 and were included to simulate the recharge from infrequent surface runoff in these areas. Recharge from seepage was linearly increased from 0.05 $(ft^3/s)/mi$ at the upstream ends of the reaches to 0.10 $(ft^3/s)/mi$ at their lower ends for the southern two reaches and twice these rates for the reach in Perkins County. These arbitrary rates are considered physically reasonable for runoff from these areas.

Solutions for steady-state head computed in this step of testing were also used to refine hydraulic-conductivity distributions after acceptable values of effective soil depth were determined. These changes were accomplished by recontouring the point values shown on figure 24.

Base flow computed using the computed steady-state heads resulting from the iterative solution to equation 22 was also compared to measured flows and those computed using the measured 1952 water-level distribution (table 13).

The total recharge rate computed using the direct solution to equation 22c is 235,000 acre-ft/yr. The significant discrepency between this value and those derived from the soil model of 117,000 acre-ft/yr and basinwide water balance methods of 126,800 acre-ft/yr can be attributed to the following, ranked in order of likelihood:

1. Errors inherent in the solution method used for equation 22c and the arbitrary smoothing method used.

2. Errors in the initial head used for the solution to equation 22c (including the possibility that these heads do not represent steady-state conditions).

3. Sources of recharge other than those used to compute the steady-state heads.

4. Discharge of ground water not accounted for in the basinwide water balance. For example, water may be discharging into the Chadron sand aquifer in some areas.

Since the same hydraulic-conductivity distribution was used for the solutions to both equations 22 and 22c, errors in this parameter are not possible causes for the discrepency. Aquifer discharge to streams computed from steady-state head solutions (table 17) agrees fairly closely with solutions to equation 22c and observed discharges. The low heads and low modeled discharge to Frenchman Creek computed from those heads are attributable to low simulated recharge in the area underlain by sandy soils in the ground-water basin of Frenchman Creek. Thus, solutions to equation 22c are reasonably error free in known areas of ground-water discharge to streams where hydraulic head values remain constant or fluctuate considerably less than in recharge areas. Values for net flux computed elsewhere are more susceptible to error in input heads, transient conditions, and numerical approximation.

It is concluded that the solution to equation 22c gives reasonable values of aquifer discharge in areas where the hydraulic head is relatively constant, such as along hydraulically connected streams and lakes not subject to stage fluctuations of more than a few tenths of feet. Absolute values of net flux values computed elsewhere may be considerably in error because true steady-state conditions do not exist due to seasonal fluctuations in hydraulic head of 4 to 5 ft. The areal distribution of recharge and discharge areas (fig. 25) computed with equation 22c is probably reasonable. However, due to aforementioned conditions, absolute values of net flux shown on this figure may be too large in recharge areas. Some discharge areas shown in figure 25 which are not associated with known areas of discharge may be anomalous for the above reasons. Net flux distributions determined from solutions to equation 22c should be used with caution as the only supporting evidence in identifying the magnitude of ground-water discharge in the absence of other lines of evidence.

Transient testing of model parameters

The remaining model parameters to be tested were the drainable porosity (ϵ_D) and the net withdrawal (W). Hydrologic response of the saturated zone used in transient testing was measured by (1) changes in water levels over the entire study area from 1952 to 1975 (fig. 16); (2) changes in water levels for the period of record between 1952 and 1975 at selected observation wells (pl. 8); (3) changes in base flow of streams as measured by seepage runs made in 1952 and in 1975 (pl. 2); and (4) changes in base flow at selected gaging stations between 1952 and 1975 (table 13).

Net withdrawal.--Net withdrawal used in transient testing was computed using equation 25. Values of Q_D and Q_I in this equation were computed using the model of the soil zone. This model and its parameters determined during steady-state tests were assumed to be correct for transient analyses. Parameters used to compute Q_I that were not evaluated in the steady-state testing were the effective soil depth under irrigated conditions and the value of 50 percent of available moisture used as a minimum allowed before a required irrigation. The minimum allowable soil moisture was not changed during transient testing. This is a reasonable value required to keep plant stress at a minimum (D. Watts, University of Nebraska, oral commun., 1975).

Consumptive-irrigation requirements were computed on a monthly basis and summarized into seasonal values for use in transient tests. The irrigation season simulated was June-August and the nonirrigation season was September-May. The net withdrawal or recharge applied at each finite-difference cell was the difference between average 1925-52 recharge under dryland conditions and the seasonal consumptive-use requirement under irrigated conditions. Averages of these values for the 1952-75 transient test period are shown in table 15. Also shown in this table is the variability of consumptive-irrigation requirements using different values of effective soil depth. The values of effective depth determined in the steady-state tests were also used for transient analyses.

Table 15.--Average June through August consumptive-irrigation requirements, 1952-75

a: By soils

Explanation: Soil depth: A = 60 inches for all soils. B = 30 inches for all soils. C = 60 inches for soils 1, 2, 3, 4, and 8; 30 inches for soil 5; 12 inches for soils 6 and 7; 48 inches for soil 9. All weather stations used; all crops except small grains used.

Index No.	Soil type	Soil depth used (inches)	Consumptive-irrigation requirement (in./season)
1	Silt-loam, flat, upland	A	18.3
		B	18.3
		C	18.3
2	Silt-loam, rolling to hilly	A	20.0
		B	20.0
		C	20.0
3	Sandy loam, flat	A	18.3
		B	18.3
		C	18.3
4	Silt-loam, flat bottom land	A	18.6
		B	18.6
		C	18.6
5	Sandy loam, rolling	A	18.5
		B	18.7
		C	18.7
6	Sand and dune sand, hilly	A	18.4
		B	18.7
		C	18.9
7	Loamy sand, flat, interdune valleys	A	18.5
		B	18.8
		C	19.1
8	Rough broken land	A	19.9
		B	19.9
		C	19.9
9	Loam, rolling	A	19.2
		B	19.4
		C	19.4

Table 15.--Average June through August consumptive-irrigation requirements, 1952-75--continued

b: By weather stations

Explanation: Soil depth: A = 60 inches for all soils. B = 30 inches for all soils. C = 60 inches for soils 1, 2, 3, 4, and 8; 30 inches for soil 5; 12 inches for soils 6 and 7; 48 inches for soil 9. All soils used; all crops except small grains used.

Weather station	Soil depth used (inches)	Consumptive-irrigation requirement (in./season)
	A	18.8
Benkelman	B	19.0
	C	19.0
	A	17.4
Imperial	B	17.6
	C	17.6
	A	18.1
Madrid	B	18.3
	C	18.3
	A	16.8
Ogallala	B	17.0
	C	17.0
	A	20.0
Holyoke, Colo.	B	20.2
	C	20.2
	A	19.8
Wray, Colo.	B	20.0
	C	20.0
	A	20.8
Julesburg, Colo.	B	20.9
	C	20.9

Table 15.--Average June through August consumptive-irrigation requirements, 1952-75--continued

c: By crops

Explanation: Soil depth: A = 60 inches for all soils. B = 30 inches for all soils. C = 60 inches for soils 1, 2, 3, 4, and 8; 30 inches for soil 5; 12 inches for soils 6 and 7; 48 inches for soil 9. All soils used; all weather stations used.

Crop	Soil depth used (inches)	Consumptive-irrigation requirement (in./season)
Row crop	A	13.7
	B	14.0
	C	14.0
Alfalfa	A	22.2
	B	22.3
	C	22.3
Small grains	A	7.7
	B	8.6
	C	8.2
Pastureland	A	20.6
	B	20.7
	C	20.7

Simulated net stress on the aquifer was modified in the extreme northwestern part of the study area to account for lands irrigated from the Western Canal (pl. 7). In this area, the consumptive-irrigation requirement was simulated as a combination of surface water and ground water. Surface-water availability was determined annually by dividing the total measured diversions of the Western Canal by the acreage reported by the Nebraska Department of Water Resources (annual) to be irrigated. Any consumptive-irrigation requirements in excess of these unit amounts were simulated as being derived from ground water. The canal itself was modeled as nonleaky.

Storage properties of the aquifer.--Initial estimates of areal distribution of storage properties of the aquifer were modified during this step of model testing. Minor changes were made by recontouring the data shown on figure 21, principally in the northwest corner of the study area. Modifications were made to the original values used for the storage properties to account for semiartesian conditions in the north-central part of Perkins County and the south-central part of Keith County. As discussed previously, analysis of measured water-level responses in a recorder well in this area indicated a storage coefficient of 0.002. This value was used throughout the area shown in figure 21. Agreement between simulated and observed water-level response for areas of water-level rise was apparent only in the western part of Perkins County. The remaining water-level rise is not adequately simulated. This disparity is attributed to inadequate simulation of increased recharge in the eastern part of the area of rise.

Field documentation of the extent and storage properties of the semiartesian area in Perkins and Keith Counties is limited to 1 year of water-level measurements in recorder well 12N-38W-31CCCC together with reported well interference problems. These conditions are assumed to exist for the predictive analyses discussed subsequently. The adequacy of the model in predicting water levels in this area should be established by field documentation of the assumed semiartesian conditions.

Because of the uncertainty in the areal extent of the semiconfined area and in the unknown vertical variation of these conditions as heads decline, a transient analysis was also made for the 1952-75 period using drainable porosity values rather than the semiartesian storage coefficient of 0.002 in this area. Simulated water-level changes from 1952 to spring 1975 for both unconfined and semiartesian conditions are shown in figure 41.

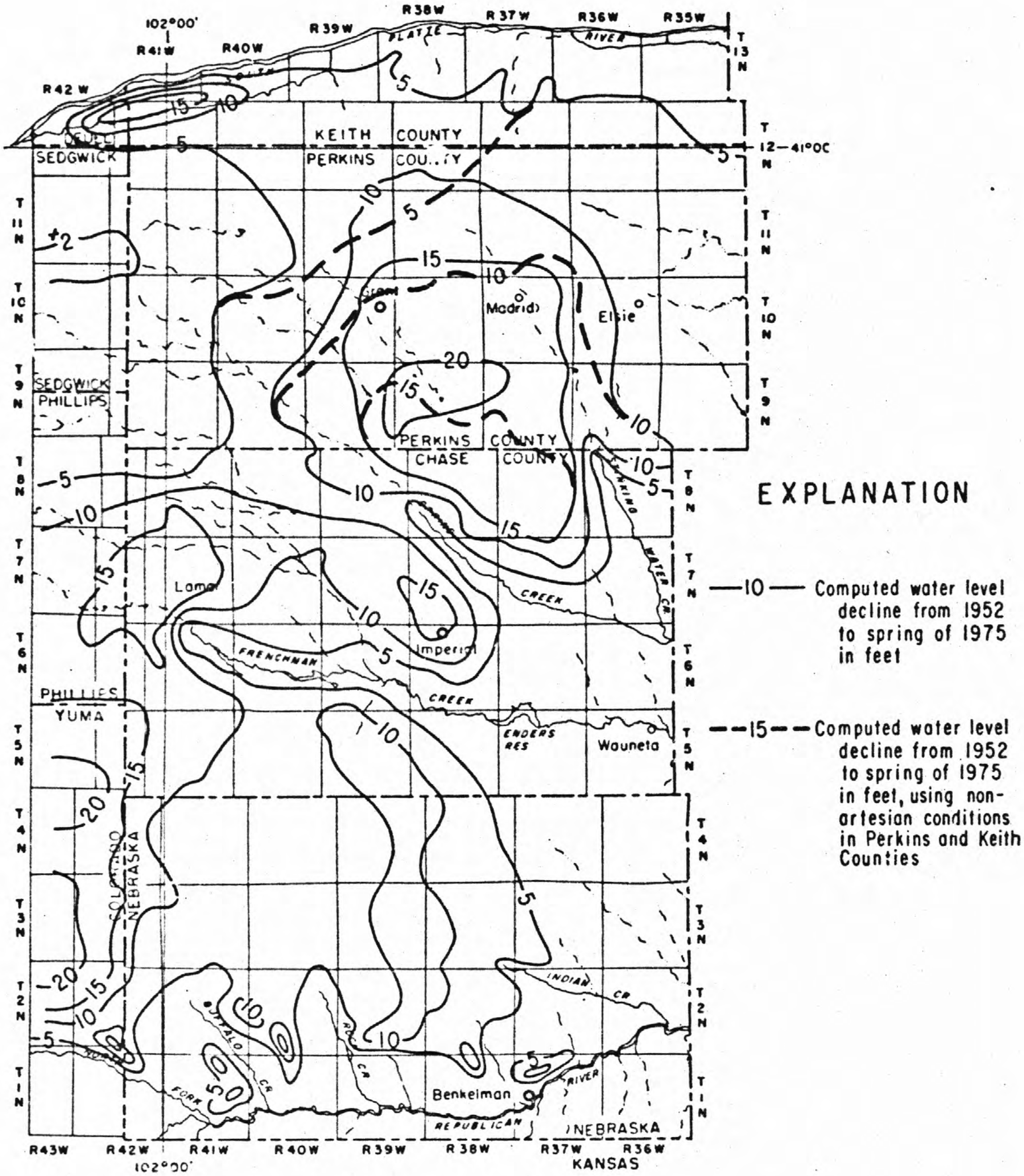


Figure 41.--Modeled change in water levels from January 1952 to May 1975.

For the apparent semiartesian area around Lamar (fig. 21), transient testing indicated that a value of the storage coefficient representative of semiartesian conditions (0.001) was too small. Simulated seasonal water-level fluctuations were of the same order of magnitude as those in the Lamar recorder well. However, the computed areal water-level declines from 1952 to 1975 in general were twice too large. The value of drainable porosity shown on figure 21 gave the best fit to these changes and was used for predictive analyses.

The parameters in models of the soil and saturated zones that resulted from model testing are considered reasonable in the context of available field data. They by no means represent a unique set of values. The saturated-zone model is the most representative of field conditions in Chase and Dundy Counties due to the availability of historical water levels used to test the model. The representativeness of the model in Perkins and southern Keith Counties is somewhat less due to lack of adequate field data to define the areal and stratigraphic extent of the semiartesian area and the prevailing storage properties of the aquifer in this area.

The model of the soil zone is considered adequate for evaluating long-term recharge rates and yearly consumptive-irrigation requirements. Evaluation of changes in recharge and irrigation requirements due to changing crop distributions and climatic conditions on less than a seasonal (3-month) basis are less accurate due to assumptions made in constructing the model.

Most of the discrepancy between simulated and observed water levels can be attributed to inadequate definition of net withdrawal or recharge in these areas. Due to the method used to generate initial water levels at the start of a simulation, the model of the saturated zone can adequately evaluate changes from a given condition, even if the absolute values of hydraulic heads are in error.

PREDICTIVE ANALYSES

The tested models of the soil and saturated zones are considered acceptable representations of the hydrogeologic system in the study area within the limitations of some data inadequacies as discussed previously. These models can be used to predict the effect of combinations of inputs and outputs on water levels and streamflow. However, this report only describes two such predictive analyses which cover the period 1976 to 2000. The first analysis was the null management alternative which assumes no regulation of consumptively used ground-water withdrawals from existing or future wells and is the baseline against which the effects of proposed management schemes may be evaluated. In addition to the null alternative, one analysis was made to evaluate the effects of cessation of new irrigation well installation after 1976. Both analyses are described below.

It is important to note that the effects of withdrawals from the aquifer that are consumptively used are determined in these analyses. Total pumpage that can be directly measured may or may not be equivalent to the consumptive use of ground water.

Due to the uncertainty over the lateral extent and storage properties of the semiartesian area in north-central Perkins and south-central Keith Counties, predictive analyses have been made using both unconfined and semiartesian storage properties in this area. As discussed subsequently, the significantly different water-level declines under both conditions strongly suggest that the extent and storage properties of this area be defined by further field observations.

For evaluation of further analyses, the following statistics are provided on computer requirements for the two analyses that follow. All runs were made on the U.S. Geological Survey IBM 370/155 Computer.

Period of simulation: 25 years

Number of time steps required: 82

Central Processor Unit time required: 91.5 minutes

Core storage required: 400 K bytes

The Null Alternative

Analysis of any combination of inputs and outputs to the saturated zone model requires specification of their timing, location, and rates. For the null alternative, withdrawals were assumed to occur over the same 3-month (June-August) period used in model testing. For the past 5 years about 75 irrigation wells have been registered each year in each county of the District (fig. 27). To predict the effect of the continuation of this rate of development, the study area was considered to be represented by three subareas divided along the Chase-Perkins and Chase-Dundy County lines. In each of the three subareas 75 wells per year were added, subject to constraints described below.

The locations of new irrigation wells were determined for each year by randomly locating wells in finite-difference cells containing non-irrigated irrigable land until either 75 wells had been located in each of the three subareas, or the total irrigable acreage in each area was reached. For these analyses, land was considered irrigable unless it was included in the rough and broken category shown on plate 1. Unless more wells were present in 1975, a maximum of five wells was permitted in each finite-difference cell. The area irrigated with each well was set at 133 acres, which is the average area covered with a center-pivot irrigation system, with the total acreage in each cell constrained to be less than 560 acres. The maximum additional irrigated area in each of the three subareas used as an ultimate limit to well installation was computed as the difference between land irrigated in 1975 and the total irrigable land. Table 16 summarizes the additional irrigable land and the potential number of wells.

Table 16.--Remaining irrigable land and equivalent number of wells[1/]

County	Total irrigable land (acres)	Irrigated land 1975 (acres)	Remaining irrigable land (acres)	Remaining number of wells
Chase	468,160	112,034	356,126	2,678
Dundy	384,720	67,244	317,476	2,387
Perkins and Keith	684,320	99,271	585,049	4,399

[1/] Assuming 133 acres per well.

Some justification for the random location of new wells on irrigable land is warranted. A comparison for Dundy County of the percentage distribution of irrigation suitability classes (as defined by the Soil Conservation Service) occupied by center-pivot systems with the county-wide distribution of these classes shows no significant difference as determined by R. O. Hoffman, University of Nebraska (written commun., 1975). This comparison shows that irrigation suitability was not a significant factor in location of wells used to supply center pivots over the 1973-75 period.

The amount of net withdrawal of ground water used in the null alternative analysis was set at a constant value of 16 in. This value is close to the average consumptive-irrigation requirement weighted for 82 percent row crops and 18 percent alfalfa for the 1952-75 period. A constant value was used to provide a more constant base against which to compare other net withdrawal schemes.

Predicted water-level changes produced by the null alternative are summarized at the end of 5-year periods on plate 9 and predicted base flows of selected streams are shown in table 17.

Table 17.--Predicted base flows at selected locations under conditions of: (A) No irrigation-well installation after 1976 and (B) Continued irrigation-well installation at the 1970-75 rate (All wells simulated as withdrawing 177 acre-feet of consumptively used water during the June through August period each year)

Stream and location of tabulated base flows	Base flow as of May 30 each year (ft^3/s)									
	1980		1985		1990		1995		2000	
	A	B	A	B	A	B	A	B	A	B
South Platte R., reach gain from south side, Julesburg, Colo., to Keith County, Nebr., line	2	2	1	0	0	0	0	0	0	0
Stinking Water Cr. plus Spring Cr. at confluence near Chase-Hayes County line, NE¼ sec. 13, T. 6 N., R. 36 W.	5	5	1	1	1	0	1	0	1	0
Frenchman Cr. above Champion, NW¼ Sec. 19, T. 6 N., R. 39 W.	12	12	6	3	1	0	0	0	0	0
Frenchman Cr. below Champion, SW¼ sec. 22, T. 6 N., R. 39 W.	18	18	11	7	5	1	2	0	0	0
Frenchman Cr. near Imperial (Enders Reservoir inflow), NE¼ sec. 3, T. 5 N., R. 38 W.	30	30	19	14	10	2	5	0	3	0
Indian Creek near Max, SE¼ sec. 16, T. 2 N., R. 36 W.	4	4	3	3	3	2	2	2	2	2
Buffalo Creek near Haigler, SW¼ sec. 26, T. 2 N., R. 41 W.	2	2	1	1	1	1	1	0	1	0
Rock Cr. above State recreation area, NW¼ sec. 31, T. 2 N., R. 39 W.	10	10	10	9	9	7	8	5	8	3
Horse Creek near Parks, SE¼ sec. 2, T. 1 N., R. 39 W.	1	1	1	1	0	0	0	0	0	0
Spring Creek near Benkelman, SE¼ sec. 2, T. 1 N., R. 38 W.	1	1	1	1	1	0	1	0	0	0

Water-level declines shown on plate 9 represent average declines over each finite-difference cell. Additional declines in individual pumping irrigation wells will be lower.

Ground water in storage would be reduced by 3.7 percent by 2000 under this simulated alternative. Saturated thickness of the Ogallala aquifer would be less than 50 ft over 16 percent of the study area compared to 6 percent in 1975. By 2000, under this alternative, 80.3 percent of the net withdrawals would be derived from ground-water storage and 19.7 percent from stream depletion.

The simulated stream depletions shown in table 17 indicate that surface-water supply of Frenchman Creek could be essentially eliminated by 1992. Further support of this prediction can be found by extrapolating the trend since 1969 of the hydrograph of Frenchman Creek near Imperial (pl. 1). Extrapolations of both the annual peaks and troughs of the hydrograph intersect the zero axis about 1992. This agreement with the simulated effect of the null alternative is not coincidental. If development were to cease, stream depletion would eventually level off as a new equilibrium between ground-water discharge to wells and streamflow was reached. However, continued development at the rate of the last few years has not and will not allow a new equilibrium to be reached before the streamflow at this point ceases. The total annual undepleted flow of Frenchman Creek (prior to 1967) is equivalent to the annual consumptive use from about 290 wells using the same acreage and unit consumptive demands used for the null alternative. Irrigated and irrigable lands within 2 mi of Frenchman Creek above Enders Reservoir are sufficient to emplace this number of wells with four wells to a section.

No Future Development

The second predictive analysis made with the model simulated the effects of the cessation of well installation after 1976. The same seasonal consumptive use or net withdrawal used for the null alternative was used for this analysis. The simulated water-level changes under this alternative are shown on plate 10, and stream depletions are shown in table 17. This alternative is perhaps unrealistic, but the simulated results show reduced rates of water-level declines and stream depletions. It is important to note, however, that within the time frame simulated, no new equilibrium was reached. Streamflow continued to be depleted and water levels continued to decline. Local equilibrium could be reached where wells are widely scattered. Ground water in storage would be reduced by 2.8 percent by 2000 under this simulated alternative. Saturated thickness of the Ogallala aquifer would be less than 50 ft over 10 percent of the study area compared to 6 percent in 1975. By 2000, under this alternative, 87.4 percent of the net with-drawals would be derived from ground-water storage and 12.6 percent from stream depletion.

Effect of Changing the Consumptive-Irrigation Requirement

A value of 16 inches over the June through August irrigation season was used as the consumptive-irrigation requirement (CIR) for both predictive analyses. This figure is reasonably close to the weighted average for the 1975 distribution of irrigated crops. However, it may be of interest to determine the water-level declines and stream depletions for the two predictive analyses if the CIR were different from 16 inches. These can be simply determined without further computer runs as outlined below.

These effects can be reasonably well approximated by multiplying the water-level decline or stream depletion at a given location at a given point in time by the ratio between the desired CIR and 16 inches. For example, the drawdown from 1975 to 1995 under conditions of no future development in the Imperial area was about 40 ft using a CIR of 16 inches. If a CIR of 12 inches were applied everywhere instead, the approximate drawdown would be (12/16) * 40 = 30 ft. This method of determining the effects of varying CIR values is applicable with the following restrictions:

1. The same CIR value is applied over the entire model, or at least over areas considerably larger than the point of interest.

2. The same temporal and spatial distribution of wells is used as was used for either of the two predictive analyses made using a CIR of 16 inches.

3. The saturated thickness is sufficiently large in relation to the drawdown at the time of interest that the principle of linear superposition is applicable.

Streamflow depletions due to diversion by wells of natural ground-water discharge under both tested futures will reduce availability of water to historical users of surface water. Provisional surface-water grants used to irrigate lands within the study area are listed in table 18. Downstream prior rights to the waters of Frenchman Creek dating back to 1890 are held by irrigation districts in the McCook area. These districts also hold most of the water storage rights in Enders Reservoir. If all the provisional grants from Frenchman Creek above Enders Reservoir were exercised only to irrigate lands within the District, they would about equal the average 1968-75 flow at the gaging station near Imperial (table 13).

Table 18.--Surface-water provisional grants for lands within the Upper Republican NRD

Source of water	Total provisional grants, in ft^3/s		
	Canal	Pump	Total
Frenchman Creek above Enders Reservoir	40.66	15.75	56.41
Frenchman Creek, Enders Reservoir to Chase County line	0	13.38	13.38
Spring and Stinking Water Creeks	11.08	18.34	29.42
Horse Creek	3.72	0	3.72
Indian Creek	0	11.73	11.73
Rock Creek	4.44	5.21	9.65
Buffalo Creek	10.57	6.86	17.43
Spring Creek (Dundy County)	.14	2.37	2.51
Muddy Creek	0	2.42	2.42
Republican River	85.11	10.62	95.73
South Platte River	188.29	9.44	1/ 197.73

1/ May include canals and pumps irrigating small tracts north of the Platte River.

SUMMARY

This study quantifies the hydrogeologic system of the area including the Upper Republican Natural Resources District and surrounding land in southwest Nebraska and northeast Colorado.

Mathematical models of the soil and saturated zones were developed in this study to enable quantification of the hydrogeologic system. The response of the models was tested against measured states of the hydrogeologic system under both preirrigation-development (1952) and present irrigation-development conditions. Differences between measured and computed hydrologic response were reduced to an acceptable level by reinterpretation of field measurements, model parameters, and modeled boundary conditions. The accepted models of the soil and saturated zones reproduced water levels and ground-water discharge to streams that were generally within 10 percent of measured values.

The largest possible error in analyses made with the models is the treatment of the semiconfined area in Perkins and Keith Counties. Little measured field documentation of the lateral and vertical extent of these conditions exists. The boundaries for this area used in testing were defined using drillers' reports of well interference problems and may be considerably in error. The semiartesian storage coefficient used in model tests in this area was based upon an analysis of water-level response of one recorder well under less than controlled conditions. Irrigation development in most of this semiartesian area has occurred only in the past 2 years (1974-75), and adequate water-level measurements to define the artesian conditions do not exist.

