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Prepared in cooperation with the

Arkansas Soil and Water Conservation Commission and the
U.S. Army Corps of Engineers, Memphis District

RECALIBRATION OF A GROUND-WATER FLOW MODEL OF THE MISSISSIPPI RIVER VALLEY ALLUVIAL AQUIFER OF NORTHEASTERN ARKANSAS, 1918-1998, WITH SIMULATIONS OF WATER LEVELS CAUSED BY PROJECTED GROUND-WATER WITHDRAWALS THROUGH 2049

Water-Resources Investigations Report 03-4109

U.S. Department of the Interior
U.S. Geological Survey

Front Cover: Photograph of rice fields and well by Paul W. McKee, U.S. Geological Survey.

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Little Rock, Arkansas
2003

U.S. DEPARTMENT OF THE INTERIOR
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Denver, Colorado 80225

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Multiply	By	To obtain
inch (in.)	25.4	millimeter (mm)
inch per year (in/yr)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
foot per mile (ft/mi)	0.3048	meter per kilometer (m/km)
foot per day (ft/d)	0.3048	meter per day (m/d)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)
square foot per day (ft^2/d)	0.09290	square meter per day (m^2/d)
cubic foot per day (ft^3/d)	0.02832	cubic meter per day (m^3/d)
billion gallons per day (Bgal/d)	0.04381	cubic meter per second (m^3/s)

In this report, vertical coordinate information is referenced to the National Geodetic Vertical Datum of 1929 (NGVD of 1929). Horizontal coordinate information is referenced to North American Datum of 1983 (NAD 83).

RECALIBRATION OF A GROUND-WATER FLOW MODEL OF THE MISSISSIPPI RIVER VALLEY ALLUVIAL AQUIFER OF NORTHEASTERN ARKANSAS, 1918-1998, WITH SIMULATIONS OF WATER LEVELS CAUSED BY PROJECTED GROUND-WATER WITHDRAWALS THROUGH 2049

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ABSTRACT

A digital model of the Mississippi River Valley alluvial aquifer in eastern Arkansas was used to simulate ground-water flow for the period from 1918 to 2049. The model results were used to evaluate effects on water levels caused by demand for ground water from the alluvial aquifer, which has increased steadily for the last 40 years. The model results showed that water currently (1998) is being withdrawn from the aquifer at rates greater than what can be sustained for the long term. The saturated thickness of the alluvial aquifer has been reduced in some areas resulting in dry wells, degraded water quality, decreased water availability, increased pumping costs, and lower well yields.

The model simulated the aquifer from a line just north of the Arkansas-Missouri border to south of the Arkansas River and on the east from the Mississippi River westward to the less permeable geologic units of Paleozoic age. The model consists of 2 layers, a grid of 184 rows by 156 columns, and comprises 14,118 active cells each measuring 1 mile on a side. It simulates time periods from 1918 to 1998 along with further time periods to 2049 testing different pumping scenarios. Model flux boundary conditions were specified for rivers, general head boundaries along parts of the western side of the model and parts of Crowleys Ridge, and a specified head boundary across the aquifer further north in Missouri.

Model calibration was conducted for observed water levels for the years 1972, 1982, 1992, and 1998. The average absolute residual was 4.69 feet and the root-mean square error was 6.04 feet for the hydraulic head observations for 1998.

Hydraulic-conductivity values obtained during the calibration process were 230 feet per day for the upper layer and ranged from 230 to 730 feet per day for the lower layer with the maximum mean for the combined aquifer of 480 feet per day. Specific yield values were 0.30 throughout the model and specific storage values were 0.000001 inverse-feet throughout the model. Areally specified recharge rates ranged from 0 to about 30 inches and total recharge increased from 1972 to 1998 by a factor of about four.