Hydrogeologic Conditions as of 1975

The hydrogeologic system receives ground-water inflow from Colorado, precipitation on the study area, and surface flow of the South Platte, Arikaree, and the North and South Forks of the Republican Rivers. Outflow consists of consumptive use of soil moisture and ground water, streamflow in the South Platte and Republican Rivers and Frenchman and Stinking Water Creeks and ground-water underflow across the eastern project boundary. Under pre-agricultural conditions, inflow was assumed to be balanced by outflow.

Ground water in storage and ground-water discharge to surface streams within the study area comprise most of the exploitable water supplies in the study area. More than 68 million acre-ft of recoverable ground water is presently in storage in the saturated deposits of the Tertiary Ogallala Formation and Holocene alluvial and aeolian deposits. Ground water exists principally under unconfined or water-table conditions. One area of apparent semiartesian conditions is found in north-central Perkins and south-central Keith Counties. Smaller areas of semiartesian conditions exist along the upper reaches of Frenchman Creek in western Chase County and Rock Creek in Dundy County.

Surface runoff as overland flow is negligible and ground-water discharge to streams accounts for most streamflow leaving the study area. The combined ground-water discharge to the nine perennial streams generated as ground-water discharge within the study area was 126,800 acre-ft/yr under conditions representative of minimal withdrawals of ground water until 1967. Ground-water discharge by evapotranspiration from shallow water-table areas in stream valleys was about 14,500 acre-ft/yr prior to 1967 and no detected change has occurred.

Agricultural development has changed the amount of ground water in storage and discharge to perennial streams. Dryland farming, particularly the wheat summer-fallow practice has resulted in increased recharge of precipitation to the ground-water reservoir in the northern one-third of the study area. Recharge under these conditions may be more than 3 in/yr. Under native pasture conditions, in the northern one-third of the study area recharge is less than 0.1 in/yr. The increased recharge caused water-level rises in Perkins and Keith Counties of as much as 13 ft between 1937 and 1967. An estimated 680,000 acre-ft of ground water was added to storage in this area from 1952 to 1975.

Water levels that were rising now are declining over most of the area of rise due to development of ground water for irrigation. Consumptive use of ground water for irrigation has caused water-level declines and stream depletions in the study area. Declines of as much as 19 ft from 1952 to 1975 occurred around Imperial and Lamar. The largest areas of decline are in areas of western Chase and Dundy Counties on both sides of Frenchman Creek. Ground water in storage has been reduced about 1 percent since 1952 with most of the reduction occurring since 1967. Consumptive withdrawal of ground water increased to from less than 30,000 acre-ft/yr in 1952 to an estimated 277,000 acre-ft in 1976 from 2,243 wells within the study area. For the study area as a whole, this consumptive withdrawal exceeds the estimated recharge from precipitation of 166,000 acre-ft/yr.

The study results show that ground-water discharge to perennial streams has been reduced by withdrawals for irrigation. Preirrigation development (pre-1967) flows of Frenchman Creek near Imperial were reduced by as much as 30 percent by 1975. Flows in Stinking Water and Spring Creeks were depleted by less than 5 percent by 1975.

Future Conditions

The accepted model of the saturated zone was used to predict water levels and stream depletions between 1976 and 2000 under conditions of no further development and development continued at the 1970-75 rate. The results of these analyses show that water-level declines of as much as 140 ft could occur by 2000 in the Lamar and Grant areas under conditions of continued development. Water-level declines in most of the remainder of the study area would be less than 60 ft by 2000 under simulated continued development conditions and less than 40 ft under no further development. Ground water in storage by 2000 would be reduced by about 3.7 percent of the recoverable volume in 1975 under the continued development that was simulated and 2.8 percent under conditions of no further development. Saturated thickness of the aquifer would be reduced to less than 50 ft over about 16 percent of the study area by 2000 under conditions of continued development and 10 percent under conditions of no further development compared with 6 percent in 1975.

The predictive analyses further show that the flow of some perennial streams would be essentially eliminated by 2000 if simulated continued development conditions occur. Perennial flow of Frenchman Creek near Imperial and Stinking Water and Spring Creeks at the Chase County line would be eliminated by about 1992 under continued development and would be less than 10 percent of the 1975 flow by 2000 under simulated conditions of no further development. The ground-water discharge that is about 2 percent of the flow of the South Platte River from the south side would be eliminated by 2000. The base flow of the tributaries to the north side of the Republican River would be reduced under both simulated conditions but not eliminated.

Recommendations

As a consequence of the uncertainty over the extent and storage properties of the semiartesian area in Perkins and Keith Counties, predictive analyses were made using both unconfined and confined storage properties of the aquifer. Since the predicted water levels differ by as much as 60 ft, the adequacy of the model as a predictive tool in this area is questionable. To alleviate this problem, it is suggested that several aquifer tests be run in the area to define the semiartesian area and that additional recording wells be installed within the area to monitor the seasonal water levels. Data collected from these studies should be used to further refine the storage properties used in the model.

The models developed in this study are available for continued use as a management tool for the Upper Republican Natural Resources District. The documentation of the mechanics of using these models is being prepared as a separate report. The models should be periodically reevaluated and updated as additional field data become available.

The results of this study show that existing data and analyses herein are not adequate to fully explain the relationship between water-level rises in Perkins and Keith Counties and the practice of summer fallow in this area. Further analyses should be made by attempting to simulate the vegetation change from native range to wheat fallow. Data for this analysis may be impossible to obtain over the entire area where water-level rises have occurred. However, detailed measurement of moisture contents and hydraulic gradients in the unsaturated zone under a few native pasture sites and nearby wheat-fallow sites might quantify the differences in recharge rates.

SELECTED REFERENCES

Arkin, H., and Colton, R. R., 1970, Statistical methods: New York, Burres and Noble, 344 p.

Baver, L. D., 1956, Soil physics: New York, John Wiley, 489 p.

Bear, Jacob, 1972, Dynamics of fluids in porous media: New York, American Elsevier, 764 p.

Bjorklund, L. J., and Brown, R. F., 1957, Geology and ground-water resources of the lower South Platte River Valley between Hardin, Colorado, and Paxton, Nebraska, with a section on Chemical quality of the ground water, by H. A. Swenson: U.S. Geological Survey Water-Supply Paper 1378, 431 p.

Boettcher, A. J., 1966, Ground-water development in the High Plains of Colorado, with a section on Chemical quality of the ground water, by Robert Brennan: U.S. Geological Survey Water-Supply Paper 1819-I, 22 p.

Boettcher, A. J., Hofstra, W. E., and Major, T. J., 1969, Water-level records for the northern High Plains of Colorado: Colorado Water Conservation Board Water Resources Basic-Data Release No. 20, 189 p.

Bradley, Edward, and Johnson, C. R., 1957, Geology and ground-water hydrology of the valleys of the Republican and Frenchman Rivers, Nebraska: U.S. Geological Survey Water-Supply Paper 1360-H, p. 589-713.

Brubacher, J. I., and Moore, T. R., 1971, Soil survey of Sedgwick County, Colorado: U.S. Department of Agriculture, Soil Conservation Service, Soil Survey, 61 p.

Cardwell, W. D. E., 1953, Irrigation-well development in the Kansas River Basin of eastern Colorado: U.S. Geological Survey Circular 295, 72 p.

Cardwell, W. D. E., and Jenkins, E. D., 1963, Ground-water geology and pump irrigation in Frenchman Creek Basin above Palisade, Nebraska, with a section on The chemical quality of the water by E. R. Jochens and R. A. Krieger: U.S. Geological Survey Water-Supply Paper 1577, 472 p.

Condra, G. E., 1907, Geology and water resources of the Republican River Valley and adjacent areas, Nebraska: U.S. Geological Survey Water-Supply Paper 216, 71 p.

Condra, G. E., and Reed, E. C., 1959, The geological section of Nebraska: Nebraska University-Lincoln, Conservation and Survey Division, Nebraska Geological Survey Bulletin 14A, 76 p.

Cooley, R. L., 1974, Finite element solutions for the equations of ground-water flow: Nevada University System, Center for Water Resources Research, Desert Research Institute, Hydrology and Water Resources Technical Report No. 18, 133 p.

Darton, N. H., 1905, Preliminary report on the geology and underground water resources of the central Great Plains: U.S. Geological Survey Professional Paper 32, 433 p.

DeVries, R. N., and Kent, D. C., 1973, Sensitivity of ground-water flow models to vertical variability of aquifer constants: Oklahoma State University, Water Resources Research Institute, Research Project Technical Completion Report, 45 p.

Dixon, W. J. (ed.), 1971, Biomedical computer programs, University of California publications in automatic computation No. 2: Los Angeles, Calif., University Press, 600 p.

Elder, J. A., 1969, Soils of Nebraska: Nebraska University-Lincoln, Conservation and Survey Division, Resource Report No. 2, 59 p.

Eskinazi, S., 1975, Fluid mechanics and thermodynamics of our environment: New York, Academic Press, 422 p.

Frenneman, N. M., 1931, Physiography of western United States: New York, McGraw-Hill, 534 p.

Freeze, R. A., 1975, A stochastic-conceptual analysis of one-dimensional groundwater flow in nonuniform homogeneous media: American Geophysical Union, Water Resources Research, v. 11, no. 5, p. 725-741.

Fritz, S., 1949, Solar radiation during cloudless days: Heating and Ventilating, January 1949.

Hantush, M. S., 1966, Hydraulics of wells in Advances in hydroscience, ed. V. T. Chow: New York, Academic Press, p. 281-432.

Hill, D. R., and Tompkin, J. M., 1953, General and engineering geology of the Wray area, Colorado and Nebraska: U.S. Geological Survey Bulletin 1001, 65 p.

Hofstra, W. E., Klein, J. M., and Major, T. J., 1971, Water-level changes, 1964-71, northern High Plains of Colorado: U.S. Geological Survey open-file report, 10 p.

Hofstra, W. E., and Luckey, R. R., 1973, Water-level records, 1969-73, and hydrogeologic data for the northern High Plains of Colorado: Colorado Water Conservation Board Water Resources Basic-Data Release No. 28, 52 p.

Hofstra, W. E., Major, T. J., and Luckey, R.R., 1972, Hydrogeologic data for the northern High Plains of Colorado: Colorado Water Conservation Board Water Resources Basic-Data Release No. 23, 143p.

Huntoon, P. W., 1974, Predicted water-level declines for alternative groundwater developments in the Upper Big Blue River Basin, Nebraska: Nebraska University, Conservation and Survey Division, Resource Report 6, 106 p.

Hurr, R. T., 1966, A new approach for estimating transmissivity from specific capacity: Water Resources Research, v. 2, no. 4, p. 657-664.

Hurr, R. T., and Luckey, R. R., 1973, Ground-water levels in the South Platte River Valley of Colorado, spring 1973: Colorado Water Conservation Board Water Resources Basic-Data Release No. 30, 33 p.

Hurr, R. T., and Schneider, P. A., Jr., and others, 1972, Hydrogeologic characteristics of the valley-fill aquifer in the Julesburg reach of the South Platte River Valley, Colorado: U.S. Geological Survey open-file report, 2 p., 6 maps.

Hurr, R. T., Schneider, P. A., Jr., and Minges, D. R., 1975, Hydrology of the South Platte River Valley, northeastern Colorado: Colorado Water Conservation Board Water Resources Circular 28, 24 p.

Jenkins, C. T., 1968, Computation of the rate and volume of stream depletion by wells: U.S. Geological Survey Techniques of Water-Resources Investigations, book 4, chap. D-1, 17 p.

Jensen, M. E., Wright, J. C., and Pratt, B. J., 1969, Estimating soil moisture depletion from climate, crop, and soil data: American Society of Agricultural Engineers, paper 69-941.

Johnson, A. I., 1967, Compilation of specific yield for various materials: U.S. Geological Survey Water-Supply Paper 1662-D, 74 p.

Johnson, C. R., 1960, Geology and ground water in the Platte-Republican Rivers watershed and the Little Blue River Basin above Angus, Nebraska, with a section on Chemical quality of the ground water, by Robert Brennan: U.S. Geological Survey Water-Supply Paper 1489, 142 p.

Kastner, W. M., 1974, Application of statistical techniques to the estimation of ground-water withdrawals in northwestern Kansas: U.S. Geological Survey Water-Resources Investigations 41-74, 11 p.

Kaul, R. B., 1975, Native vegetation of Nebraska (Circa, 1850): Nebraska University-Lincoln, Institute of Agriculture and Natural Resources, 1 map.

Keech, C. F., and Bentall, Ray, 1971, Dunes on the plains--the sandhills region of Nebraska: Nebraska University, Conservation and Survey Division Resource Report 4, 18 p.

Knapp, R. M., Green, D. W., Pogge, E. C., and Stanford, C., 1975, Development and testing of a basin hydrology simulator: Water Resources Research, v. 11, no. 6, p. 879-889.

Konikow, L. F., 1976, Preliminary digital model of ground-water flow in the Madison Group, Powder River Basin, and adjacent areas, Wyoming, Montana, South Dakota, North Dakota, and Nebraska: U.S. Geological Survey Water-Resources Investigations 63-75, 44 p.

Lappala, E. G., (in prep.), Ground-water hydrology of the northern High Plains of New Mexico: New Mexico State Engineer Technical Report, 200 p.

______1976, Changes in the water supply in the Upper Republican Natural Resources District, southwest Nebraska, from 1952-75: U.S. Geological Survey Open-File Report 76-498, 17 p.

Leonard, G. J., and Huntoon, P. W., 1974, Groundwater geology of Southwest Nebraska Groundwater Conservation District: Nebraska University, Institute of Agriculture and Natural Resources, Water Survey Paper 37, 37 p.

Linsley, R. K., Jr., Kohler, M. A., and Paulhus, J. L., 1958, Hydrology for engineers: New York, McGraw-Hill, 340 p.

Lohman, S. W., 1972, Ground-water hydraulics: U.S. Geological Survey Professional Paper 708, 70 p.

Lohman, S. W., and others, 1972, Definitions of selected ground-water terms--revisions and conceptual refinements: U.S. Geological Survey Water-Supply Paper 1988, 21 p.

Longenbaugh, R. A., 1970, Evaluation of the method to compute volumes of water pumped from power records: State of Colorado, Division of Water Resources, final report for State Engineer's office, 24 p.

______1975, Computer estimates of natural recharge from soil moisture data, High Plains of Colorado: Colorado State University, Environmental Resources Center, Completion Report, ser. no. 64, 58 p.

Luckey, R. R., 1973, Digital model of the hydrologic system, northern High Plains of Colorado - A preliminary report: Colorado Water Conservation Board Water Resources Circular 19, 40 p.

Luckey, R. R., and Hofstra, W. E., 1974, Digital model of the Ogallala aquifer of the northern part of the northern High Plains of Colorado: Colorado Water Conservation Board Water Resources Circular 24, 22 p.

McLaughlin, T. G., 1948, Ground water in the Julesburg area, Colorado: Colorado Water Conservation Board Water Resources Circular 1, 20 p.

Missouri River Basin Commission, 1975, Stream-aquifer hydrology, Platte River Basin, Nebraska, Level B Study: Missouri River Basin Commission technical paper, 67 p.

National Oceanic and Atmospheric Administration, Nebraska and Colorado climatological data: NOAA annual and summary reports.

Nebraska Department of Agriculture, 1941-75, Nebraska agricultural statistics: Nebraska Department of Agriculture in cooperation with U.S. Department of Agriculture Statistical Reporting Service annual reports.

Nebraska Department of Water Resources, 1955-74, Biennial hydrographic reports: Lincoln, Nebr.

______(annual), Reports of diversions, Frenchman River and tributaries and Republican River and tributaries west of Range 30: Nebraska Water Resources Department unpublished records.

Nebraska Natural Resources Commission, 1976, Republican River Basin water quality management plan: Nebraska Natural Resources Commission draft report, 246 p.

Nie, N. H., Hull, C. H., Jenkins, J. G., Steinbrenner, K., and Bent, D. H., 1975, Statistical package for the social sciences, 2d edition: New York, McGraw-Hill, 675 p.

Osborn, C. A., and Caswell, A. E., 1957, Final geological report, Enders Dam, Frenchman-Cambridge Unit, Missouri Basin Project: U.S. Bureau of Reclamation, 19 p.

Osborn, C. A., Dirmeyer, R. D., and Nesbitt, A. P., 1947, Geologic report of the Wray Damsite, North Republican Unit, Missouri Basin Project: Indianola, Nebr., U.S. Bureau of Reclamation, 10 p.

Pikul, M. F., Street, R. L., and Remson, Irwin, 1974, A numerical model based on coupled one-dimensional Richards and Boussinesq equations: Water Resources Research, v. 10, no. 2, p. 295-303.

Pinder, G. F., 1969, An iterative digital model for aquifer evaluations: U.S. Geological Survey open-file report, 43 p.

Redell, D. L., 1967, Distribution of ground-water recharge: Colorado State University, Agricultural Engineering Department technical report, 132 p.

Reed, E. C., and Pickin, Rauf, (no date), Table for determining hydraulic conductivity from grain-size descriptions: Nebraska University, Institute of Agriculture and Natural Resources, unpublished table.

Rogers, R. F., and Wolfanger, L. A., 1919, Soil survey of Chase County, Nebraska: U.S. Department of Agriculture, advance sheets of Field Operations of Bureau of Soils, 1917, 66 p.

Rosenberg, N. J., 1964, Solar energy and sunshine in Nebraska: Nebraska University, Agriculture and Home Economics College Research Bulletin 213, 31 p.

Sautter, H. E., Huber, N. W., Kolbo, A. D., Lebruska, R. L., and Cunningham, R. L., 1963, Soil survey of Dundy County, Nebraska: U.S. Department of Agriculture, Soil Conservation Service, Soil Survey, ser. 1959, no. 20, 86 p.

Searcy, J. K., and Hardison, C. H., 1960, Double mass curves: U.S. Geological Survey Water-Supply Paper 1541-B, 66 p.

Thomssen, D. W., 1976, Geologic studies, spring flow improvement project, Rock Creek fish hatchery, Parks, Nebraska: Lincoln, Nebr., Hoskins-Western-Sonderegger, Inc., consulting engineers' report prepared for Nebraska Games and Parks Commission, 19 p.

Trescott, P. C., Pinder, G. F., and Larson, S. P., 1976, Finite difference model for aquifer simulation in two dimensions with results of numerical experiments: U.S. Geological Survey Techniques of Water-Resources Investigations, book 7, chap. 1, 116 p.

Upper Republican Natural Resources District, 1976, One and six year plan, Perkins, Chase, and Dundy Counties, 82 p.

U.S. Bureau of Reclamation, 1974, Supplemental water-supply study, Frenchman-Cambridge Division, Nebraska: U.S. Bureau of Reclamation appraisal report, 25 p.

U.S. Department of Commerce, 1968, Climatic atlas of the United States: U.S. Department of Commerce, ESSA environmental data service, 80 p.

U.S. Geological Survey, 1958, Compilation of records of surface waters of the United States through September 1950, Part 6-B: U.S. Geological Survey Water-Supply Paper 1310.

______1964, Compilation of records of surface waters of the United States, October 1950 to September 1960, Part 6-B: U.S. Geological Survey Water-Supply Paper 1730.

______1969, Surface water supply of the United States 1961-65, Part 6, Volume 3: U.S. Geological Survey Water-Supply Paper 1918.

______1973, Surface water supply of the United States, 1966-70, Part 6, Volume 3: U.S. Geological Survey Water-Supply Paper 2118.

______1971-75, Water resources data for Nebraska, Part 1: U.S. Geological Survey annual reports.

Waite, H. A., Reed, E. C., and Jones, D. S., Jr., 1946, Ground water in the Republican River Basin in Nebraska: Nebraska University, Conservation and Survey Division, Nebraska Water Resources Survey Water-Supply Paper 1, pt. 4, 38 p.

Waite, H. A., and others, 1948, Progress report on the ground-water hydrology of the Republican and Frenchman River Valleys, with a section on The chemical quality of the ground water, by H. A. Swenson: U.S. Geological Survey Circular 19, 83 p.

Wenzel, L. K., and Waite, H. A., 1941, Ground water in Keith County, Nebraska, with sections on Platte Valley Public Power and Irrigation District, Sutherland project, by E. E. Halmos, and Central Nebraska Public Power and Irrigation District, Tri-County project, by G. E. Johnson: U.S. Geological Survey Water-Supply Paper 848, 68 p.

PLATE 1 - MONTHLY STREAMFLOW HYDROGRAPHS FOR SELECTED SITES.

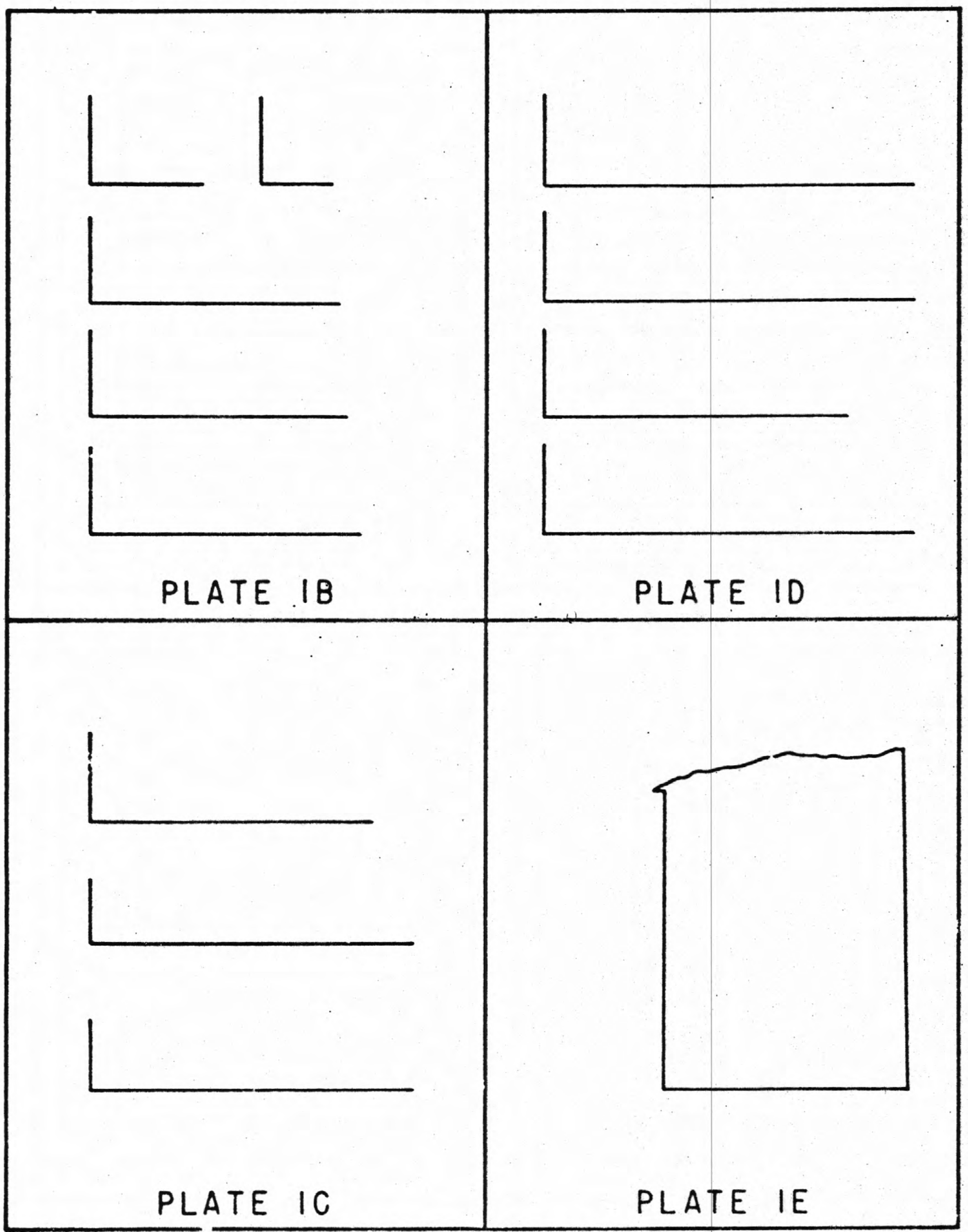


DIAGRAM SHOWING COMPONENTS OF PLATE 1

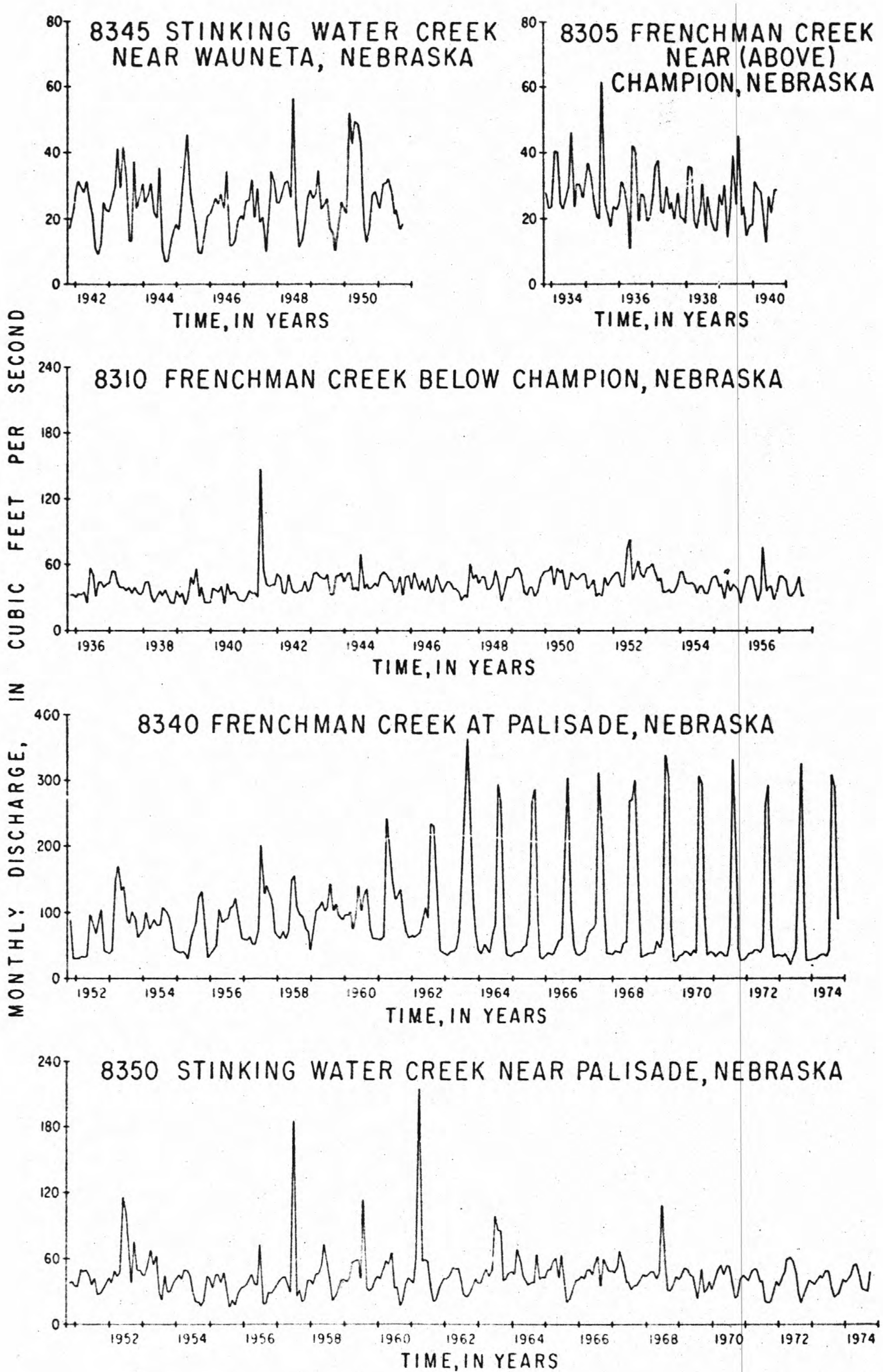
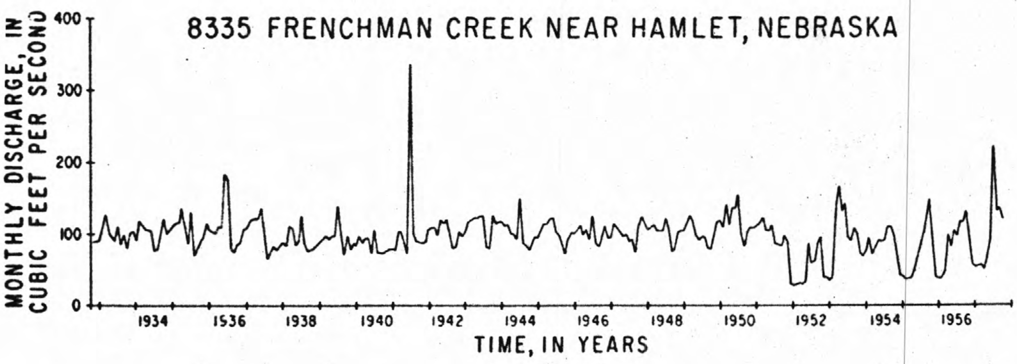

MONTHLY DISCHARGE, IN CUBIC FEET PER SECOND
8345 STINKING WATER CREEK NEAR WAUNETA, NEBRASKA
0
20
40
60
80
1942
1944
1946
1948
1950
TIME, IN YEARS
8305 FRENCHMAN CREEK NEAR (ABOVE) CHAMPION, NEBRASKA
0
20
40
60
80
1934
1936
1938
1940
TIME, IN YEARS
8310 FRENCHMAN CREEK BELOW CHAMPION, NEBRASKA
0
60
120
180
240
1936
1938
1940
1942
1944
1946
1948
1950
1952
1954
1956
TIME, IN YEARS
8340 FRENCHMAN CREEK AT PALISADE, NEBRASKA
0
100
200
300
400
1952
1954
1956
1958
1960
1962
1964
1966
1968
1970
1972
1974
TIME, IN YEARS
8350 STINKING WATER CREEK NEAR PALISADE, NEBRASKA
0
60
120
180
240
1952
1954
1956
1958
1960
1962
1964
1966
1968
1970
1972
1974
TIME, IN YEARS

PLATE 1C - MONTHLY STREAMFLOW HYDROGRAPHS FOR SELECTED SITES.



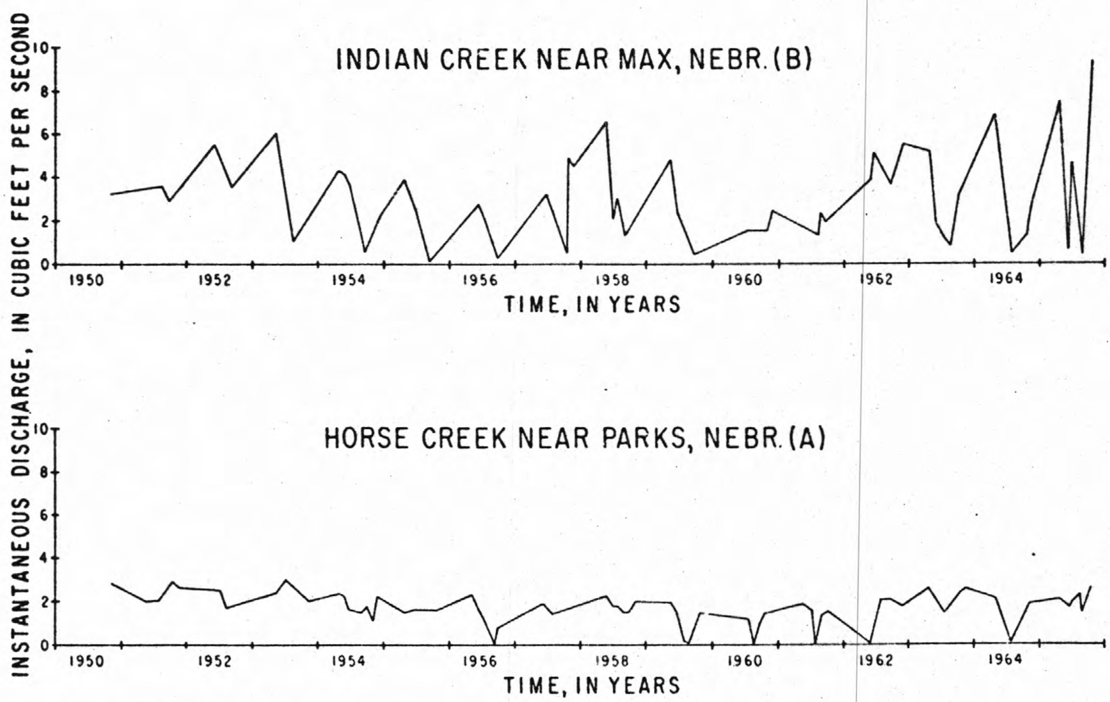


PLATE 1D - MONTHLY STREAMFLOW HYDROGRAPHS FOR SELECTED SITES.

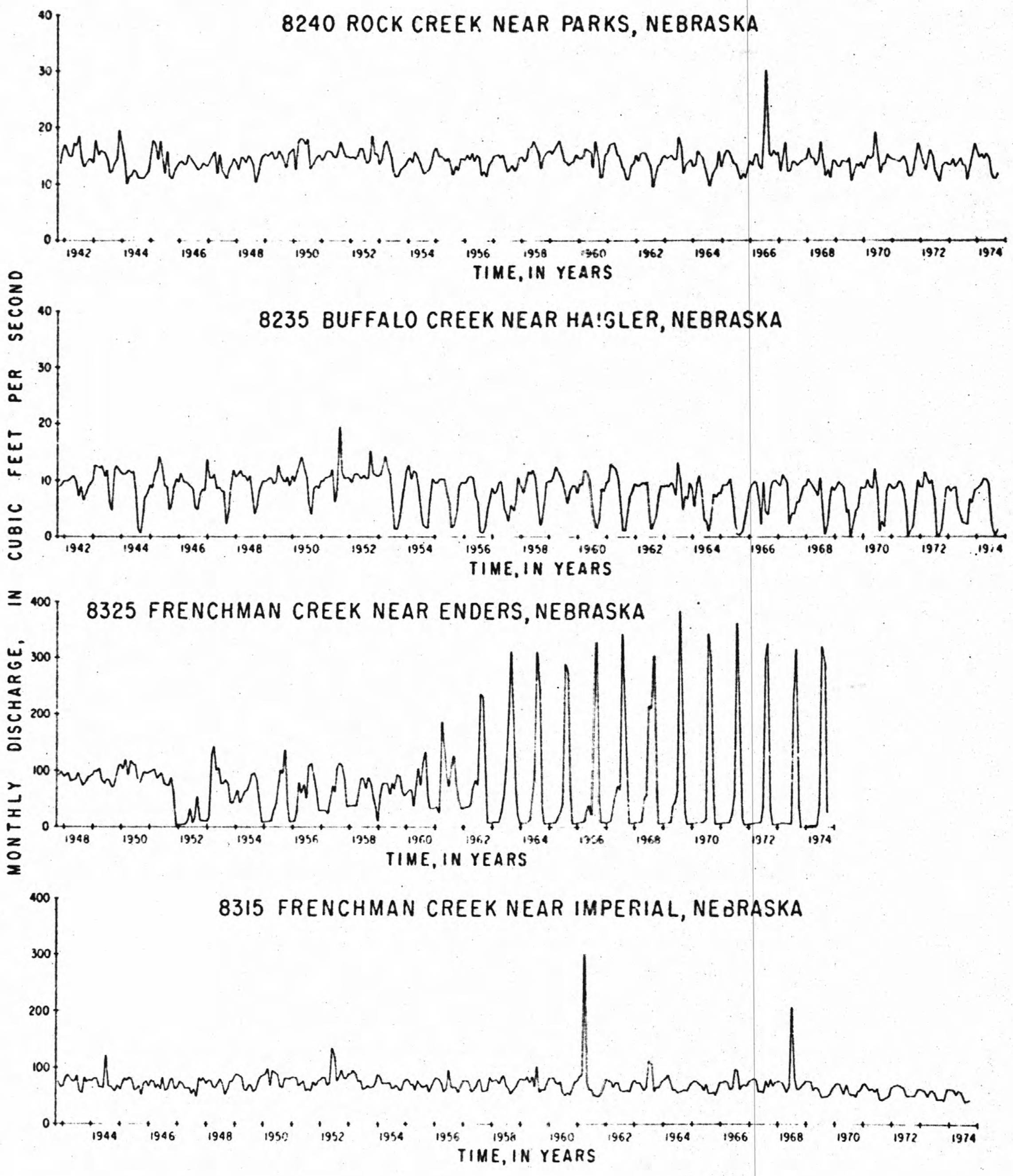


PLATE 1E - MONTHLY STREAMFLOW HYDROGRAPHS FOR SELECTED SITES.

LOCATION OF ACTIVE AND INACTIVE GAGING STATIONS

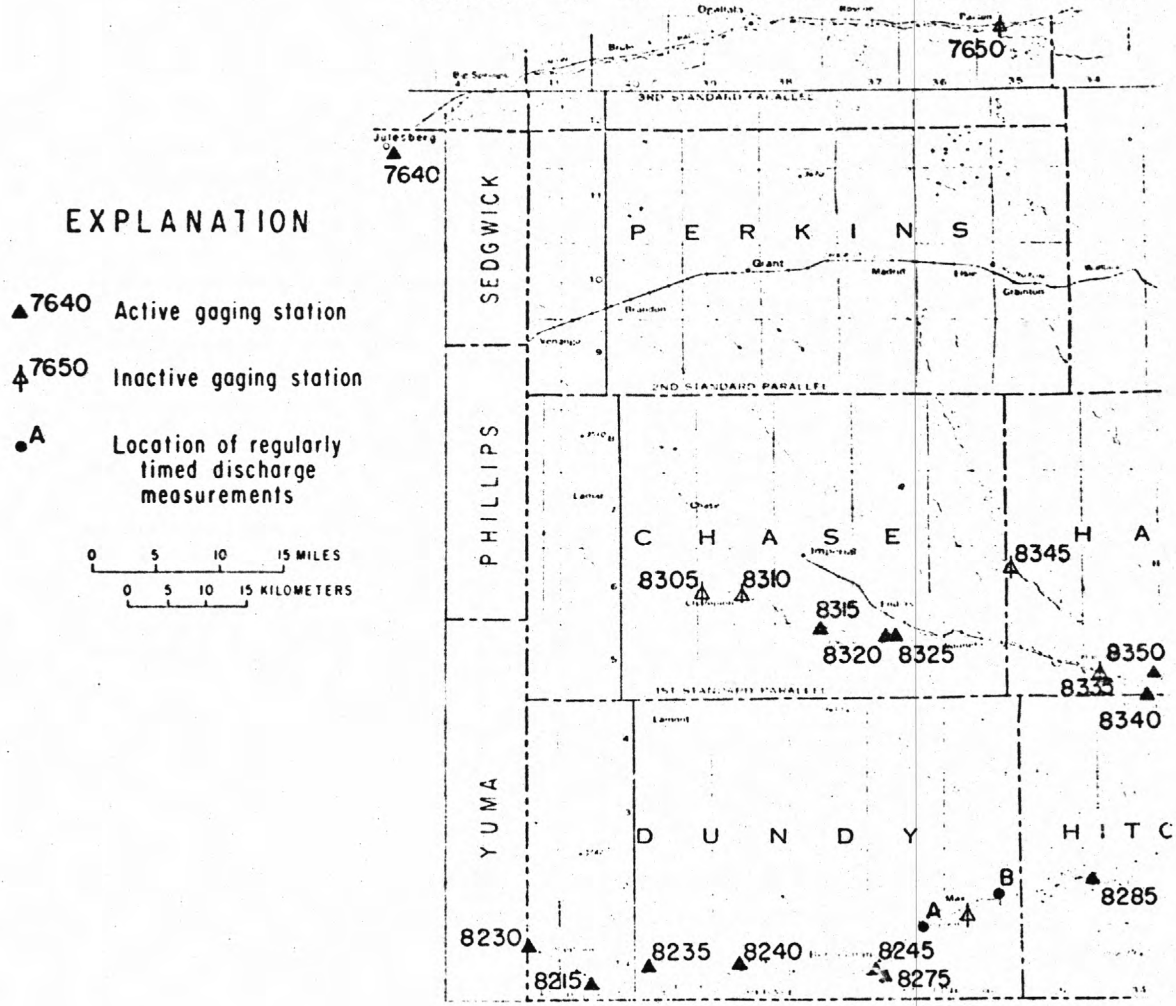


PLATE 2 - MEASURED AND MODELED STREAMFLOW GAINS AND LOSSES FOR SELECTED STREAMS.

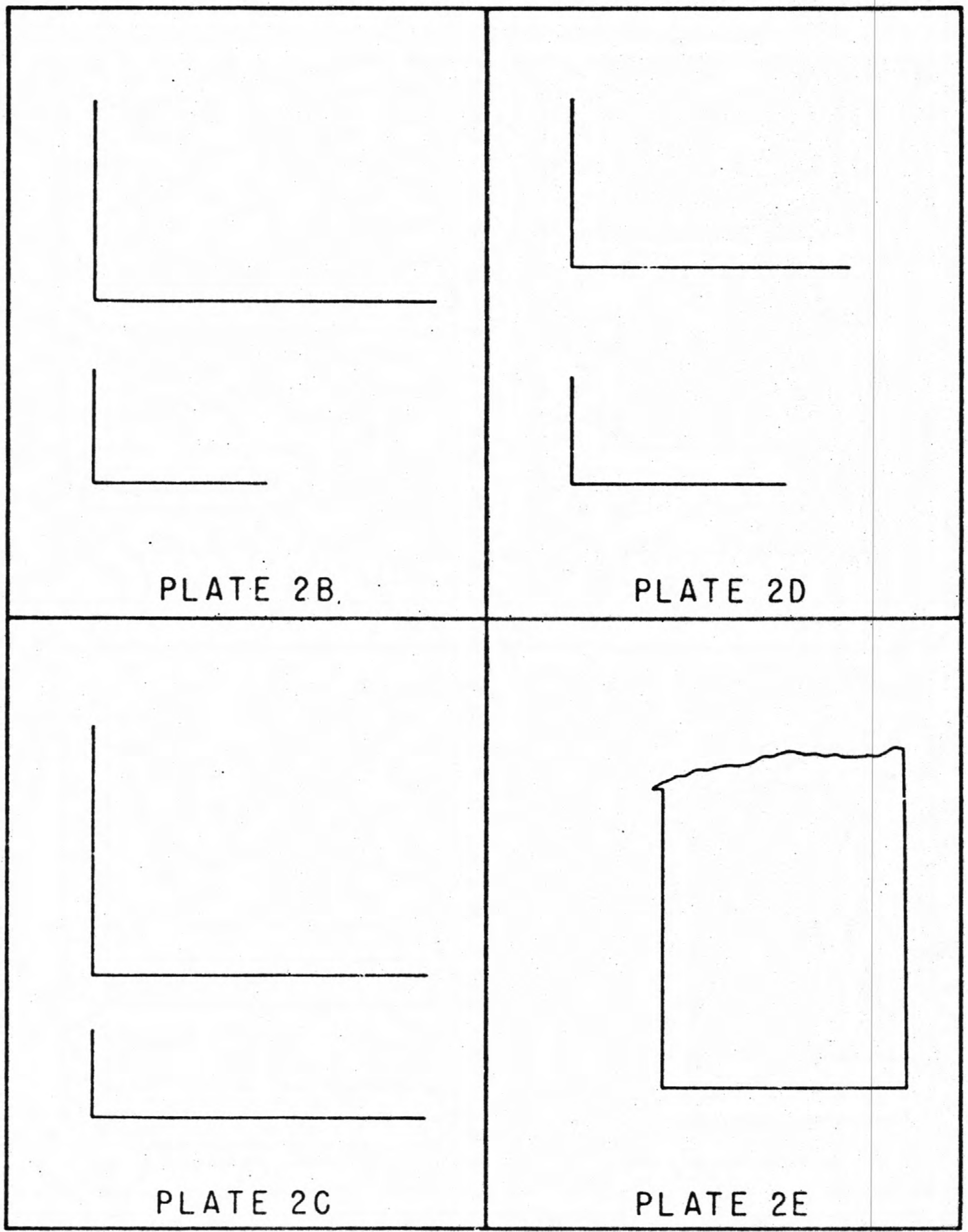


DIAGRAM SHOWING COMPONENTS OF PLATE 2

PLATE 2B - MEASURED AND MODELED STREAMFLOW GAINS AND LOSSES FOR SELECTED STREAMS.

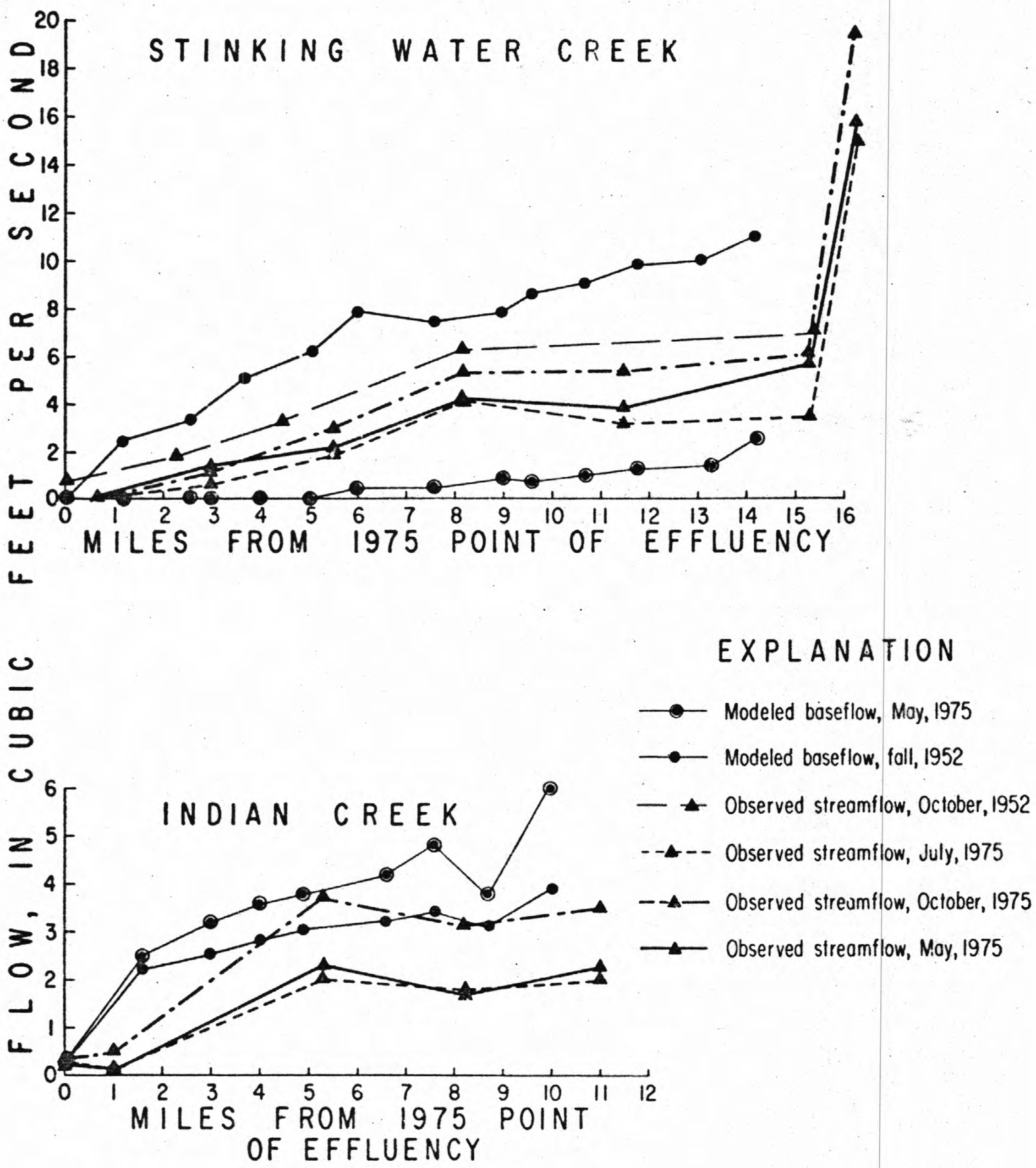
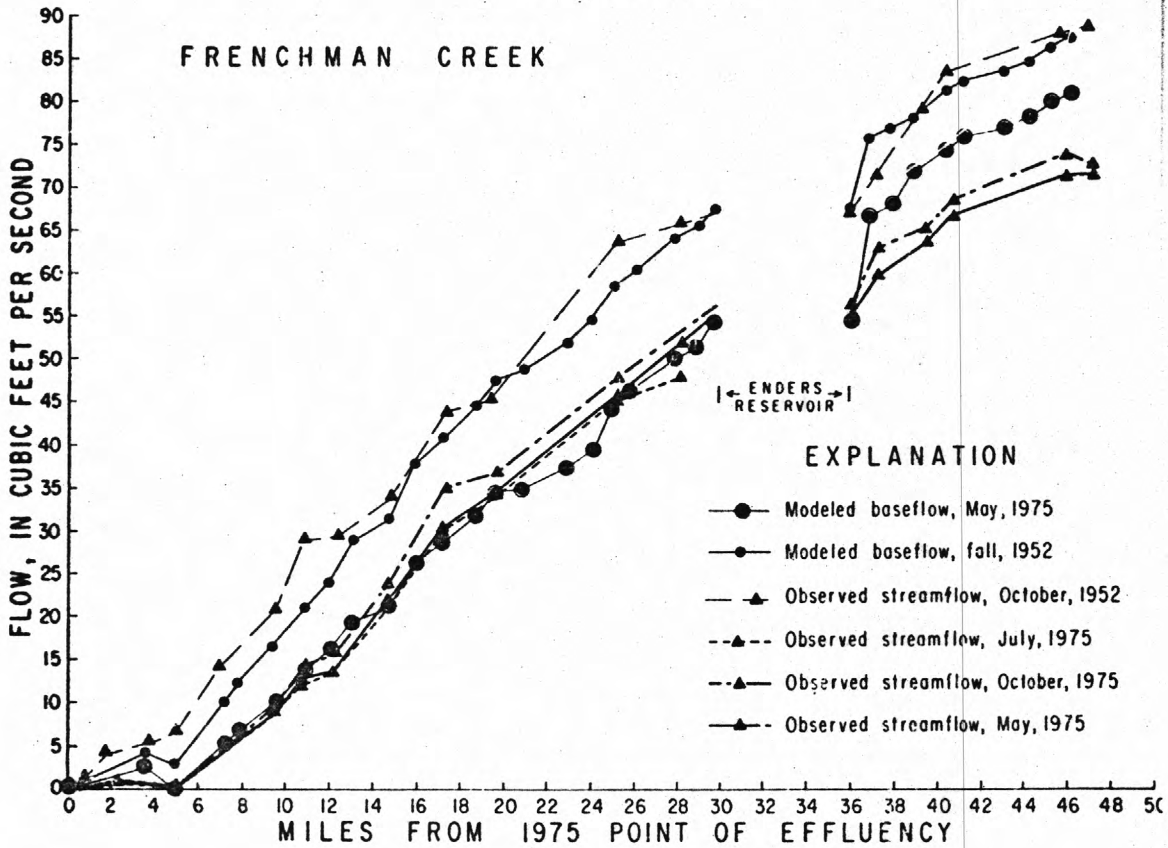


PLATE 2C - MEASURED AND MODELED STREAMFLOW GAINS AND LOSSES FOR SELECTED STREAMS.



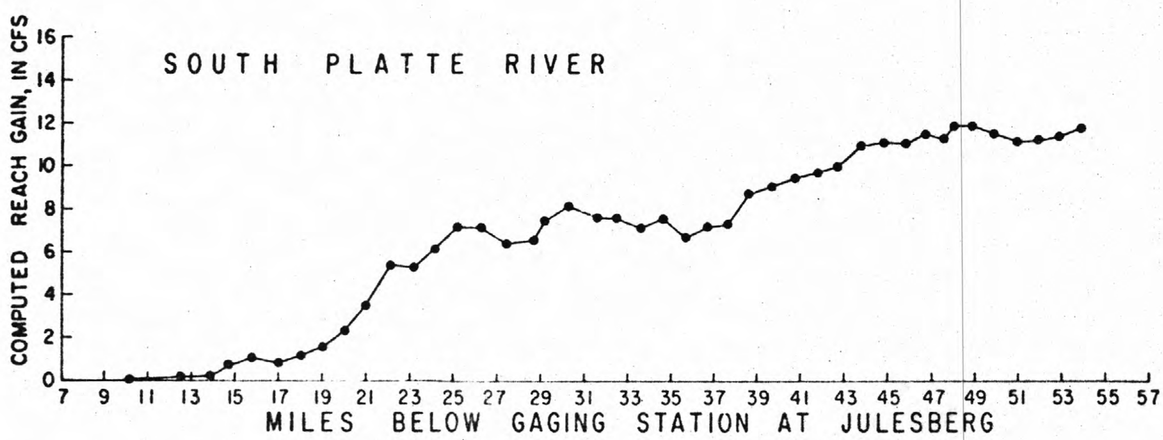


PLATE 2D - MEASURED AND MODELED STREAMFLOW GAINS AND LOSSES FOR SELECTED STREAMS.