Water levels caused by projected ground-water withdrawals were simulated using the calibrated model. Simulations represented a period of 50 years into the future in three scenarios with either unchanged pumpage, pumpage increased by historic trends, or pumpage increased by historic trends except in two areas of the Grand Prairie. If pumping remains at 1997 rates, this produces extreme water-level declines (areas where model cells have gone dry or where the water level in the aquifer is equal to or less than the original saturated thickness, assuming confined conditions in the aquifer everywhere in the formation in predevelopment times) in the aquifer in two areas of the aquifer (one in the Grand Prairie area between the Arkansas and White Rivers and the other west of Crowleys Ridge along the Cache River) with about 400 square miles going dry. Increasing the pumping rates to that which would be projected using historic data led to increased extreme water-level declines in both areas with about 1,300 square miles going dry. Declines in both scenarios generally occurred most rapidly between 2009 and 2019. Reducing the pumping rates to 90 percent of that used for projected historic rates in areas between the Arkansas and White Rivers relating to two diversion projects of the U.S. Army Corps of Engineers and other agencies did little to decrease the extreme water-level declines. However, these pumpage reductions are small (amounting to about 16 percent of the reductions that

could result from implementation of these diversion projects).

INTRODUCTION

The Mississippi River Valley alluvial aquifer (hereinafter referred to as the alluvial aquifer), located in eastern Arkansas, and parts of Missouri, Kentucky, Tennessee, Mississippi, and Louisiana, also locally called the Delta, encompasses an area of about 32,000 mi^2 and supplies an average of 4.4 Bgal/d of water (1994 data). Historically, the aquifer has been an important water resource driving agriculture, business, and community growth in eastern Arkansas by providing abundant water of high quality. However, in recent years demand has outstripped recharge of the aquifer, and water users and water-use planners are questioning the ability of the aquifer to meet increasing water demands in the long term. Withdrawals from the aquifer have caused considerable declines in aquifer water levels. The effects of current (1998) ground-water withdrawals and potential future withdrawals on water availability are major concerns of water managers and users as well as the general public. A full understanding of the behavior of the aquifer under various water-use scenarios is critical to development of viable water-management and alternative source plans.

The alluvial aquifer is the most prolific aquifer supplying water to the Grand Prairie area and other parts of the Delta of eastern Arkansas. Aquifer withdrawal rates probably are much greater than recharge; and furthermore, they cannot be sustained for the long term. Massive cones of depression in aquifer water levels have formed in numerous areas and continue to grow in area and depth. For example, in some areas the saturated thickness of the alluvial aquifer has been reduced to less than 20 ft from an original total estimated saturated thickness greater than 200 ft. The consequences of excessive pumping fo the aquifer are serious: dry wells, decreasing water quality, decreasing water availability, increased pumping costs, and damage to the ability of the aquifer to yield water.

Surface water is abundant in the region at certain times during the year, and the conjunctive use of surface water with ground water is a key element to assuring sustainable use of the alluvial aquifer and future growth in the region. However, surface water is not readily or economically available in all areas, and surface-water withdrawals are subject to limitation by mandated minimum streamflows. Surface-water availability also is affected by intense pumping of regional aquifers. Surface- and ground-water systems in the region are intimately connected; great volumes of water are transferred between surface-water bodies and the alluvial aquifer. In the early 1900’s, predevelopment conditions existed and most rivers in the region received part of their flow from ground water, especially during dry summer months. This transfer of water has reversed as development of the alluvial aquifer has increased and water levels have declined. Many rivers now lose water to the aquifer, and minimum observed streamflows in unregulated streams have decreased, especially during the summer months. Ground-water gradients tend to show flow from the rivers to the aquifer during most of the year, except for the Mississippi River.

To address these concerns, the U.S. Geological Survey (USGS), in cooperation with the Arkansas Soil and Water Conservation Commission and the U.S. Army Corps of Engineers, Memphis District, conducted an investigation using a digital ground-water flow model.

Purpose And Scope

The purpose of this report is to evaluate the regional effects on water levels in the alluvial aquifer located north of the Arkansas River caused by projected ground-water withdrawals. An existing ground-water flow model (Mahon and Poynter, 1993), discretized at a uniform 1 mi by 1 mi cell size was recalibrated using more recently developed modeling software. Additional water-use and observation well data from succeeding years provided an opportunity to verify or, as proved necessary, recalibrate this model.