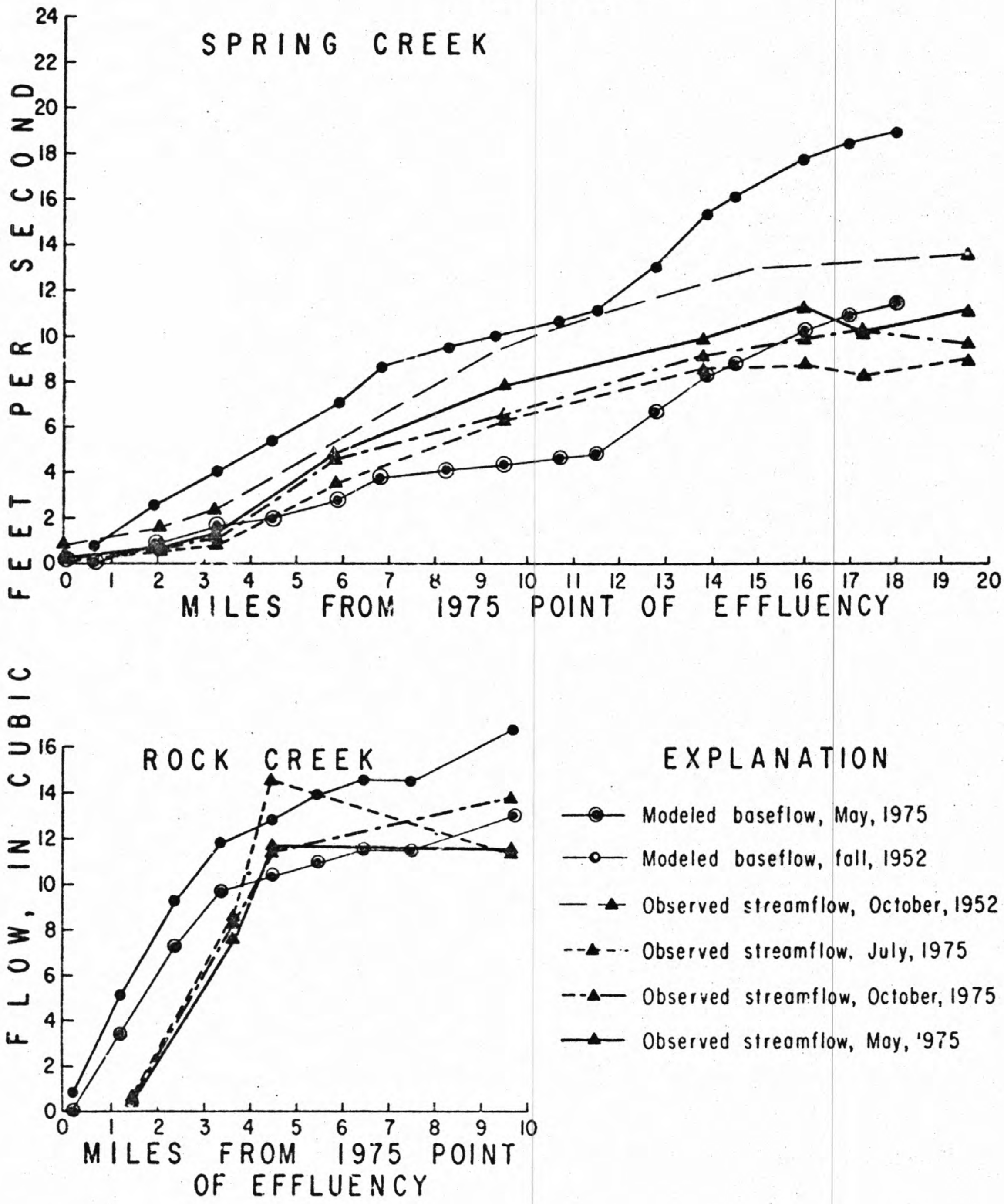


PLATE 2E - MEASURED AND MODELED STREAMFLOW GAINS AND LOSSES FOR SELECTED STREAMS.

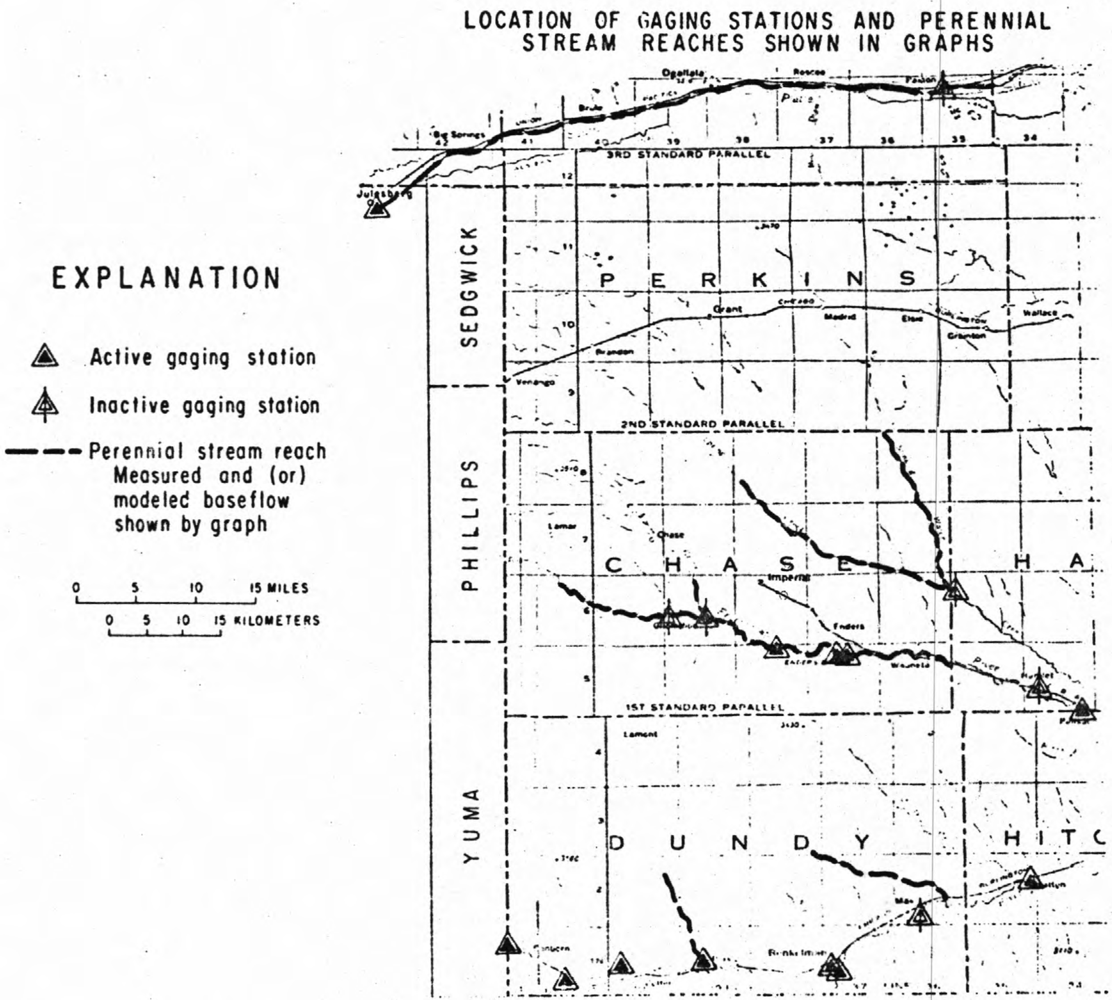


PLATE 3 - SINGLE MASS CURVES OF MONTHLY STREAMFLOW AND ANNUAL REACH GAINS FOR SELECTED STREAMS.

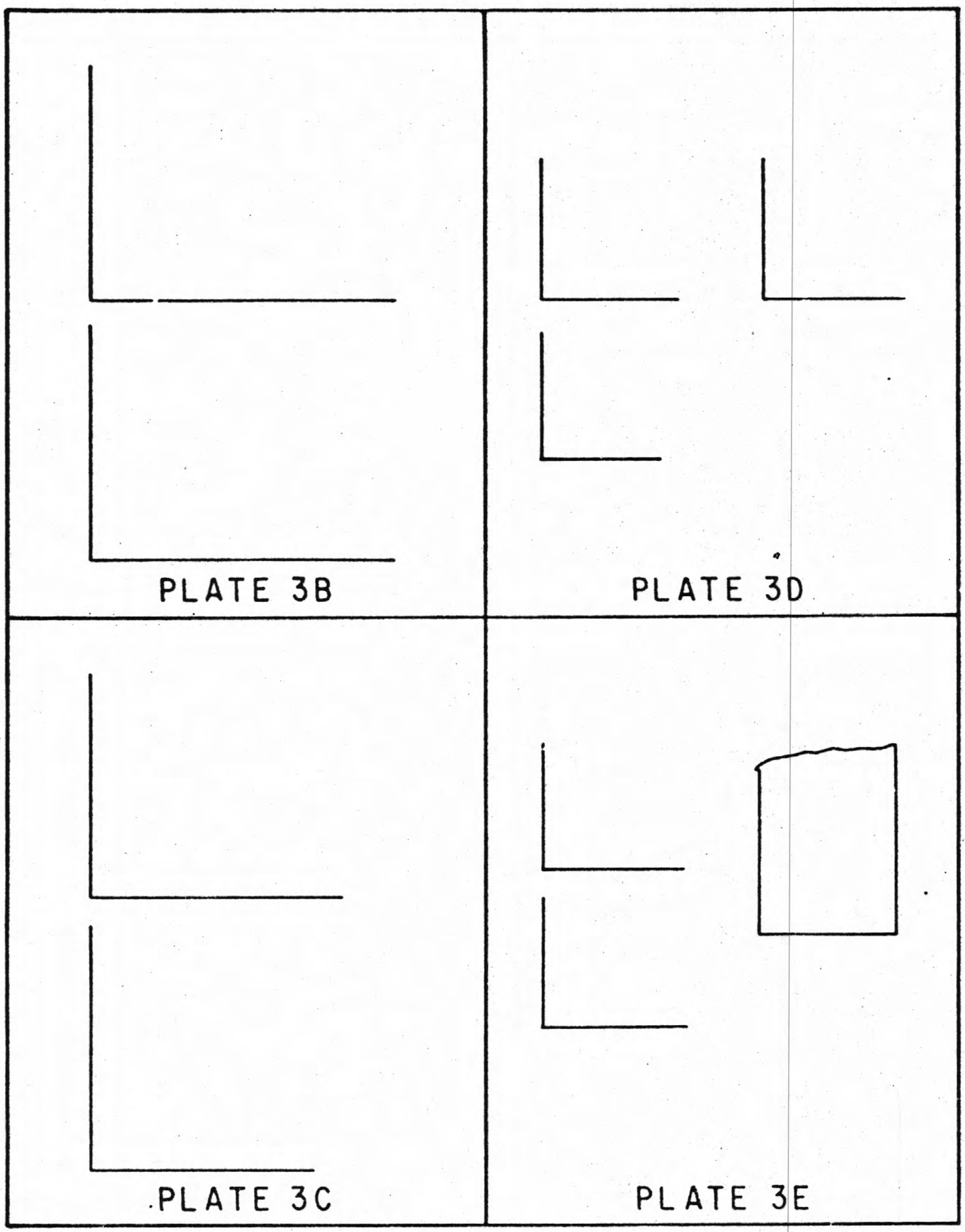
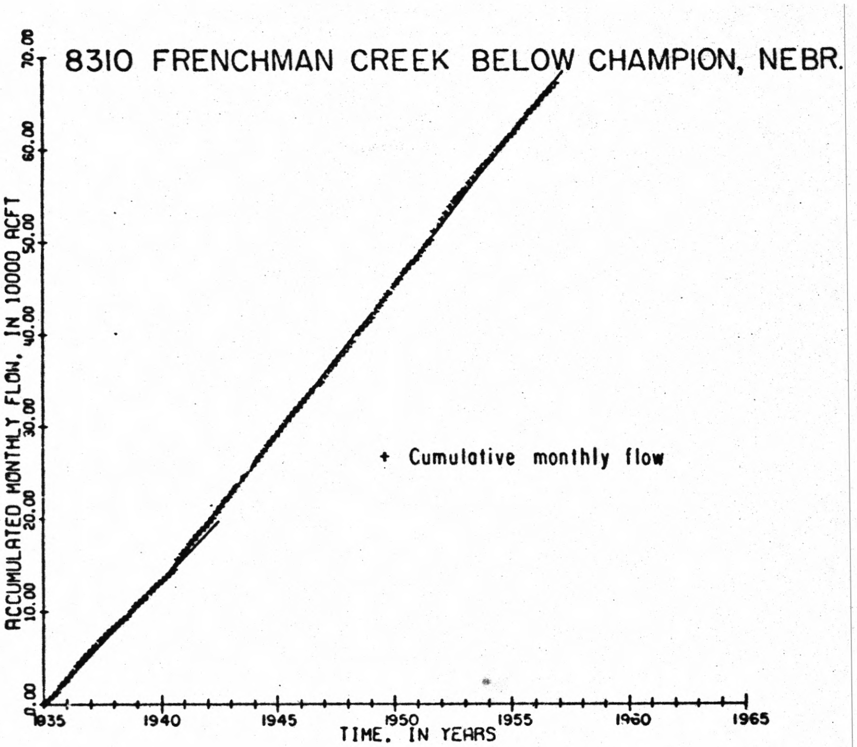


DIAGRAM SHOWING COMPONENTS OF PLATE 3

PLATE 3B - SINGLE MASS CURVES OF MONTHLY STREAMFLOW AND ANNUAL REACH GAINS FOR SELECTED STREAMS.



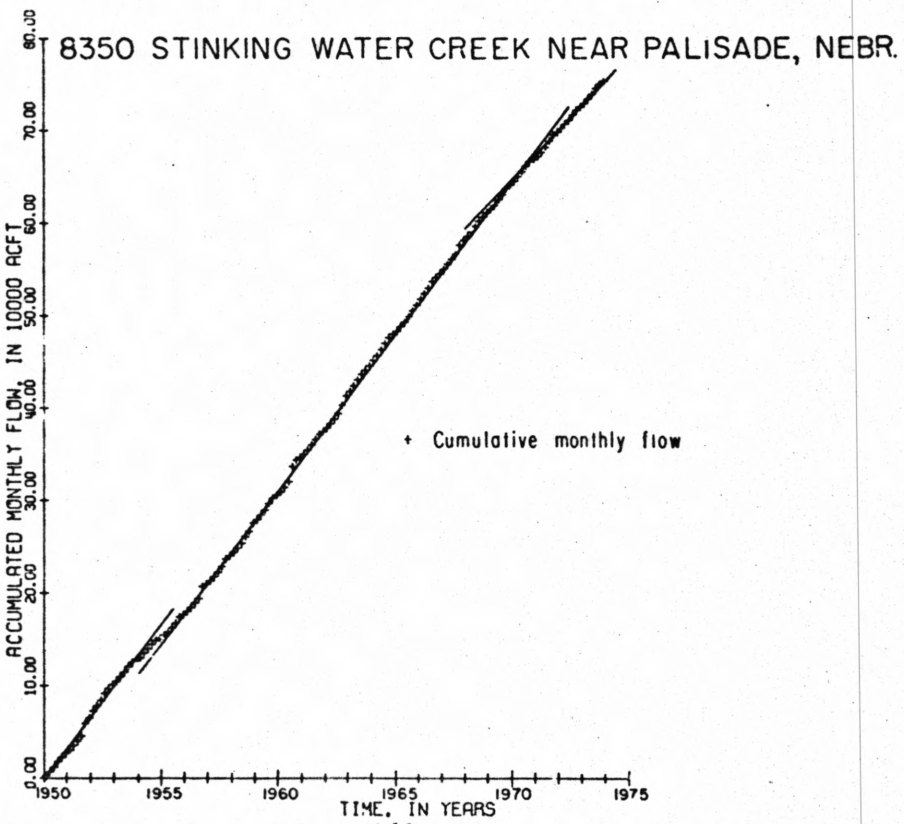
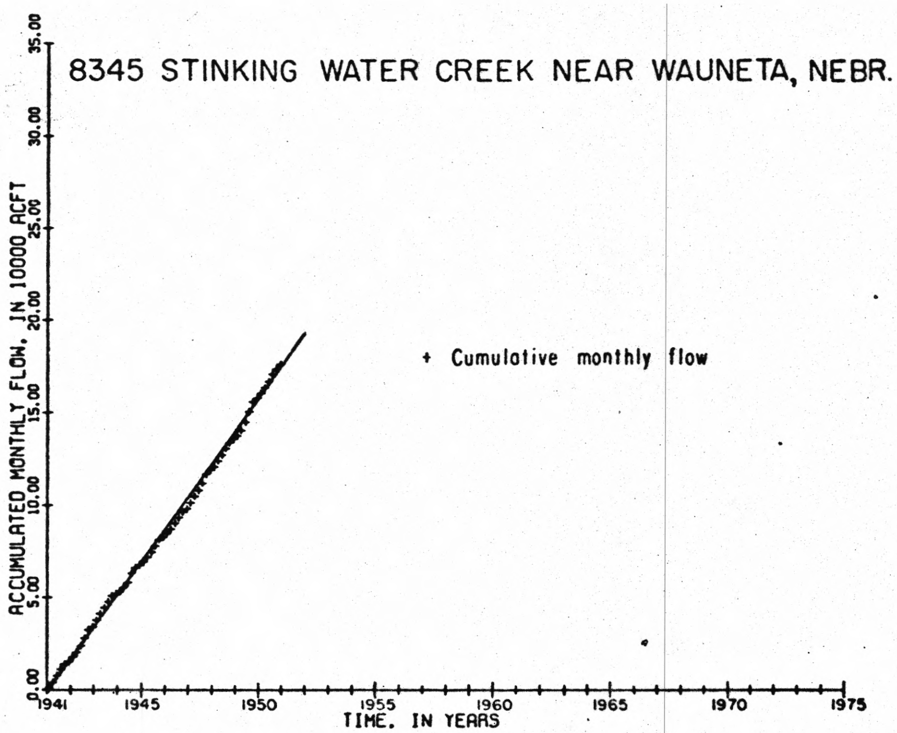
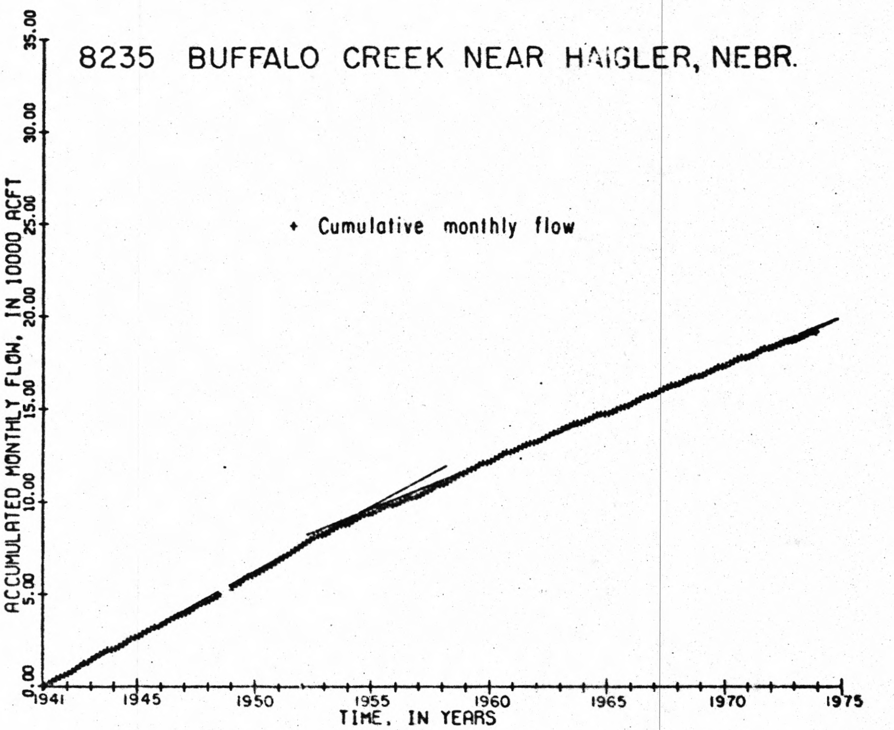


PLATE 3C - SINGLE MASS CURVES OF MONTHLY STREAMFLOW AND ANNUAL REACH GAINS FOR SELECTED STREAMS.





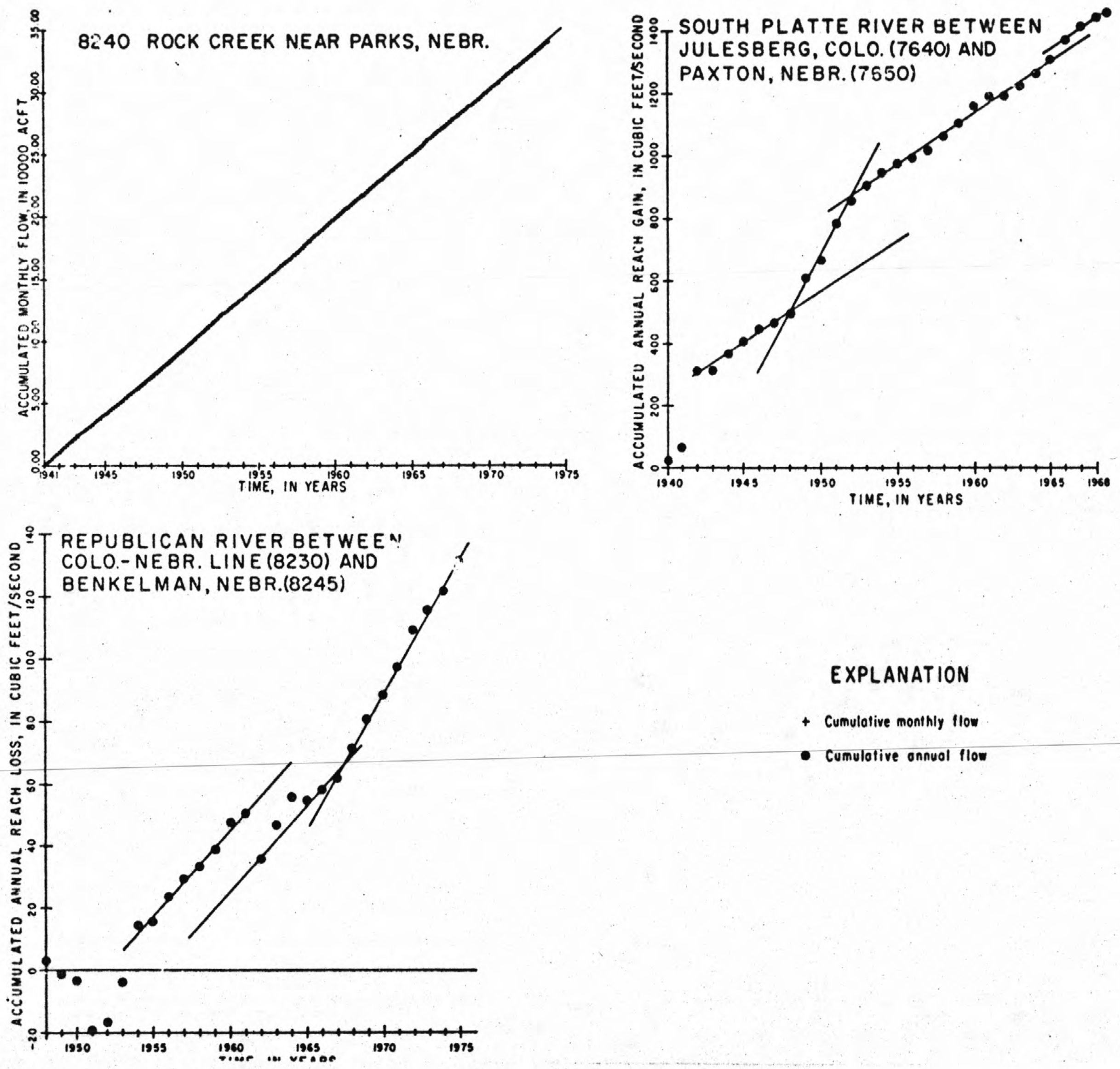
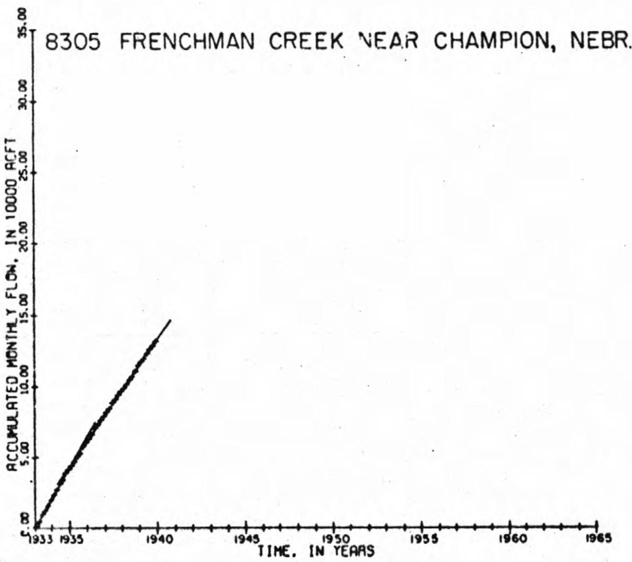
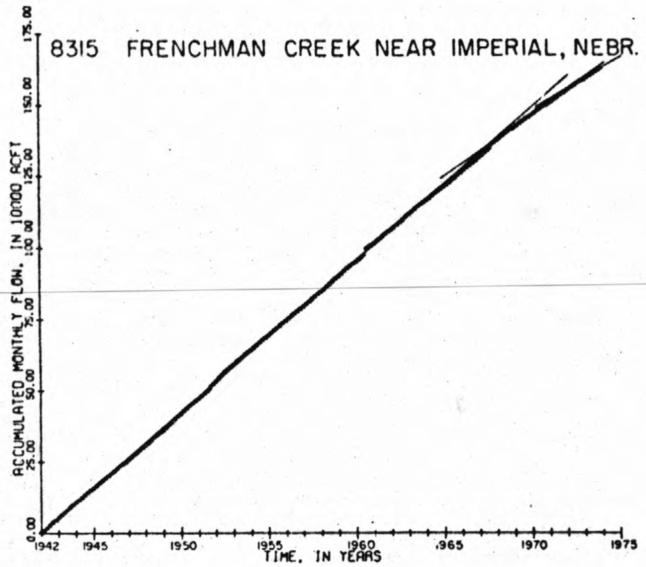


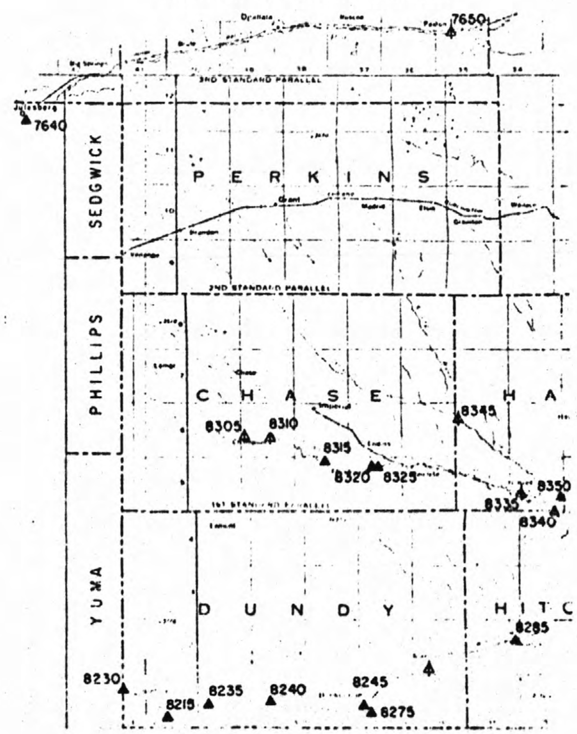
PLATE 3D - SINGLE MASS CURVES OF MONTHLY STREAMFLOW AND ANNUAL REACH GAINS FOR SELECTED STREAMS.

PLATE 3E - SINGLE MASS CURVES OF MONTHLY STREAMFLOW AND ANNUAL REACH GAINS FOR SELECTED STREAMS.





LOCATION OF GAGING STATIONS



EXPLANATION

+ Cumulative monthly flow

▲ 8305 Active gaging station, number refers to graphs

8345 Inactive gaging station, number refers to graphs

PLATE 4 - DOUBLE MASS CURVES OF MONTHLY STREAMFLOW OF DEPLETED STREAMS VS UNDEPLETED STREAMS, MONTHLY PRECIPITATION AT IMPERIAL VS MONTHLY FLOW OF FRENCHMAN CREEK NEAR IMPERIAL, AND RELATIONSHIP BETWEEN ANNUAL PRECIPITATION AT IMPERIAL AND ANNUAL FLOW OF FRENCHMAN CREEK NEAR IMPERIAL.

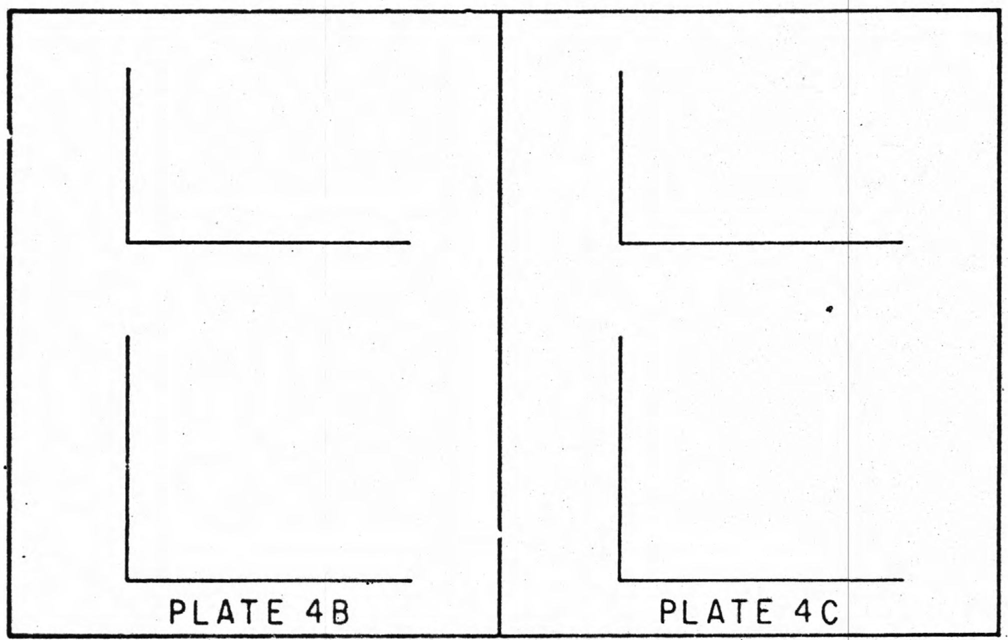
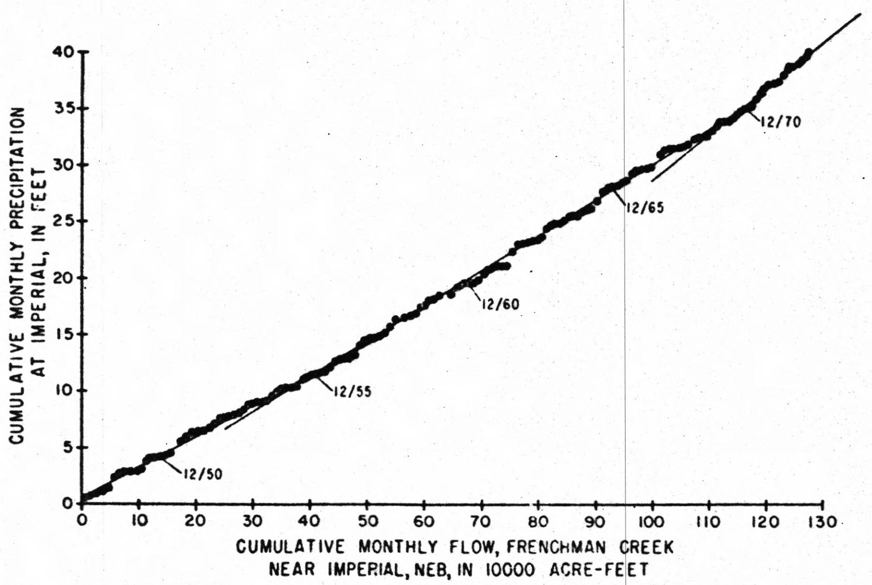
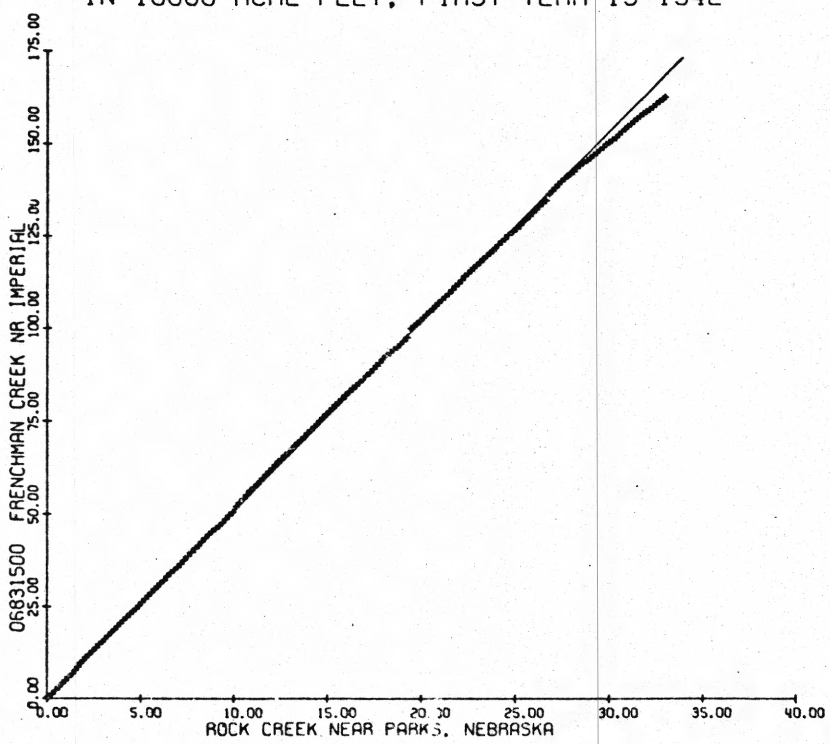


DIAGRAM SHOWING COMPONENTS OF PLATE 4

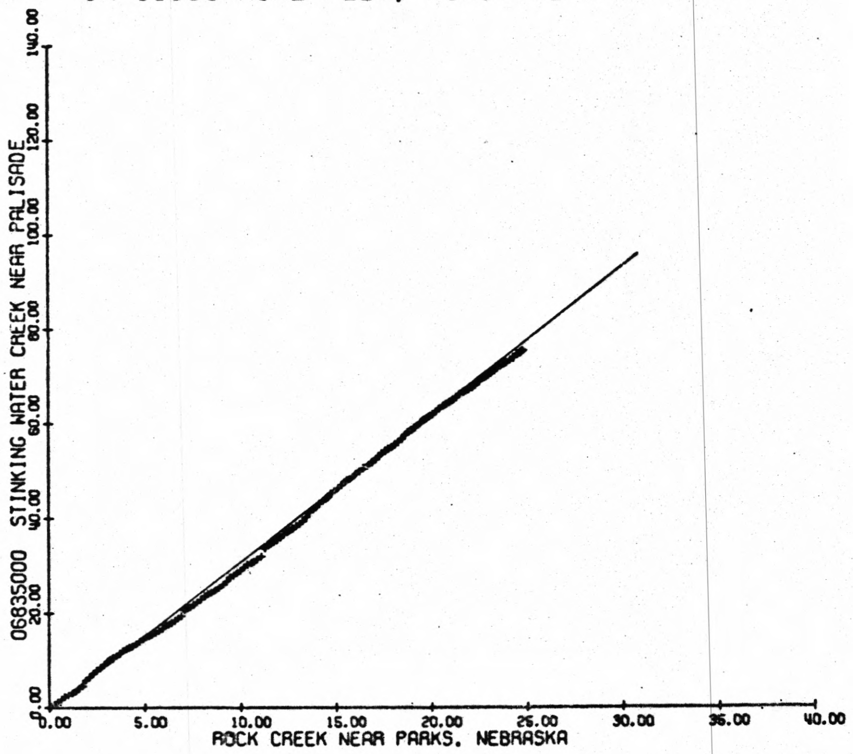
PLATE 4B - DOUBLE MASS CURVES OF MONTHLY STREAMFLOW OF DEPLETED STREAMS VS UNDEPLETED STREAMS, MONTHLY PRECIPITATION AT IMPERIAL VS MONTHLY FLOW OF FRENCHMAN CREEK NEAR IMPERIAL, AND RELATIONSHIP BETWEEN ANNUAL PRECIPITATION AT IMPERIAL AND ANNUAL FLOW OF FRENCHMAN CREEK NEAR IMPERIAL.



DOUBLE MASS CURVE OF MONTHLY FLOW
IN 10000 ACRE-FEET; FIRST YEAR IS 1942



DOUBLE MASS CURVE OF MONTHLY FLOW
IN 10000 ACRE-FEET; FIRST YEAR IS 1950



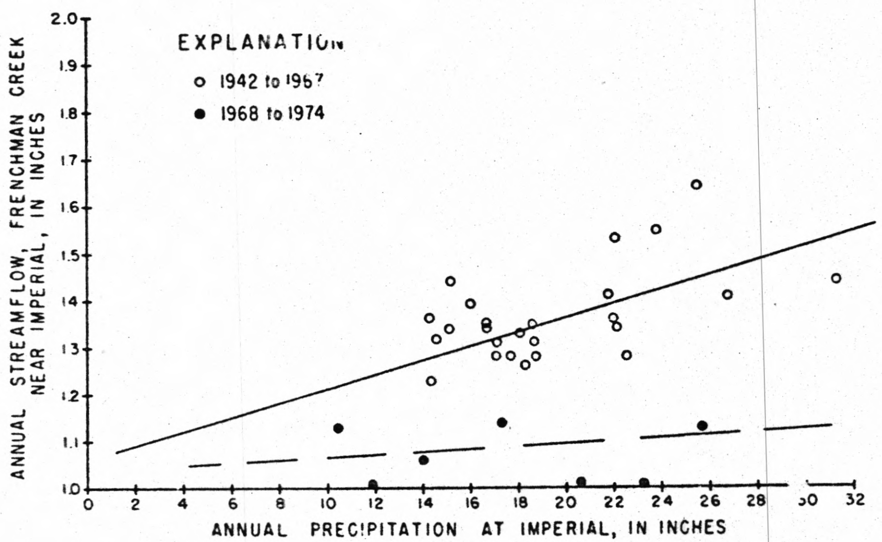


PLATE 4C - DOUBLE MASS CURVES OF MONTHLY STREAMFLOW OF DEPLETED STREAMS VS UNDEPLETED STREAMS, MONTHLY PRECIPITATION AT IMPERIAL VS MONTHLY FLOW OF FRENCHMAN CREEK NEAR IMPERIAL, AND RELATIONSHIP BETWEEN ANNUAL PRECIPITATION AT IMPERIAL AND ANNUAL FLOW OF FRENCHMAN CREEK NEAR IMPERIAL.

PLATE 5 - VEGETATIVE COVER AND LAND USE, MAY 1975

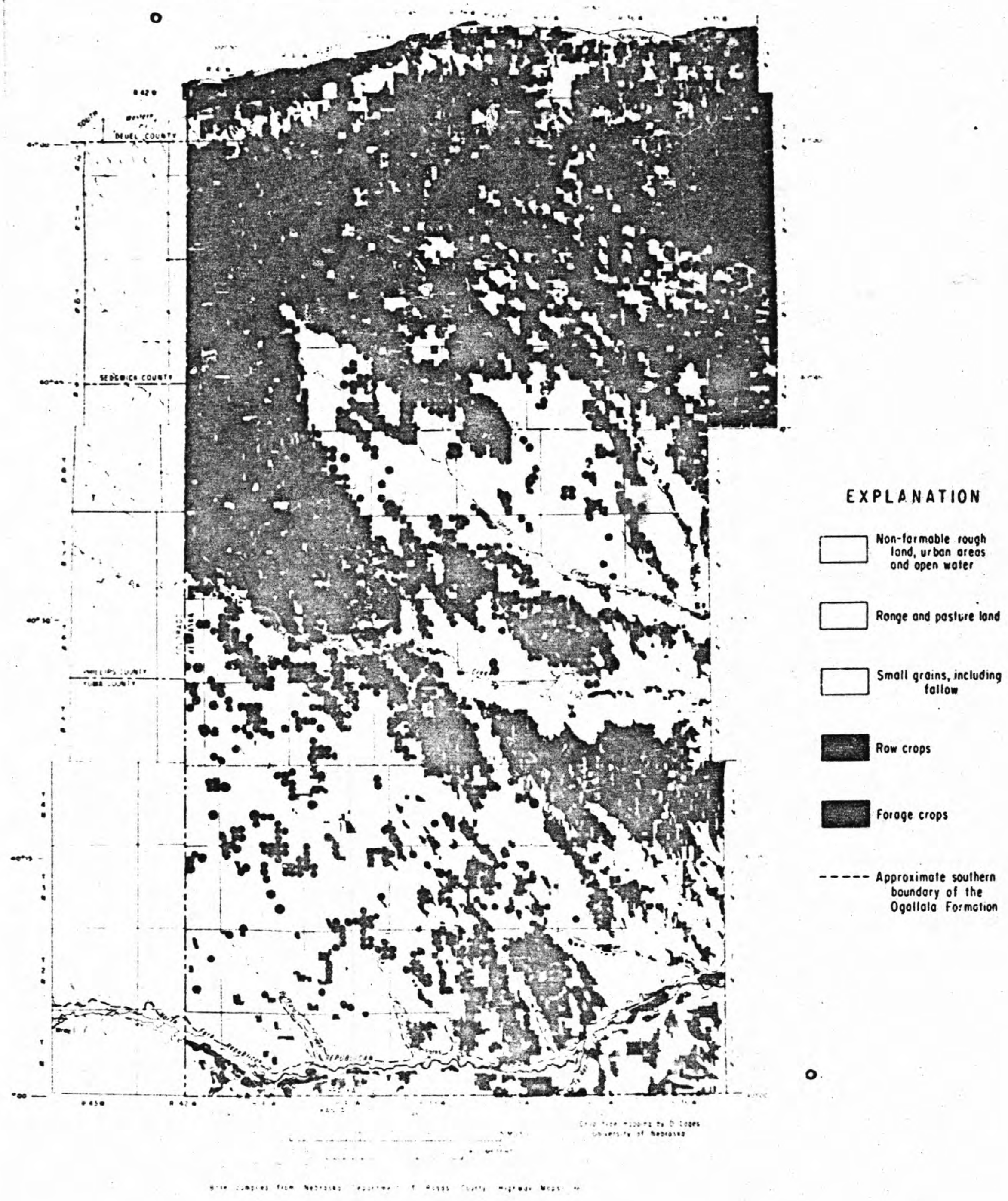


PLATE 6 - FIVE-YEAR MOVING MEAN PRECIPITATION AND ANNUAL DEPARTURE FROM THE 1925 THROUGH 1975 AVERAGE PRECIPITATION FOR BENKELMAN, IMPERIAL, MADRID, AND OGALLALA, NEBR., AND HOLYOKE, JULESBURG, AND WRAY, COLO.

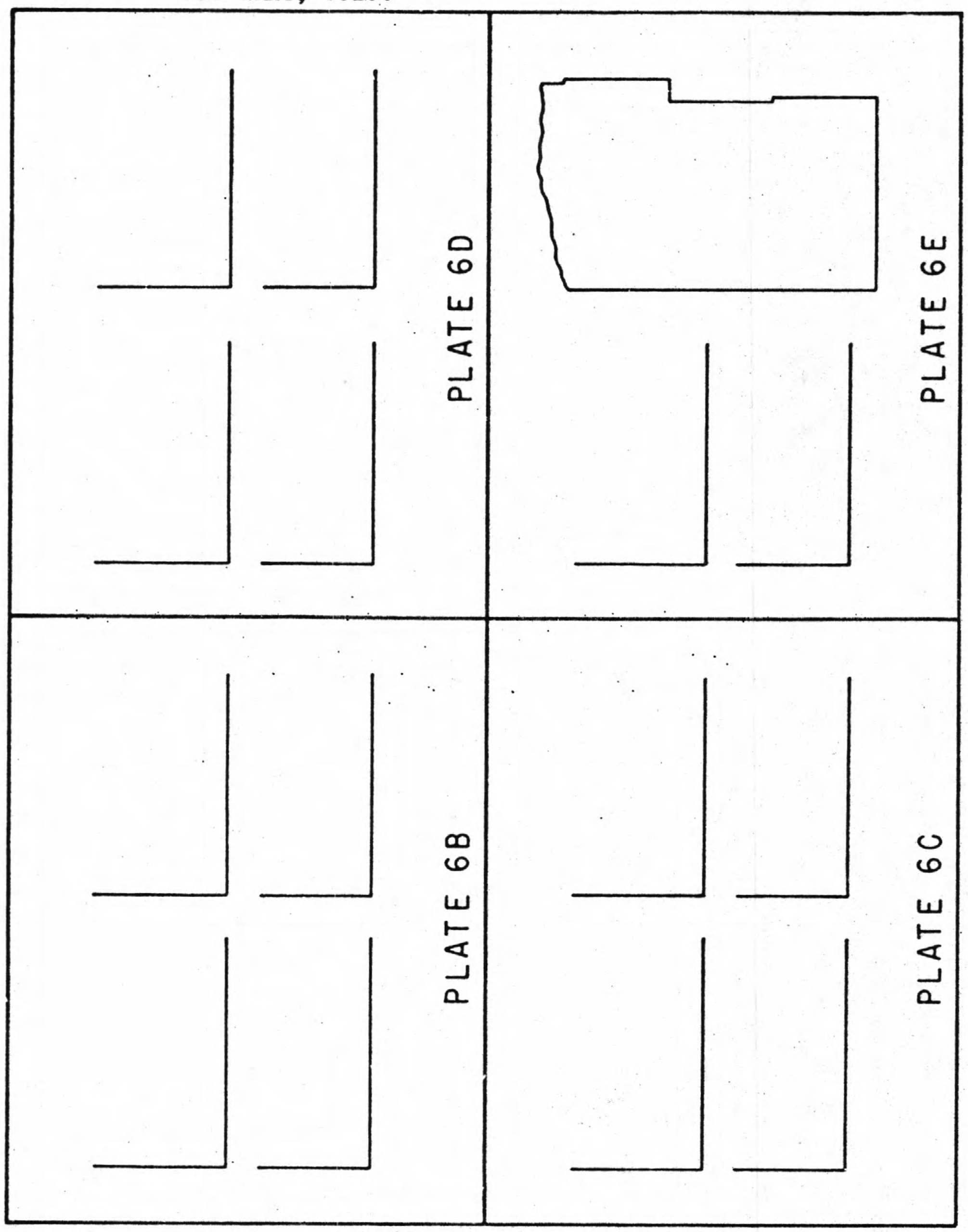


DIAGRAM SHOWING COMPONENTS OF PLATE 6

PLATE 6B - FIVE-YEAR MOVING MEAN PRECIPITATION AND ANNUAL DEPARTURE FROM THE 1925 THROUGH 1975 AVERAGE PRECIPITATION FOR BENKELMAN, IMPERIAL, MADRID, AND OGALLALA, NEBR., AND HOLYOKE, JULESBURG, AND WRAY, COLO.

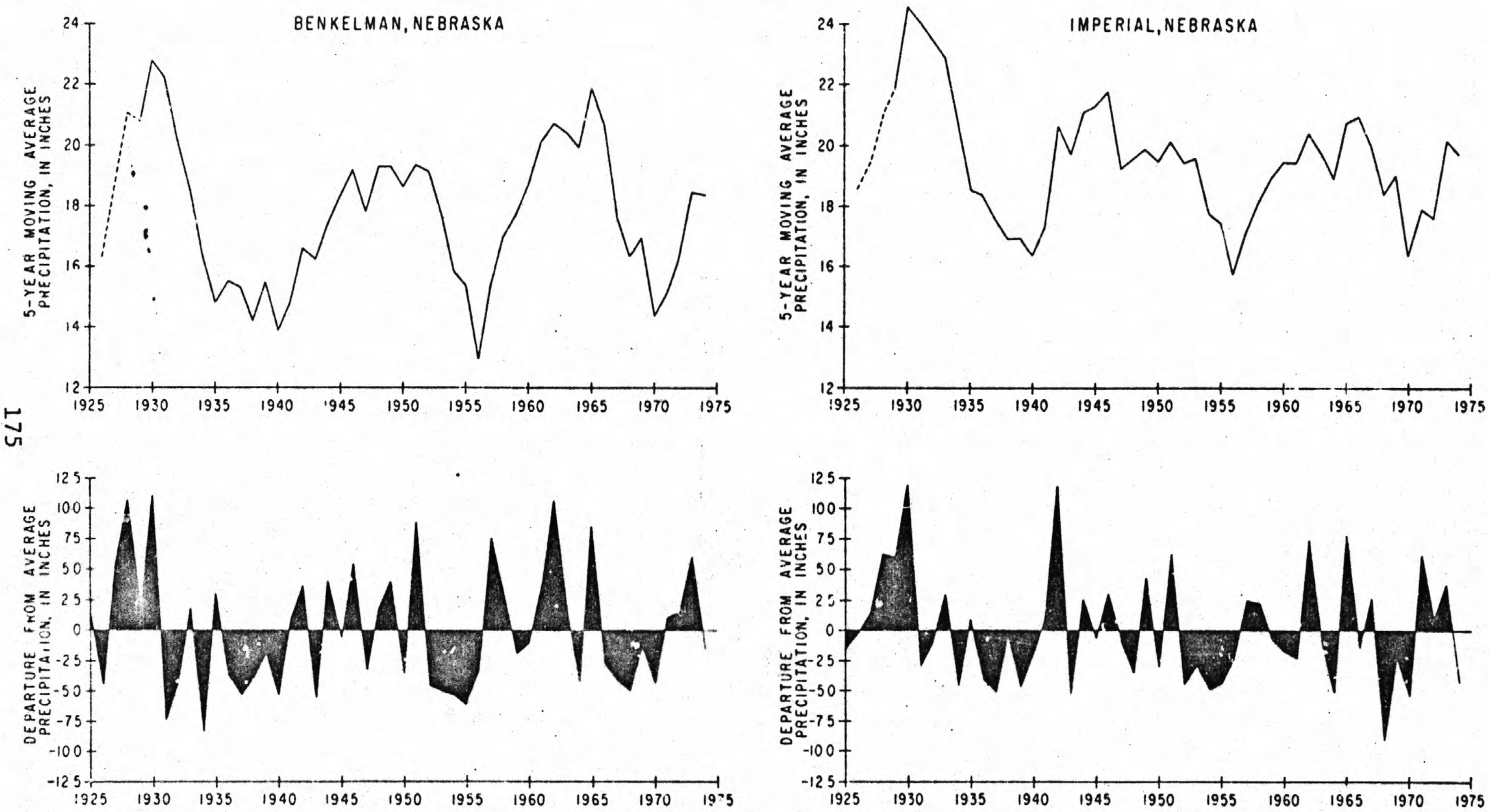


PLATE 6C - FIVE-YEAR MOVING MEAN PRECIPITATION AND ANNUAL DEPARTURE FROM THE 1925 THROUGH 1975 AVERAGE PRECIPITATION FOR BENKELMAN, IMPERIAL, MADRID, AND OGALLALA, NEBR., AND HOLYOKE, JULESBURG, AND WRAY, COLO.