This report describes the hydrogeologic setting, model design, calibration procedures, and results of simulations using the calibrated model on water levels caused by projected ground-water withdrawals. Hypothetical pumpage for 50 years into the future (through 2049) was computed and effects on water levels caused by the projected pumpage were simulated using three different pumpage scenarios.

In addition to the interpretations, tables, and figures in the printed report, an accompanying compact disk contains a pdf file of this report as well as a set of digital animations of the three scenarios simulating different rates of hypothetical pumpage through 2049. These animations show changes in water levels resulting from model simulations associated with each of the projected scenarios.

Study Area Description

The area studied and modeled includes all or parts of 23 counties north of the Arkansas River in Arkansas and all or part of 5 counties in southeastern Missouri (fig. 1). The modeled area included the area of the Mississippi River Valley alluvial aquifer north of the Arkansas River, west of the Mississippi River, east of the consolidated formations of Paleozoic age, and an area north of the Arkansas-Missouri border, encompassing a small part of southeastern Missouri.

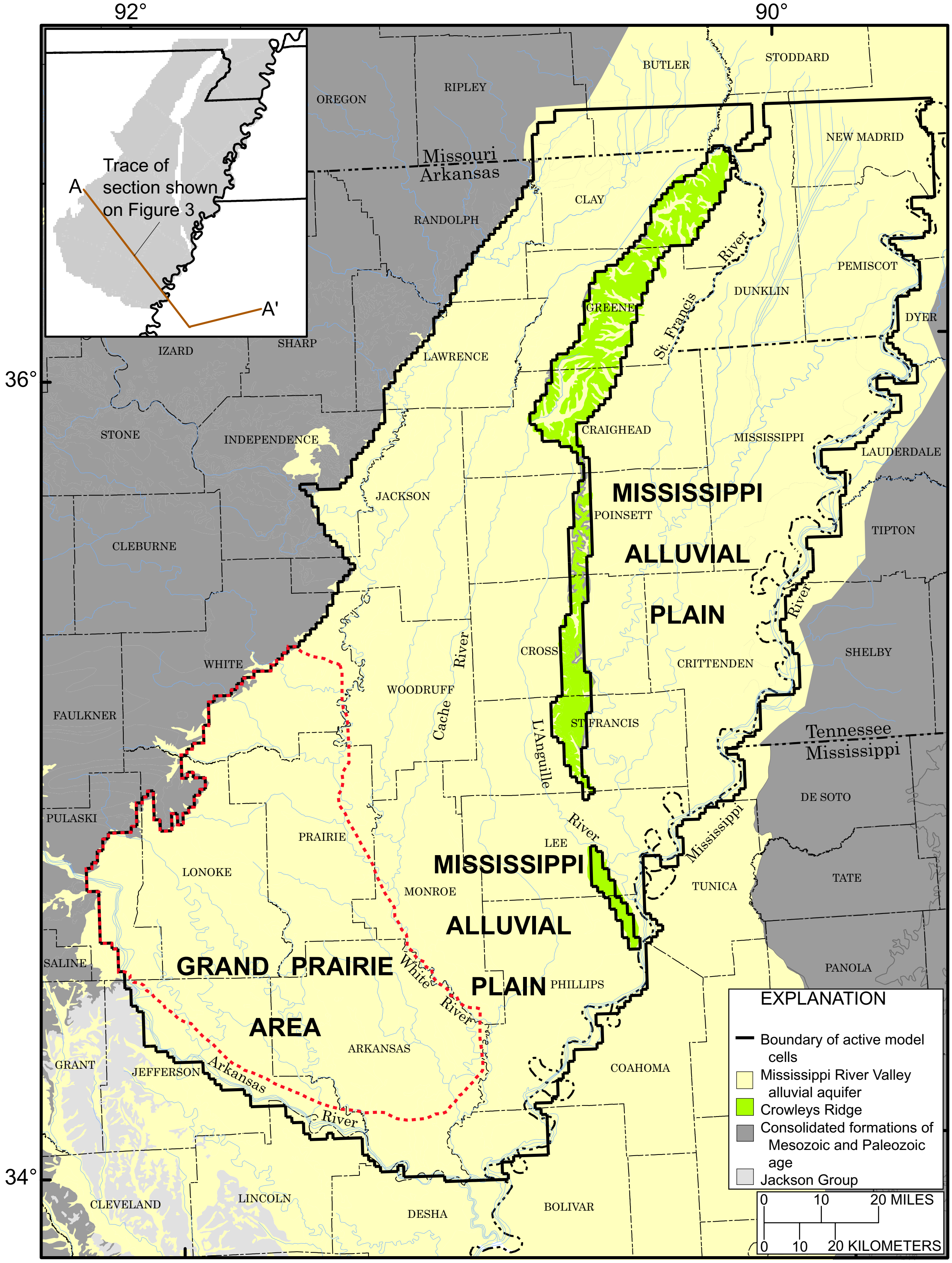
Previous Studies

Many investigators have described the underlying sediments of the Mississippi Alluvial Plain. One of the earliest reports describing subsurface geology and ground-water resources in southern Arkansas and northern Louisiana was written by Veatch (1906). Ground-water resources of northeastern Arkansas were described and a detailed inventory was provided by Stephenson and Crider (1916). Fisk (1944) reported on extensive geologic investigations along the Mississippi River Valley made by the U.S. Army Corps of Engineers between 1941 and 1944. This compilation consists of text accompanied by more than 110 illustrations describing the alluvial sediments. Counts and Engler (1954) reported on changes in water levels in the alluvial aquifer showing as early as 1938 a large cone of depression up to 60 ft deep in the Grand Prairie area covering Arkansas County (Counts and Engler, 1954, fig. 1). Krinitzsky and Wire (1964) expanded on the hydrogeologic work of Fisk with a comprehensive look at ground-water conditions. Cushing and others (1964) and Boswell and others (1968) provided an overview of the alluvial aquifer in their discussions of Quaternary-age aquifers on the Mississippi Embayment. Boswell and others (1968) first referred to the water-yielding sediments underlying the alluvial plain as the Mississippi River Valley alluvial aquifer. Joseph (1999) provided a contour map of water levels for the aquifer for measurements made in the spring of 1998 as did Schrader (2001) for measurements made in the spring of 2000.