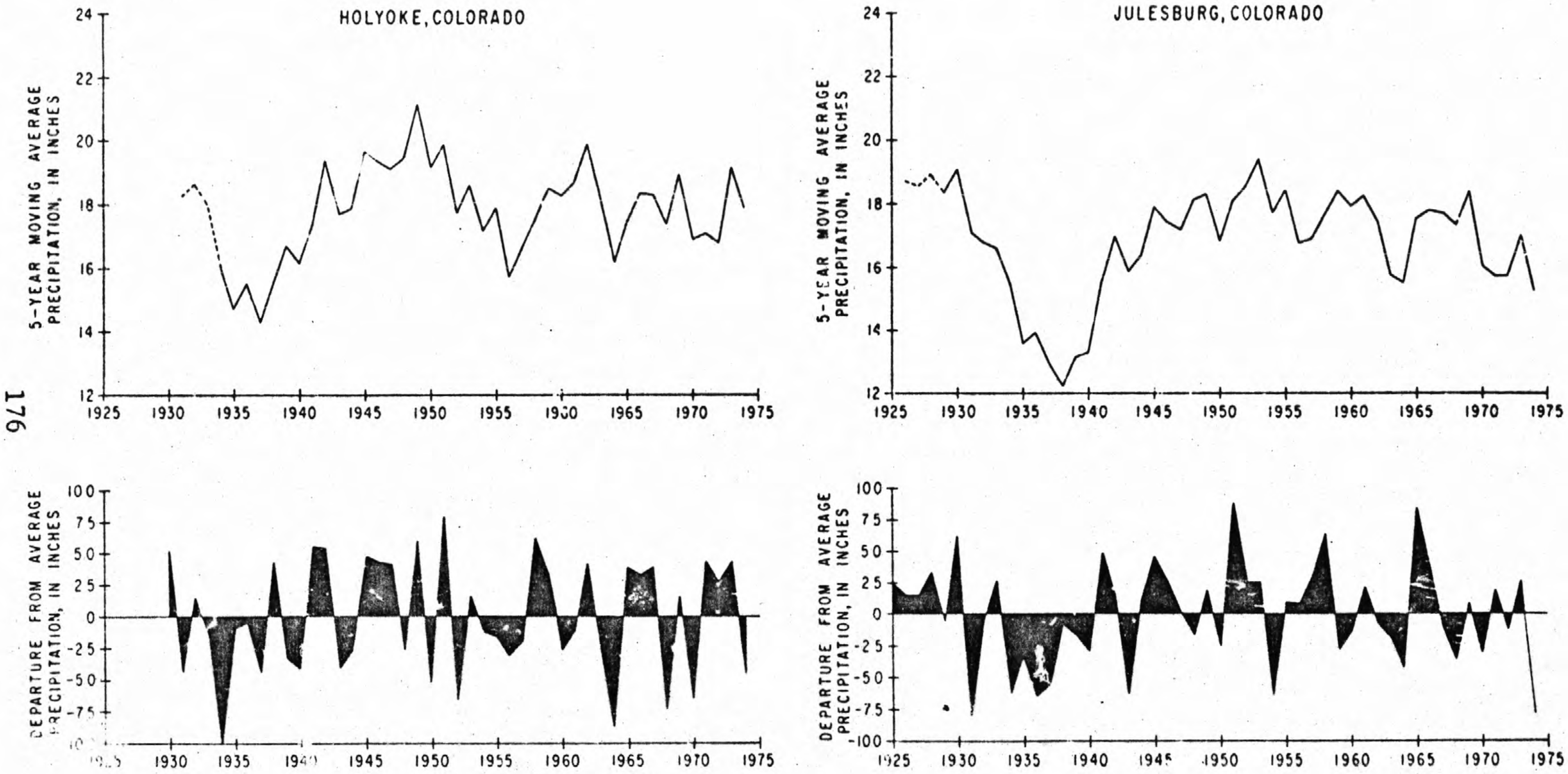


PLATE 6D - FIVE-YEAR MOVING MEAN PRECIPITATION AND ANNUAL DEPARTURE FROM THE 1925 THROUGH 1975 AVERAGE PRECIPITATION FOR BENKELMAN, IMPERIAL, MADRID, AND OGALLALA, NEBR., AND HOLYOKE, JULESBURG, AND WRAY, COLO.

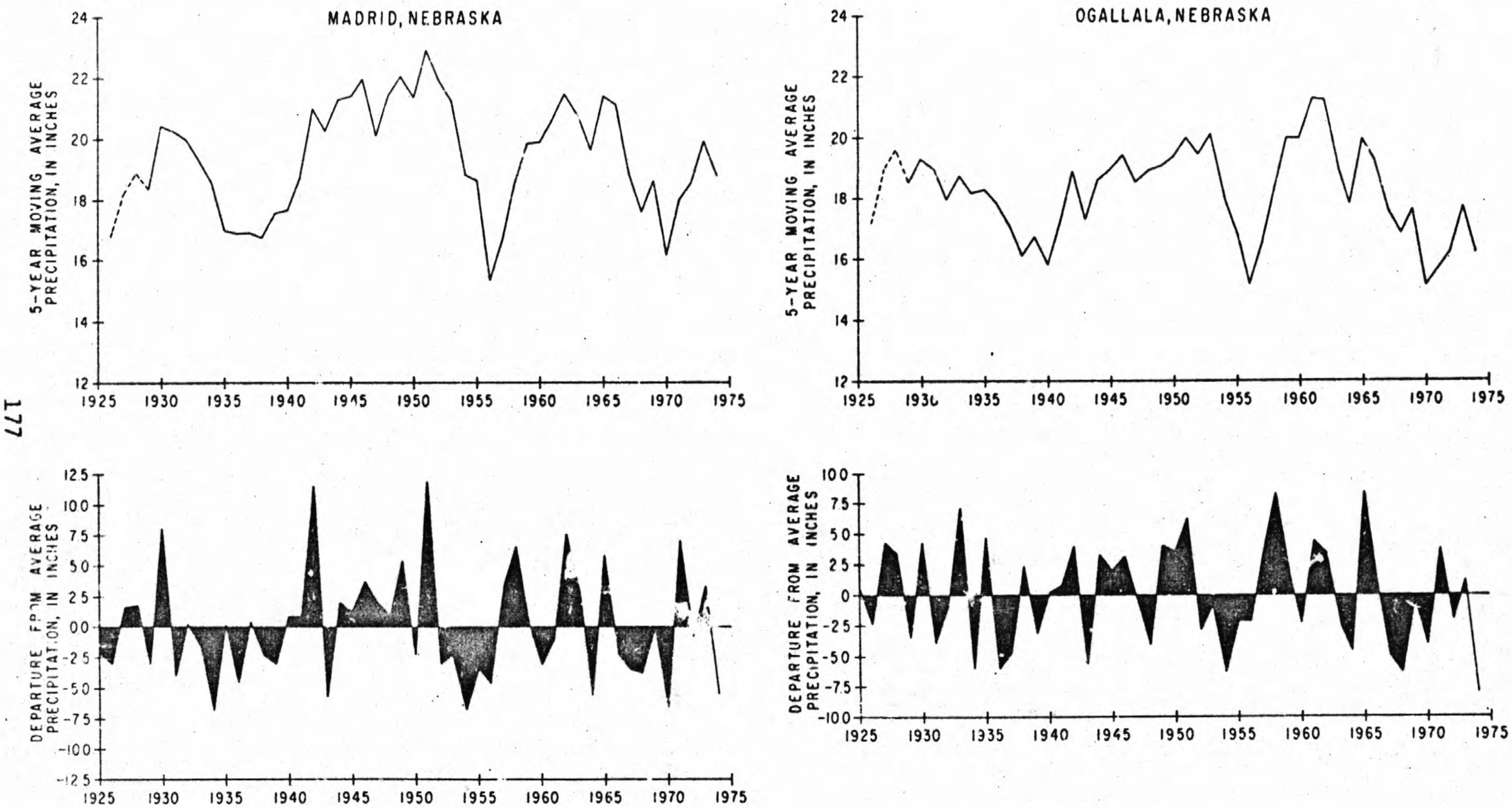
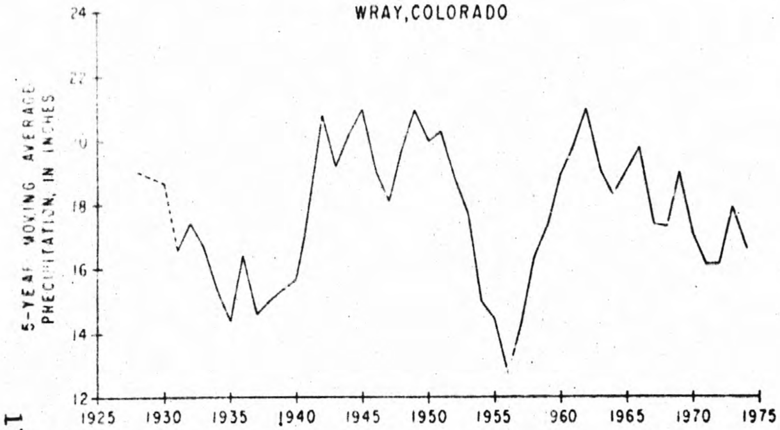
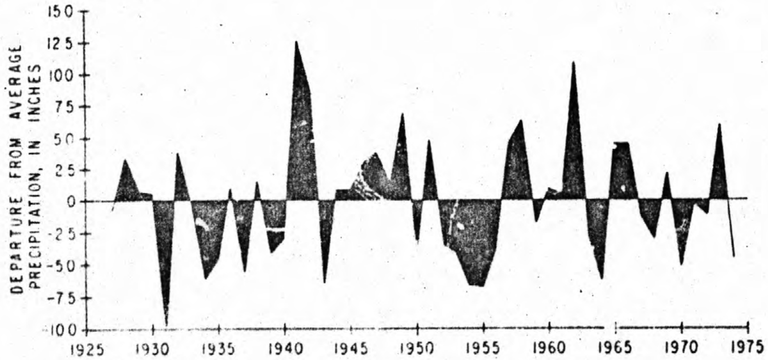
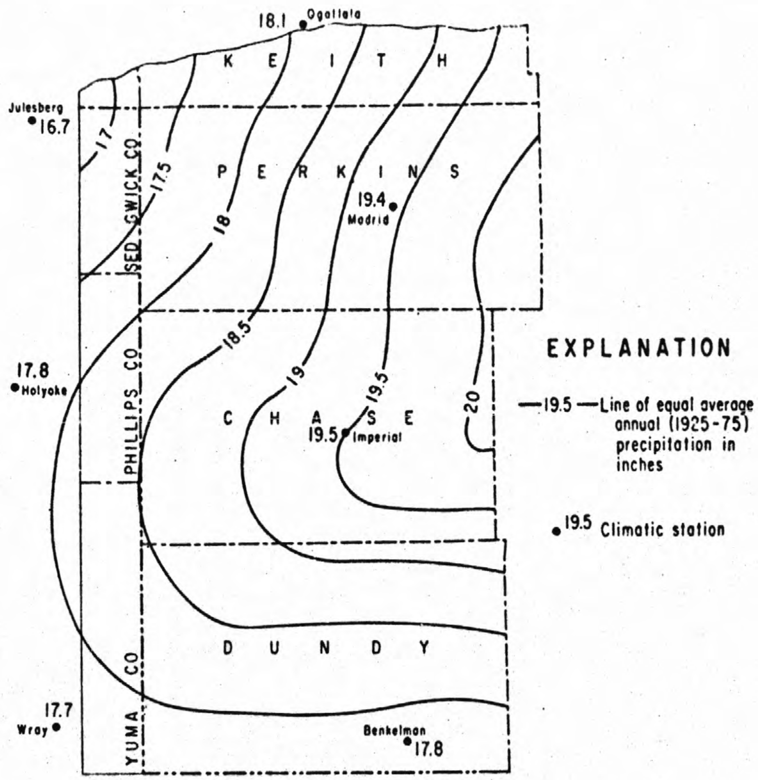


PLATE 6E - FIVE-YEAR MOVING MEAN PRECIPITATION AND ANNUAL DEPARTURE FROM THE 1925 THROUGH 1975 AVERAGE PRECIPITATION FOR BENKELMAN, IMPERIAL, MADRID, AND OGALLALA, NEBR., AND HOLYOKE, JULESBURG, AND WRAY, COLO.

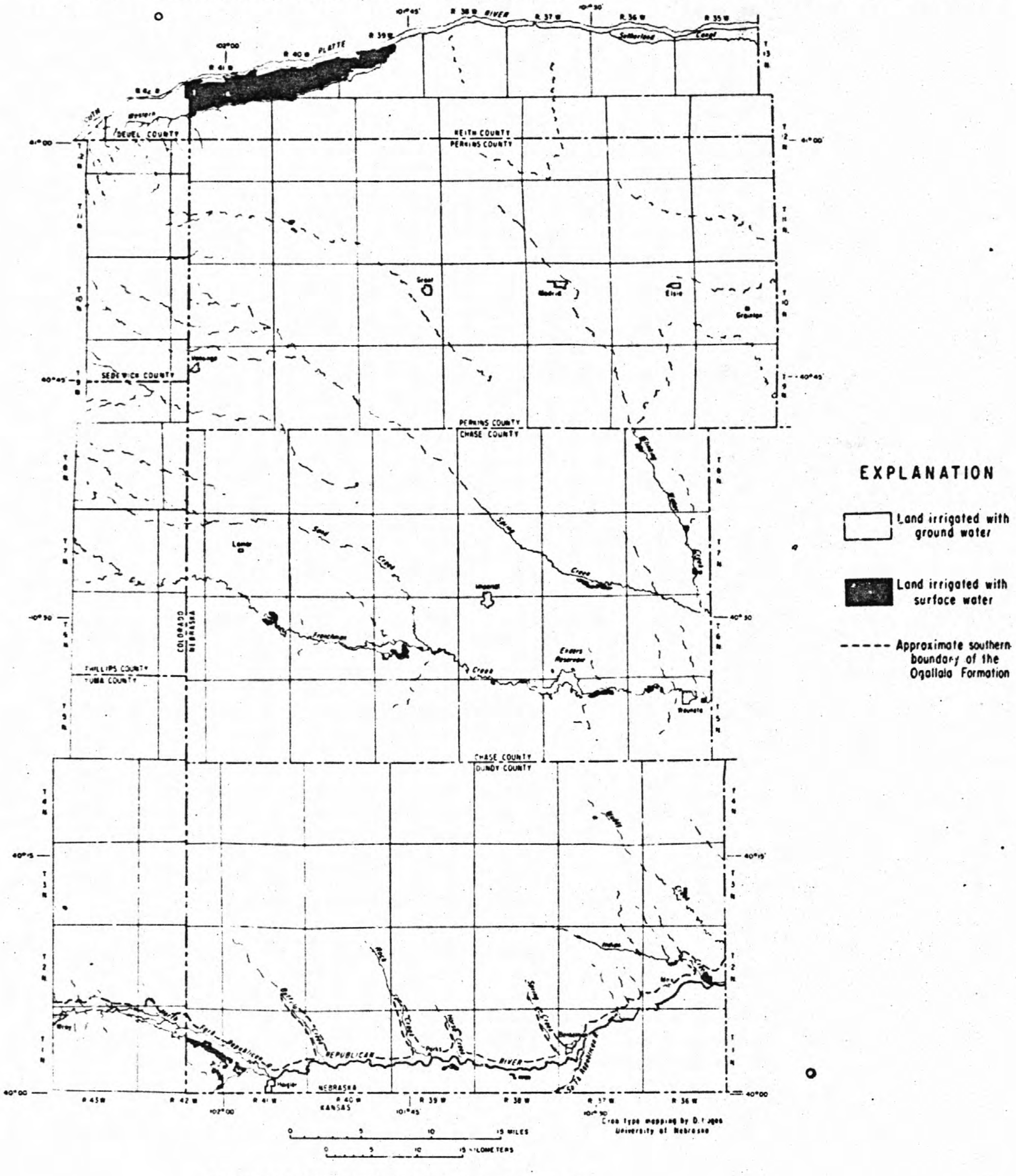






MEAN ANNUAL PRECIPITATION (1925-75)

PLATE 7 - LOCATION OF LAND IRRIGATED WITH GROUND WATER AND SURFACE WATER, MAY 1975



Base compiled from Nebraska Department of Roads County Highway Maps 1961

PLATE 8 - MEASURED AND MODELED HYDROGRAPHS OF SELECTED OBSERVATION WELLS

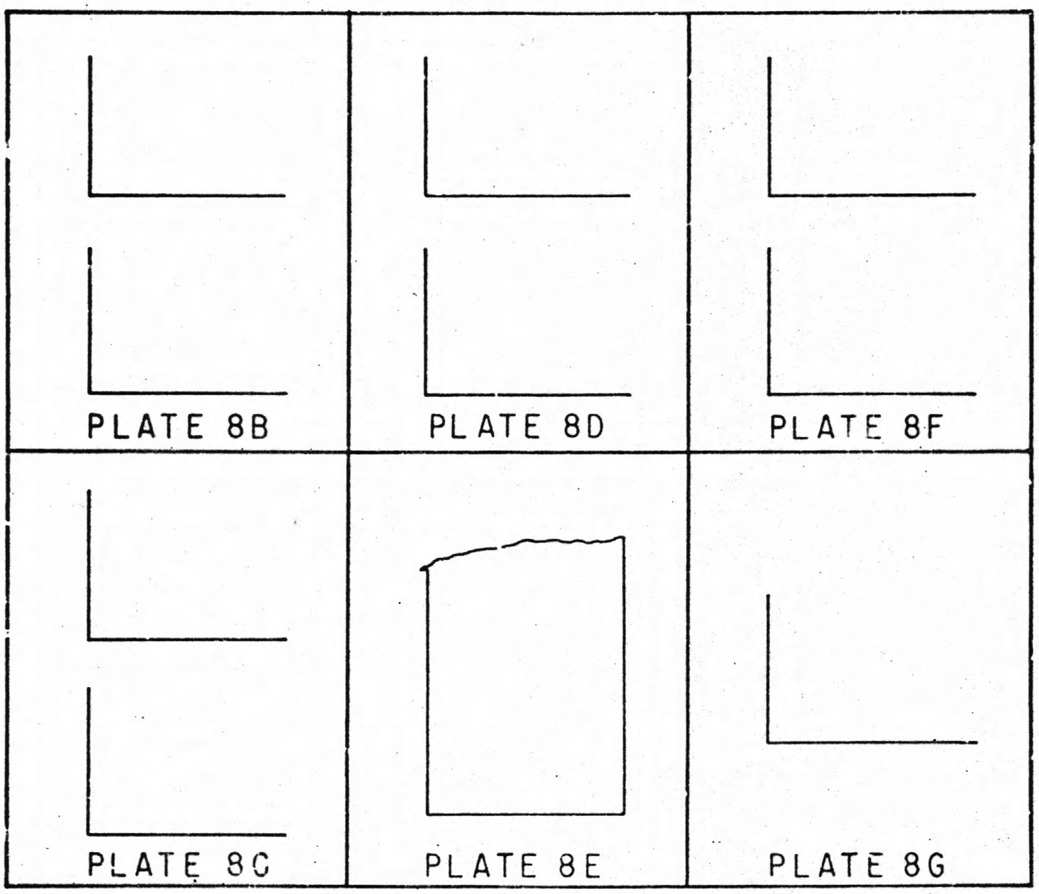


DIAGRAM SHOWING COMPONENTS OF PLATE 8

PLATE 8B - MEASURED AND MODELED HYDROGRAPHS OF SELECTED OBSERVATION WELLS

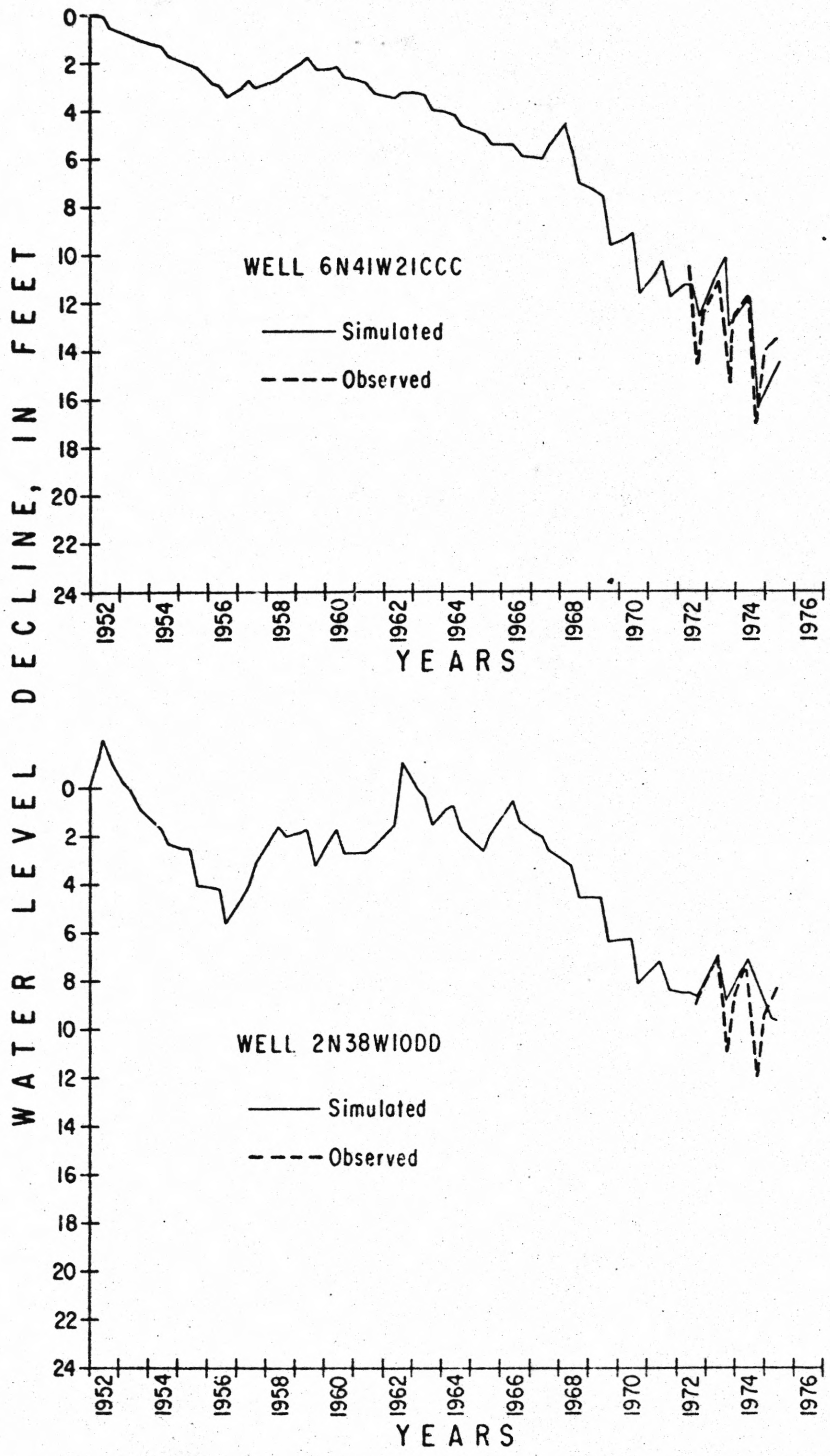
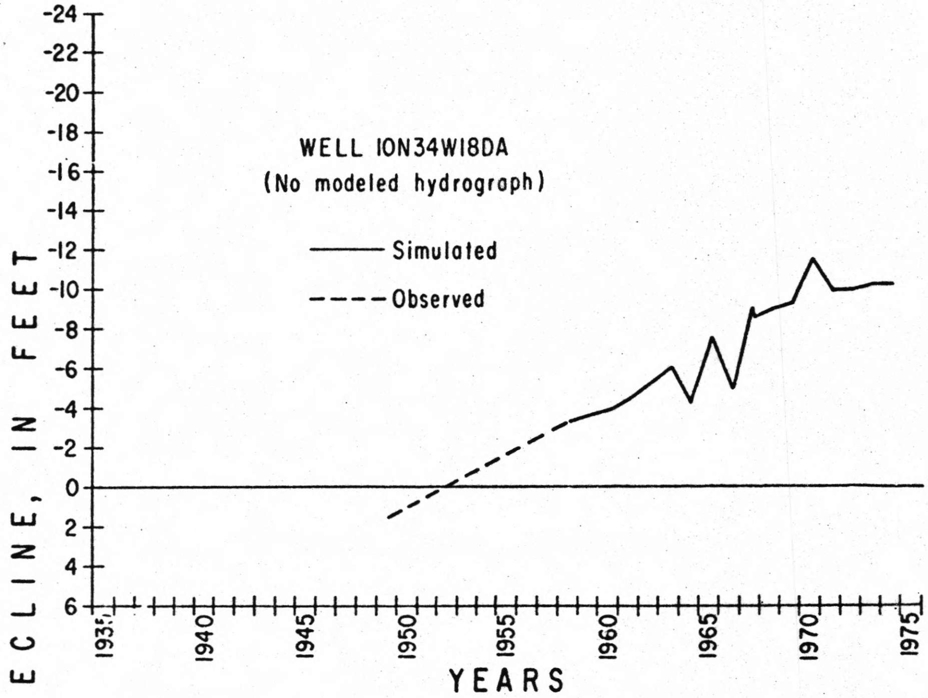


PLATE 8C - MEASURED AND MODELED HYDROGRAPHS OF SELECTED OBSERVATION WELLS



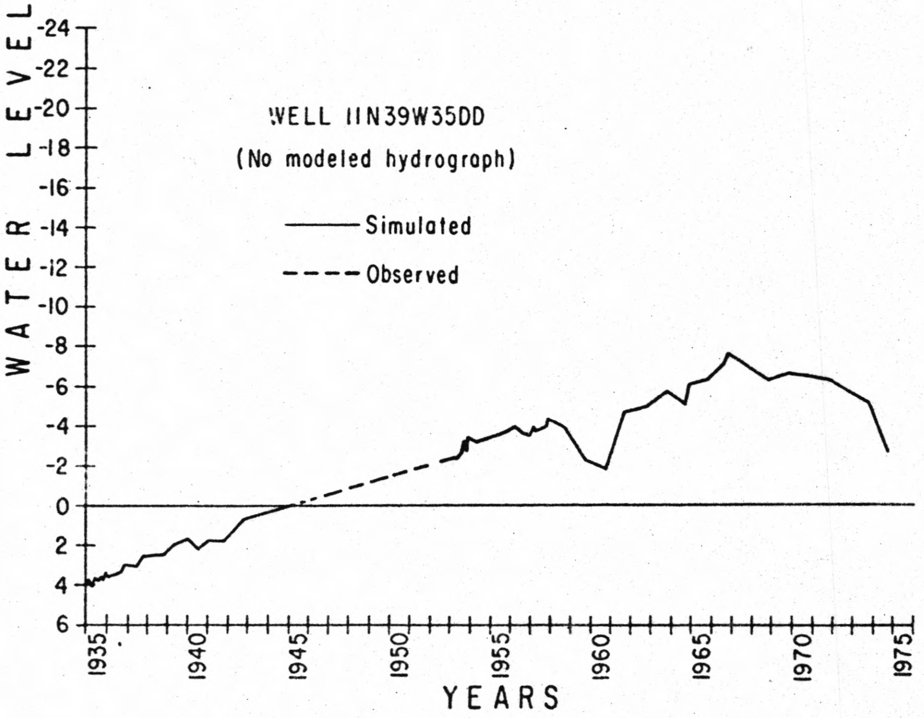
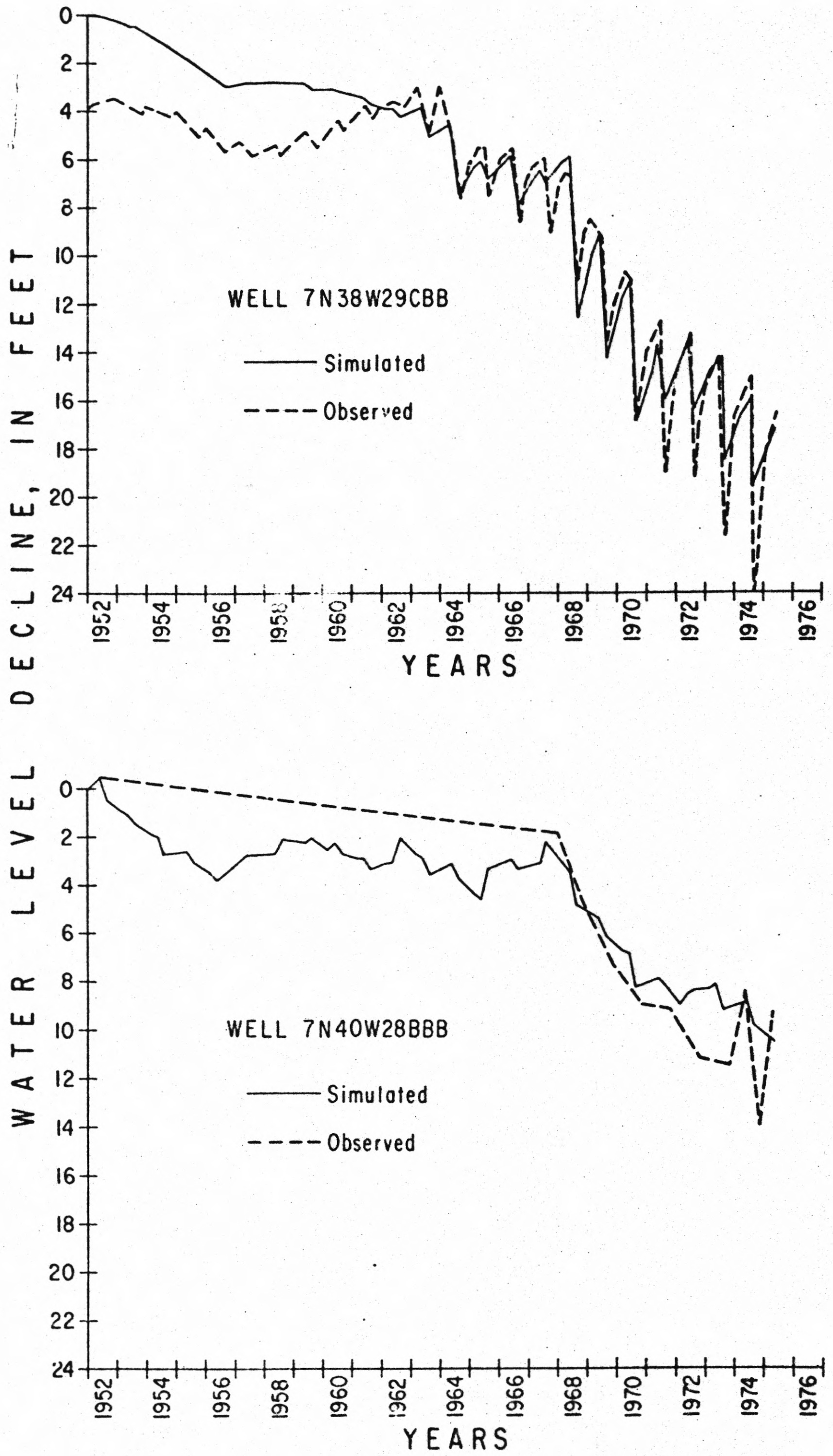


PLATE 8D - MEASURED AND MODELED HYDROGRAPHS OF SELECTED OBSERVATION WELLS



LOCATION OF WELLS FOR WHICH HYDROGRAPHS ARE SHOWN

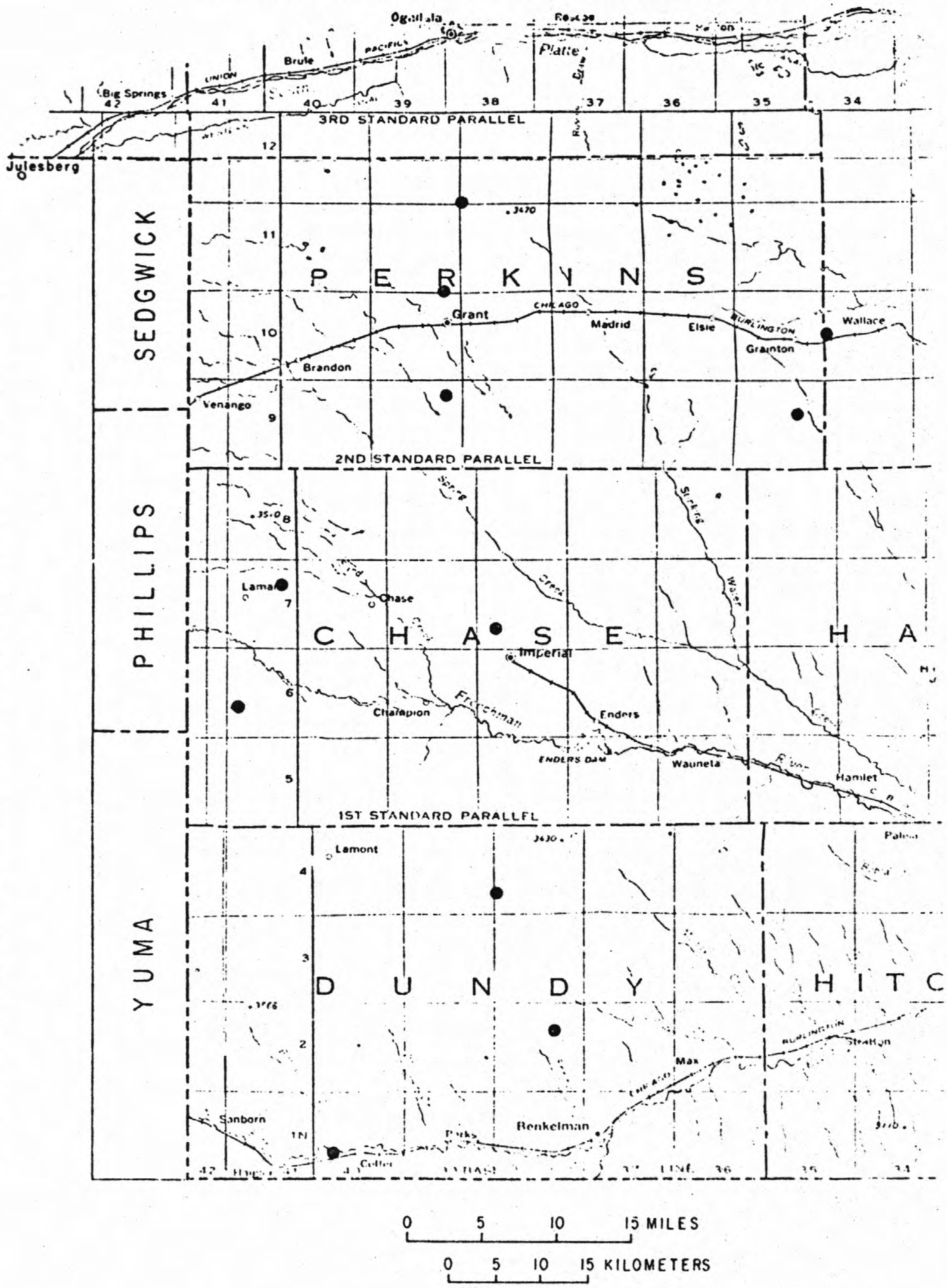


PLATE 8E - MEASURED AND MODELED HYDROGRAPHS OF SELECTED OBSERVATION WELLS

PLATE 8F - MEASURED AND MODELED HYDROGRAPHS OF SELECTED OBSERVATION WELLS

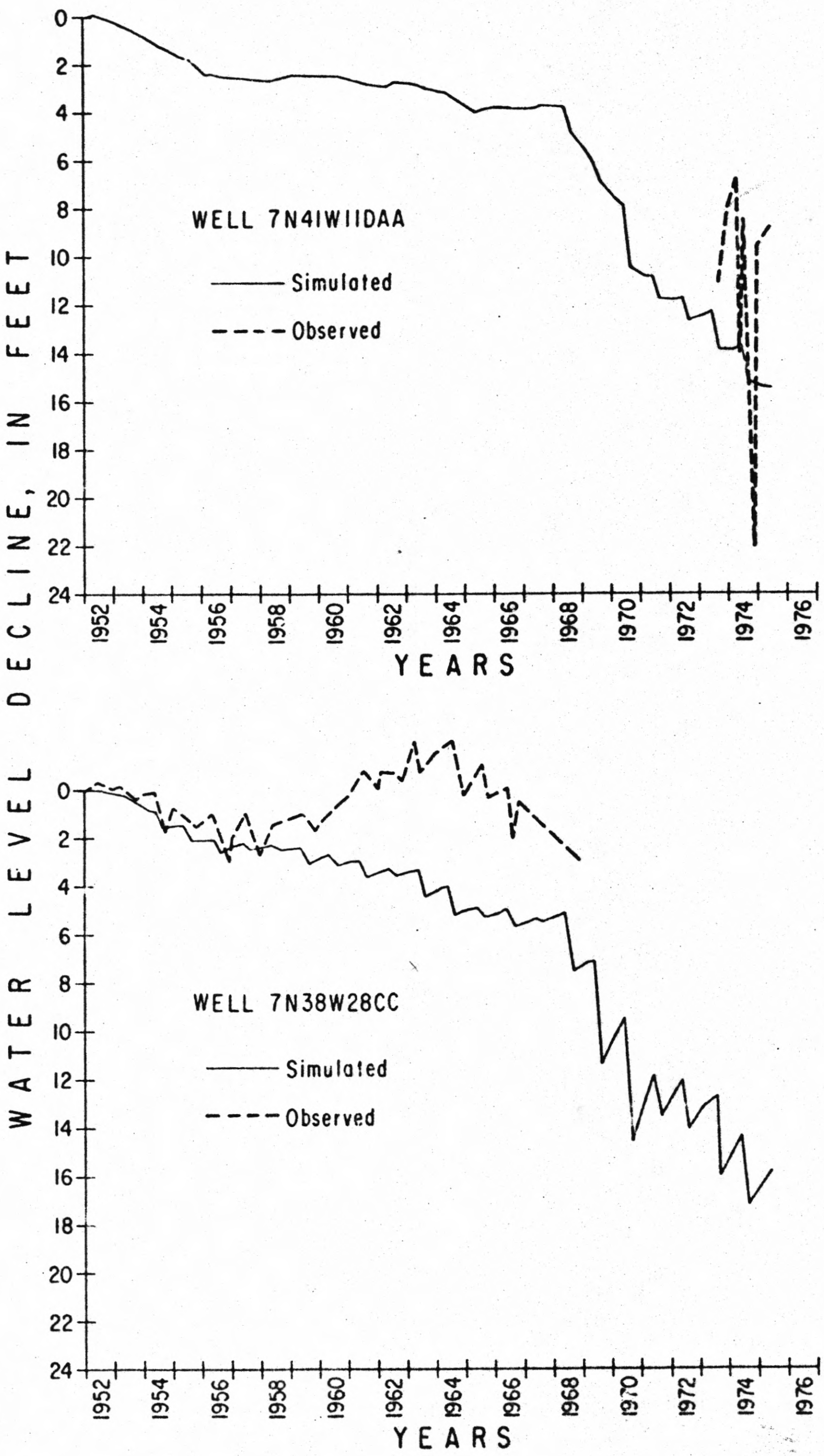


PLATE 8G - MEASURED AND MODELED HYDROGRAPHS OF SELECTED OBSERVATION WELLS

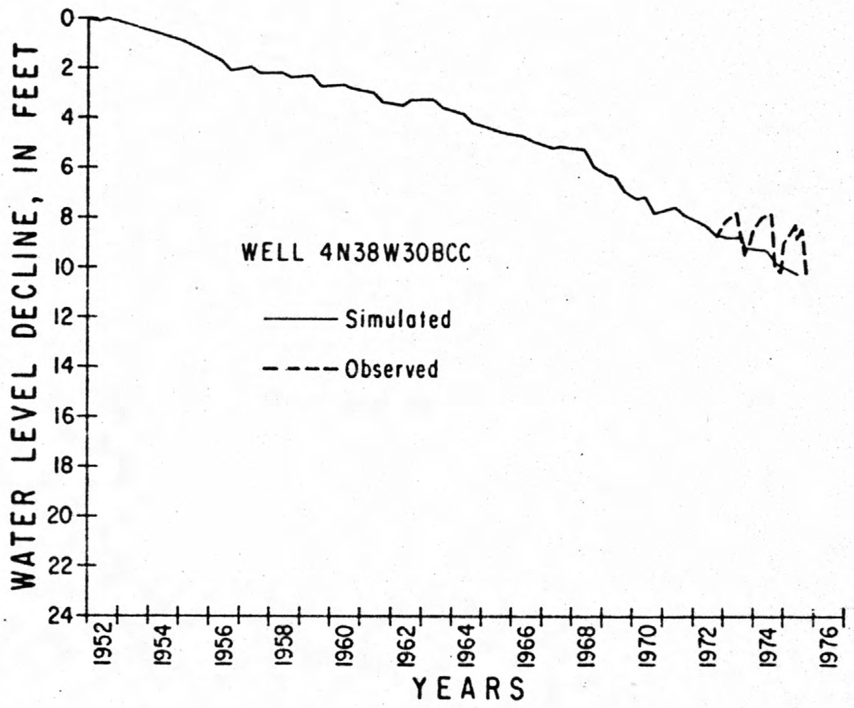


PLATE 9 - PREDICTED WATER-LEVEL DECLINES IN 1980, 1985, 1990, 1995, AND 2000 CAUSED BY WELLS INSTALLED AT THE RATE OF THE LAST 5 YEARS, EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR.

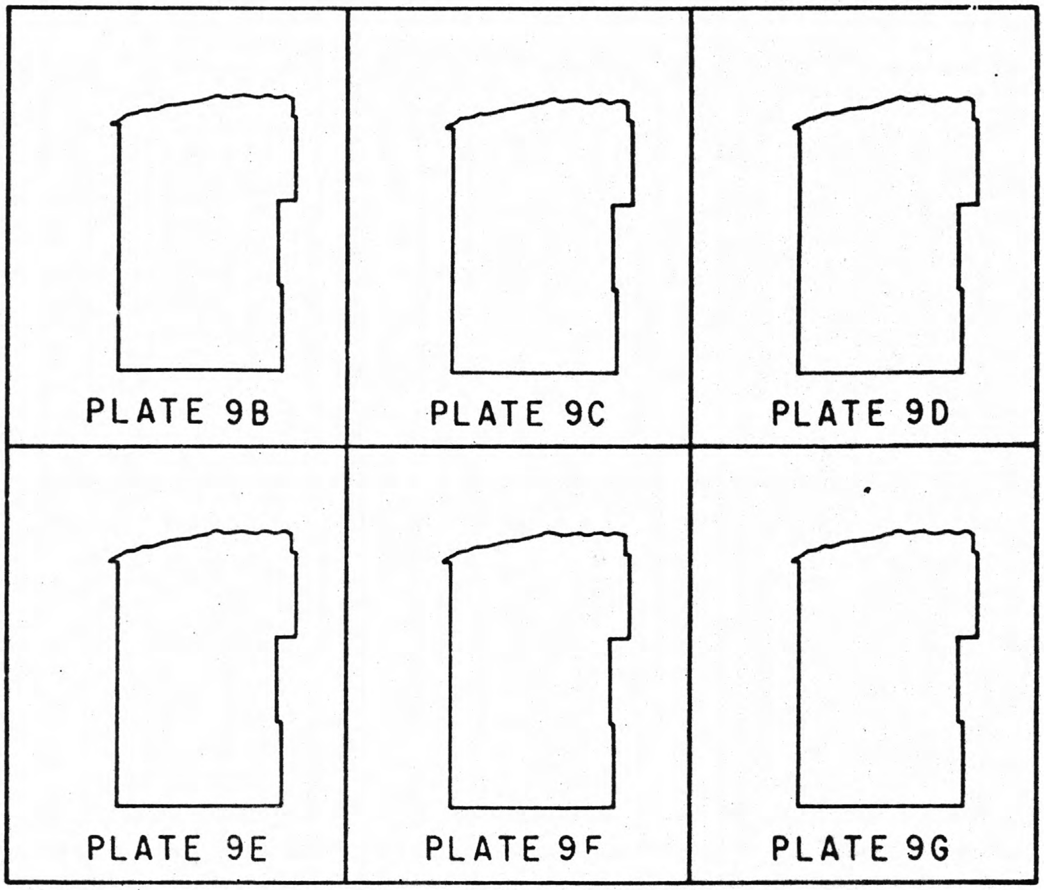


DIAGRAM SHOWING COMPONENTS OF PLATE 9

PLATE 9B - PREDICTED WATER-LEVEL DECLINES IN 1980 CAUSED BY WELLS INSTALLED AT THE RATE OF THE LAST 5 YEARS, EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR.

EXPLANATION

—20— Predicted water-level decline from 1976 to indicated year in feet

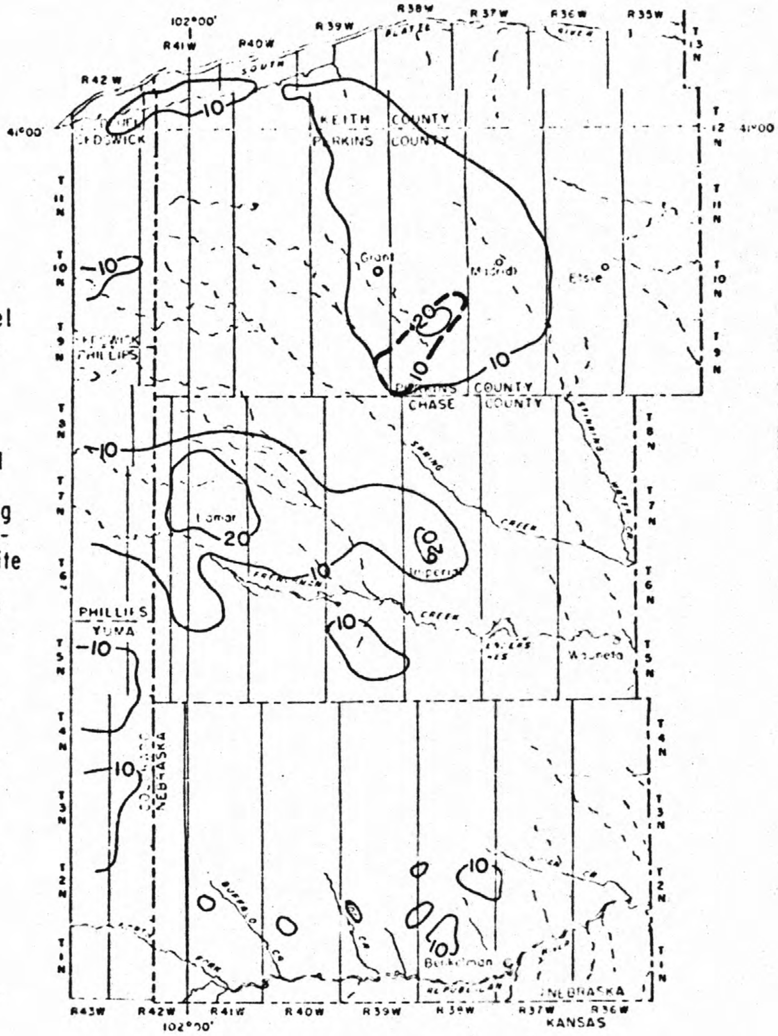
▬▬10▬▬ Predicted water-level decline, 1976 to indicated year, using non-artesian conditions in the indefinite artesian area in Perkins and Keith Counties, in feet

All predicted water levels given as of May 31, each year

Contour intervals are variable

0 5 10 15 MILES

0 5 10 15 KILOMETERS



1976 TO 1980

PLATE 9C - PREDICTED WATER-LEVEL DECLINES IN 1985 CAUSED BY WELLS INSTALLED AT THE RATE OF THE LAST 5 YEARS, EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR.

EXPLANATION

—20— Predicted water-level decline from 1976 to indicated year in feet

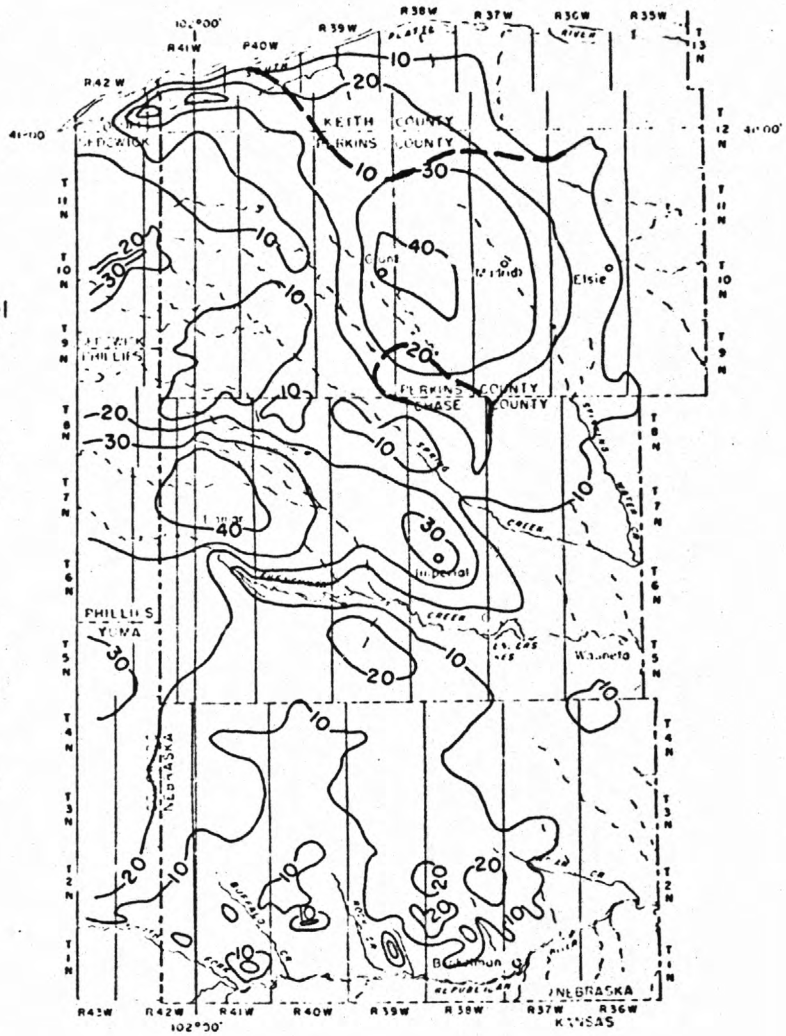
- -10- - Predicted water-level decline, using non-artesian conditions in Perkins and Keith Counties

All predicted water levels given as of May 31, each year

Contour intervals are variable

0 5 10 15 MILES

0 5 10 15 KILOMETERS



1976 TO 1985

PLATE 9D - PREDICTED WATER-LEVEL DECLINES IN 1990 CAUSED BY WELLS INSTALLED AT THE RATE OF THE LAST 5 YEARS, EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR.

EXPLANATION

—20— Predicted water-level decline from 1976 to indicated year in feet

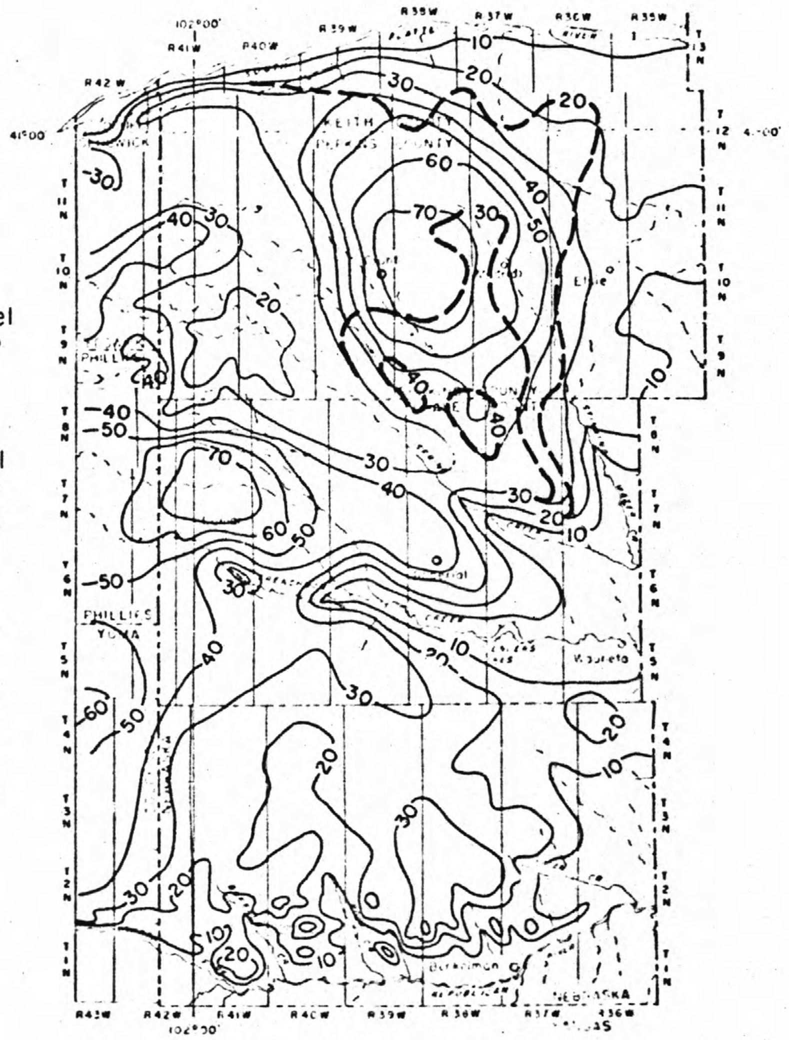
– –10– – Predicted water-level decline, using non-artesian conditions in Perkins and Keith Counties

All predicted water levels given as of May 31, each year

Contour intervals are variable

0 5 10 15 MILES

0 5 10 15 KILOMETERS



1976 TO 1990

PLATE 9E - PREDICTED WATER-LEVEL DECLINES IN 1995 CAUSED BY WELLS INSTALLED AT THE RATE OF THE LAST 5 YEARS, EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR.

EXPLANATION

—20— Predicted water-level decline from 1976 to indicated year in feet

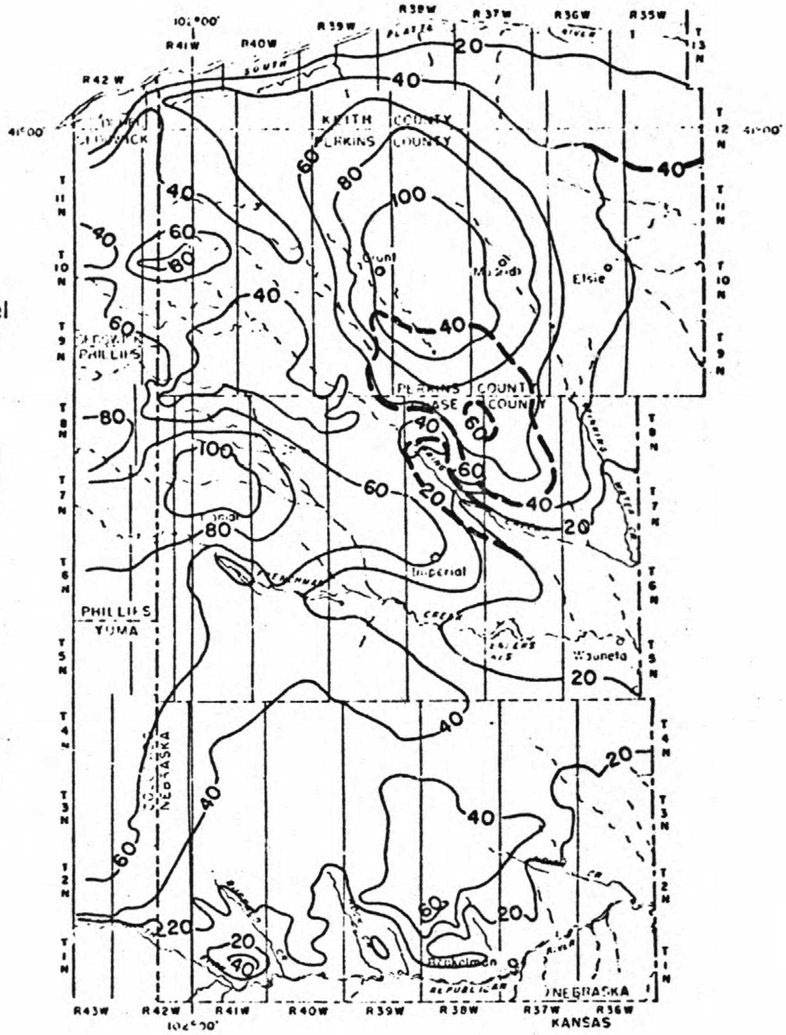
▬▬10▬▬ Predicted water-level decline, using non-artesian conditions in Perkins and Keith Counties

All predicted water levels given as of May 31, each year

Contour intervals are variable

0 5 10 15 MILES

0 5 10 15 KILOMETERS



1976 TO 1995

PLATE 9F - PREDICTED WATER-LEVEL DECLINES IN 2000 CAUSED BY WELLS INSTALLED AT THE RATE OF THE LAST 5 YEARS, EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR.

EXPLANATION

—20— Predicted water-level decline from 1976 to indicated year in feet

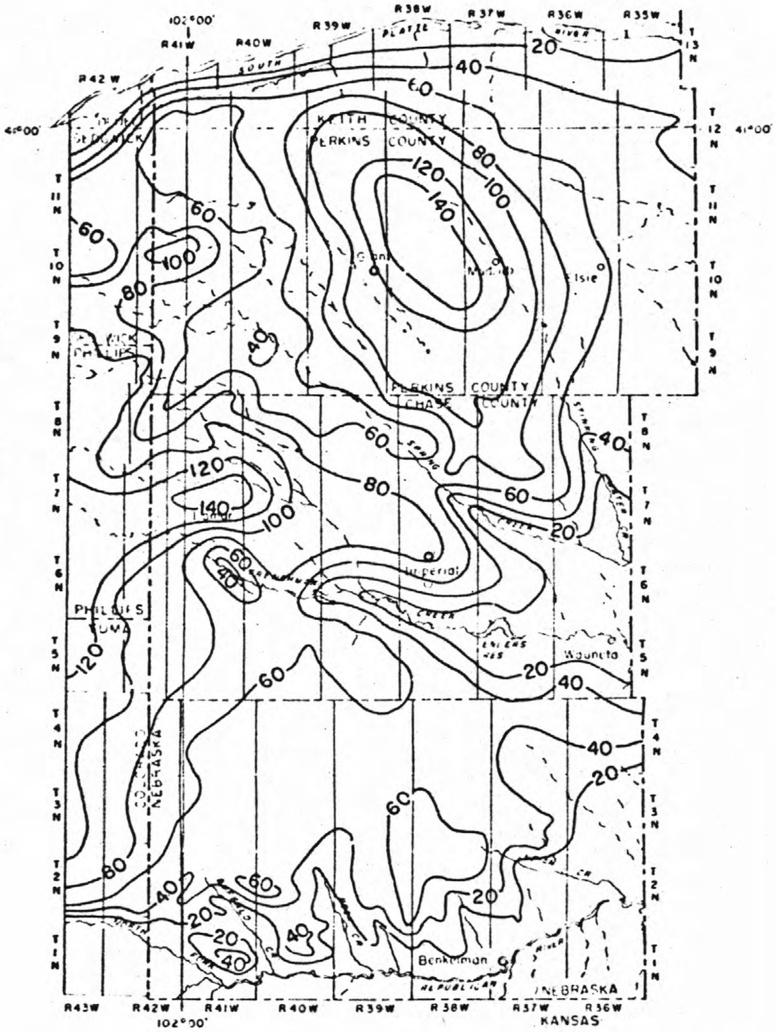
▬ ▬10▬ ▬ Predicted water-level decline, using non-artesian conditions in Perkins and Keith Counties

All predicted water levels given as of May 31, each year

Contour intervals are variable

0 5 10 15 MILES

0 5 10 15 KILOMETERS



1976 TO 2000

PLATE 9G - PREDICTED WATER-LEVEL DECLINES IN 1980, 1985, 1990, and 2000 CAUSED BY WELLS INSTALLED AT THE RATE OF THE LAST 5 YEARS EACH WITHDRAWING 177.3 ACRE-ft EACH YEAR.

NUMBER OF SIMULATED WELLS PER SQUARE MILE, BY 2000

PLATE 10 - PREDICTED WATER-LEVEL DECLINES IN 1980, 1985, 1990, 1995, AND 2000 CAUSED BY WELLS INSTALLED AS OF JANUARY 1976 EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR DURING JUNE THROUGH AUGUST.

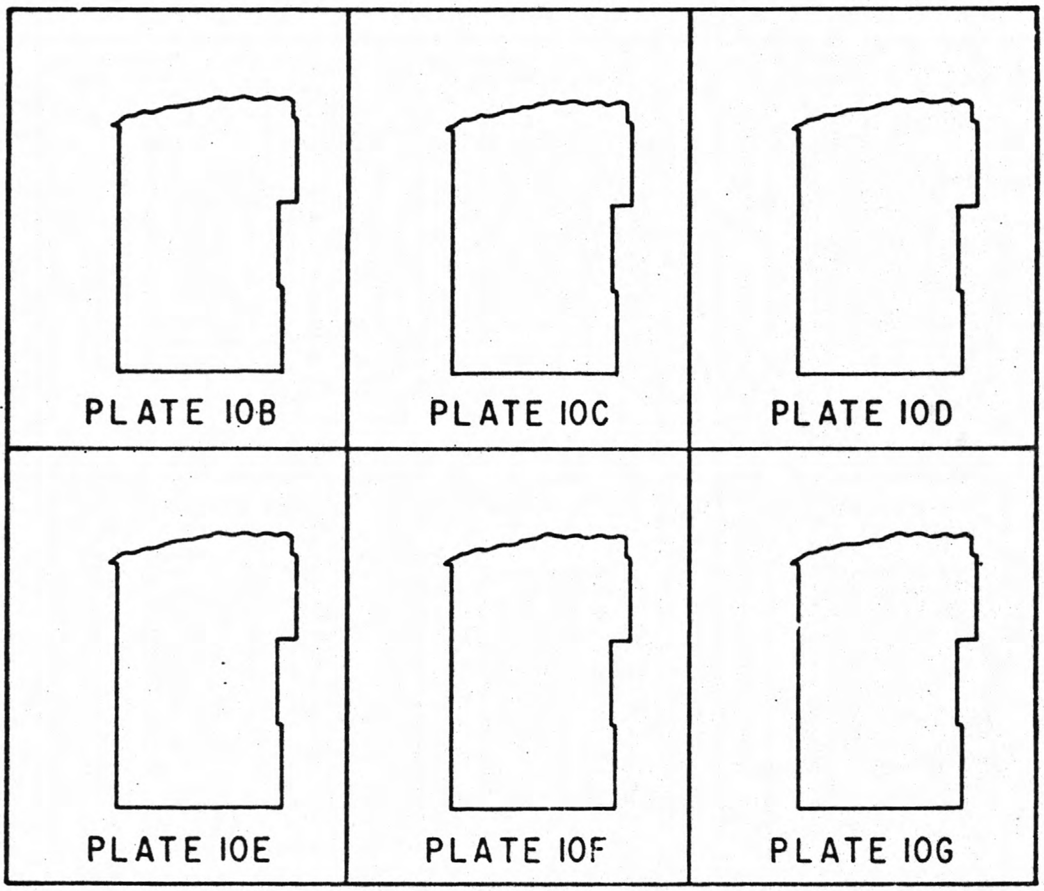


DIAGRAM SHOWING COMPONENTS OF PLATE 10

PLATE 10B - PREDICTED WATER-LEVEL DECLINES IN 1980 CAUSED BY WELLS INSTALLED AS OF JANUARY 1976 EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR DURING JUNE THROUGH AUGUST.

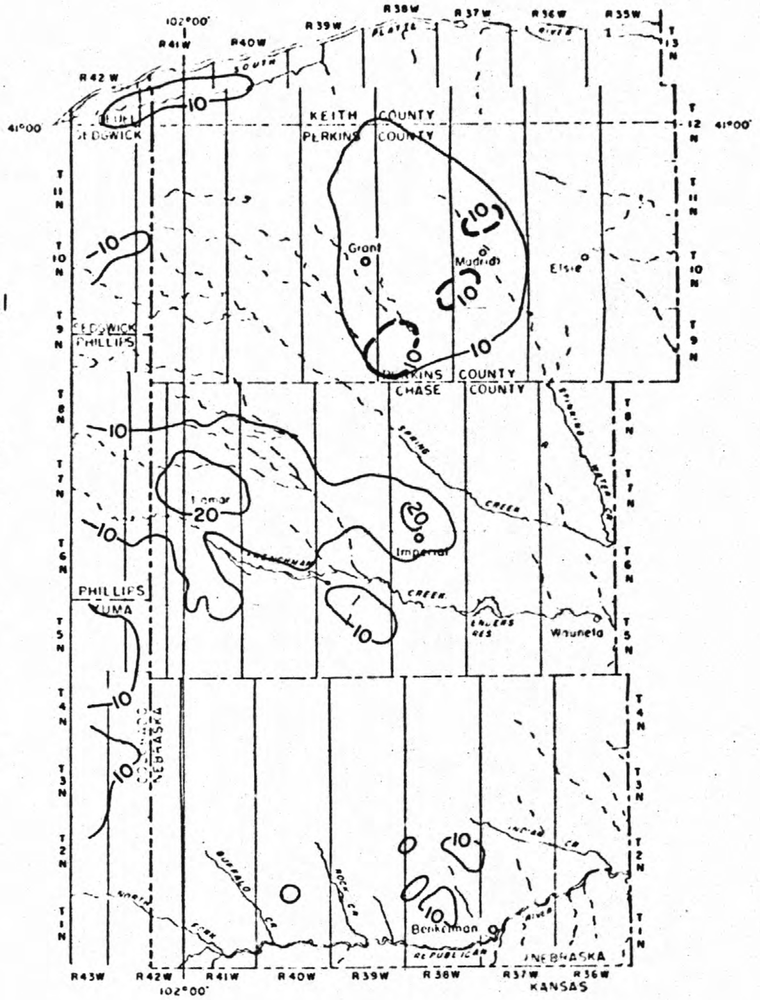
EXPLANATION

—20— Predicted water-level decline from 1976 to indicated year in feet

▬ ▬10▬ ▬ Predicted water-level decline, using non-artesian conditions in Perkins and Keith Counties

All predicted water levels given as of May 31, each year

Contour intervals are variable



1976 TO 1980

PLATE 10C - PREDICTED WATER-LEVEL DECLINES IN 1985 CAUSED BY WELLS INSTALLED AS OF JANUARY 1976 EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR DURING JUNE THROUGH AUGUST.

EXPLANATION

—20— Predicted water-level decline from 1976 to indicated year in feet

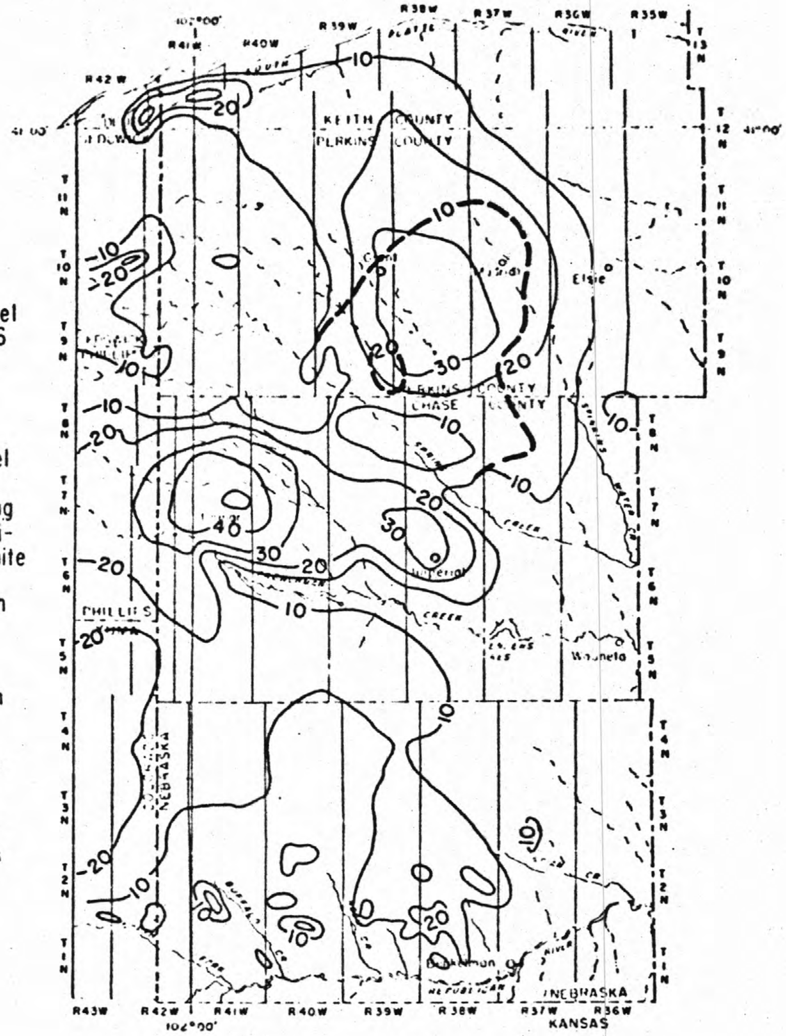
— —10— — Predicted water-level decline, 1976 to indicated year, using non artesian conditions in the indefinite artesian area in Perkins and Keith Counties, in feet

All predicted water levels given as of May 31, each year

Contour intervals are variable

0 5 10 15 MILES

0 5 10 15 KILOMETERS



1976 TO 1985

PLATE 10D - PREDICTED WATER-LEVEL DECLINES IN 1990 CAUSED BY WELLS INSTALLED AS OF JANUARY 1976 EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR DURING JUNE THROUGH AUGUST.

EXPLANATION

—20— Predicted water-level decline from 1976 to indicated year in feet

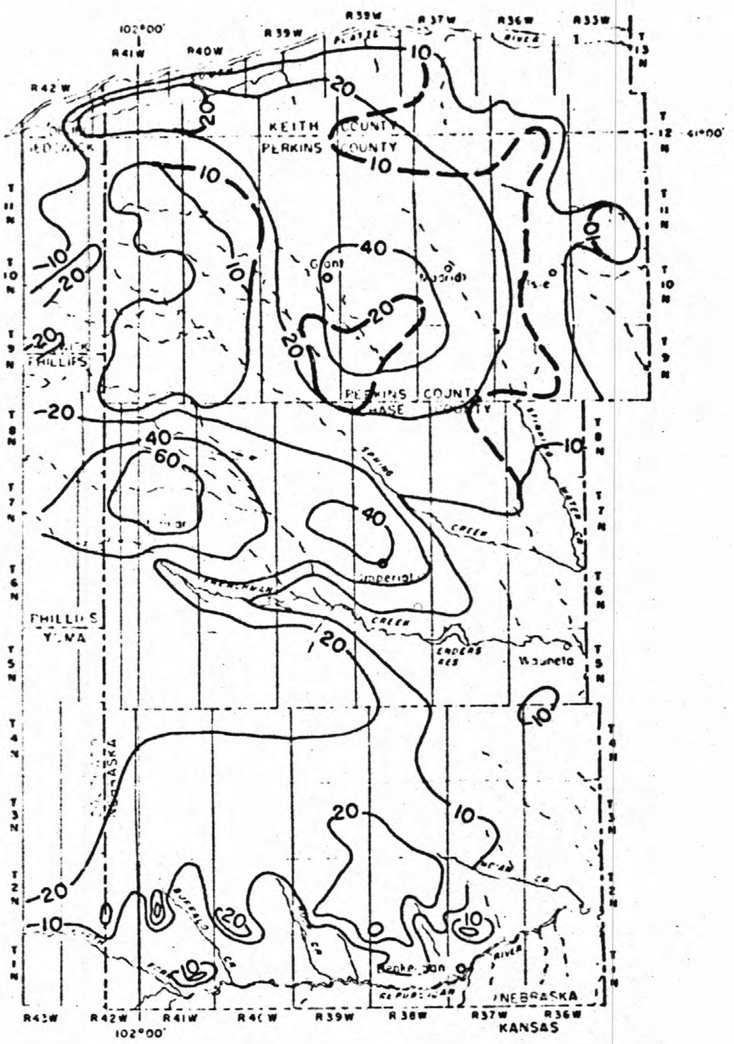
▬▬10▬▬ Predicted water-level decline, 1976 to indicated year, using non-artesian conditions in the indefinite artesian area in Perkins and Keith Counties, in feet

All predicted water levels given as of May 31, each year

Contour intervals are variable

0 5 10 15 MILES

0 5 10 15 KILOMETERS



1976 TO 1990

PLATE 10E - PREDICTED WATER-LEVEL DECLINES in 1995 CAUSED BY WELLS INSTALLED AS OF JANUARY 1976 EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR DURING JUNE THROUGH AUGUST.

EXPLANATION

—20— Predicted water level decline from 1976 to indicated year in feet

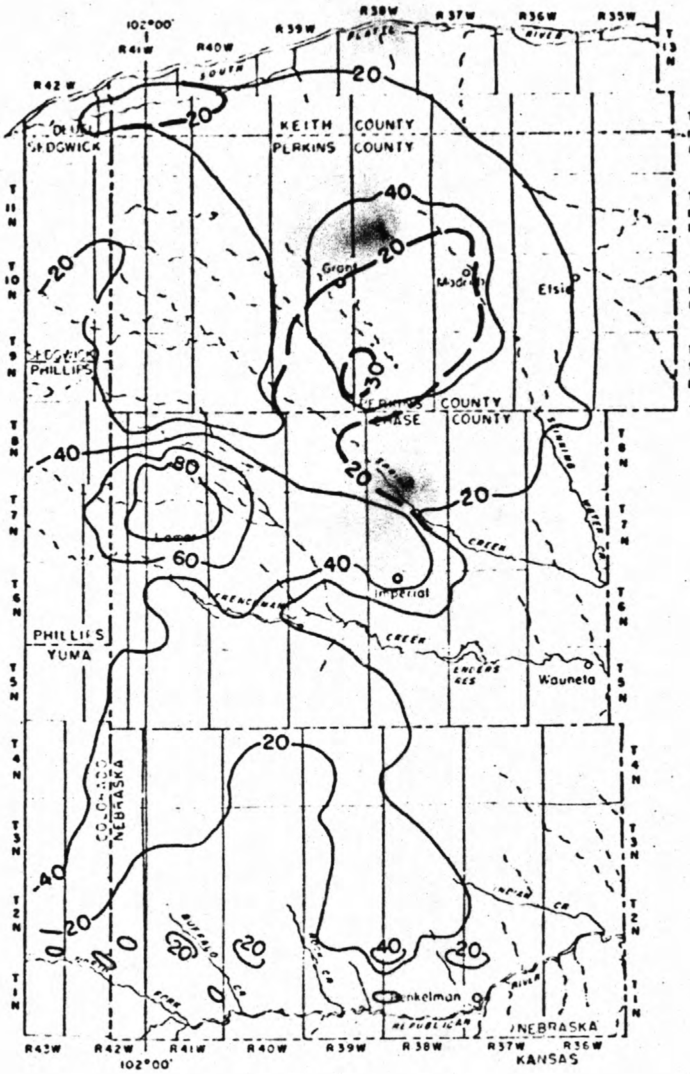
--10-- Predicted water-level decline, 1976 to indicated year, using non-artesian conditions in the indefinite artesian area in Perkins and Keith Counties, in feet

All predicted water levels given as of May 31, each year

Contour intervals are variable

0 5 10 15 MILES

0 5 10 15 KILOMETERS



1976 TO 1995

PLATE 10F - PREDICTED WATER-LEVEL DECLINES IN 2000 CAUSED BY WELLS INSTALLED AS OF JANUARY 1976 EACH WITHDRAWING 177.3 ACRE-FT EACH YEAR DURING JUNE THROUGH AUGUST.

EXPLANATION

—20— Predicted water-level decline from 1976 to indicated year in feet

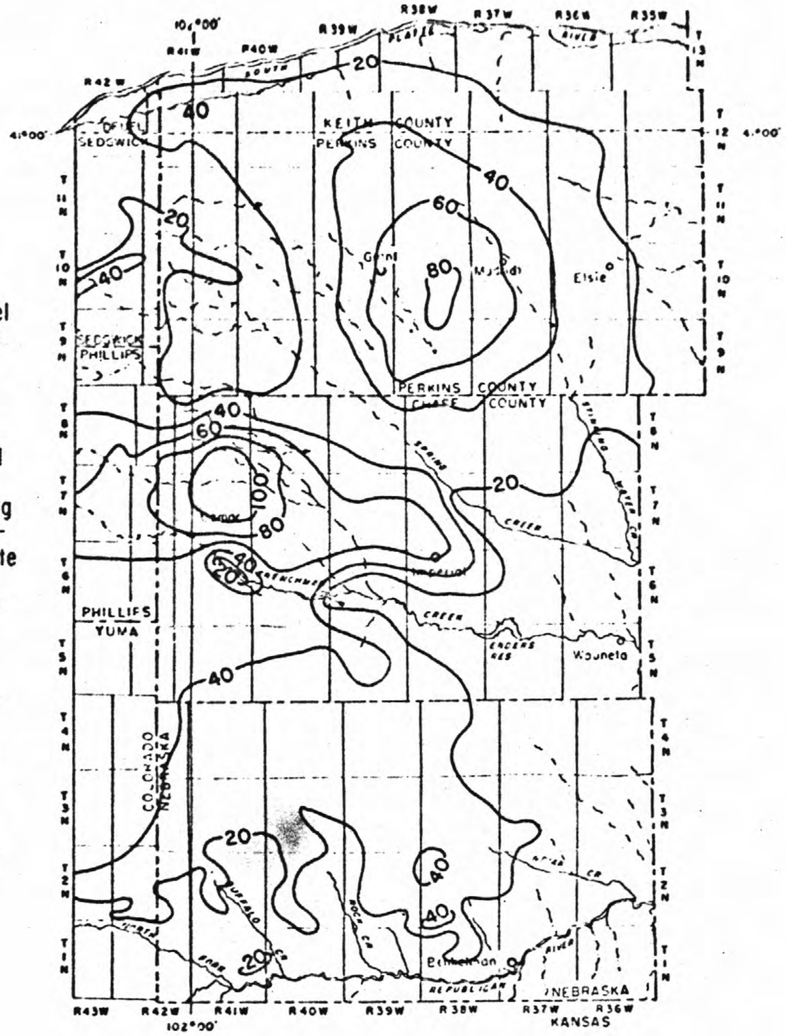
– –10– – Predicted water-level decline, 1976 to indicated year, using non-artesian conditions in the indefinite artesian area in Perkins and Keith Counties, in feet

All predicted water levels given as of May 31, each year

Contour intervals are variable

0 5 10 15 MILES

0 5 10 15 KILOMETERS



1976 TO 2000