Several reports have been published documenting the results of model simulations of the flow system within and across boundaries of the alluvial aquifer. Broom and Lyford (1981) described the flow system of the alluvial aquifer based on the results of a model of the Cache and St. Francis River Basins in northeastern Arkansas. Peralta and others (1985) estimated future Quaternary-age (alluvial aquifer) ground-water availability in the Grand Prairie area by using a flow model coupled to an optimization routine. Regional model investigations were conducted by Ackerman (1989a, 1989b, 1990) under the framework of the USGS Gulf Coast Regional Aquifer-System Analysis; these reports describe the model development and results, and show the characteristics of the flow system on a regional basis. Predictive simulations presented by Ackerman (1989a) were based on hypothetical increases in pumping.

Previous Model

Substantial water-level declines in the alluvial aquifer prompted the need to better understand the flow system in the alluvial aquifer, which led to the development of digital ground-water flow models of the alluvial aquifer. Models developed previously were either at a scale that was too large to analyze the effects of projected pumpage or they were limited in their areal extent. The model discussed in this report is based on the previous model of the alluvial aquifer documented by Mahon and Poynter (1993). Their report included model development, calibration, and results of two separate models for eastern Arkansas: one for the area north of the Arkansas River and one for the area south of the Arkansas River. The model utilized the MODFLOW finite difference numerical-modeling software (McDonald and Harbaugh, 1988) to simulate transient conditions, and simulated time from 1918 to 1987 in seven stress periods. Model grids were based on a cell size of 1 mi^2 throughout the study areas. The models consisted of one layer with no-flow boundaries below that layer. Recharge was simulated as entering the aquifer from head-dependent surface infiltration through the overlying confining unit and through seepage through underlying units and riverbeds using the river package of MODFLOW. The active model cells of the Mahon and Poynter (1993) north model encompassed the same area as the active cells calibrated for the model described in this report (fig. 2).



Base from U.S. Geological Survey digital data, 1:100,000 and 1:2,000,000
Digital geological data adapted from Haley and others (1976) and Bicker (1969)

Figure 1. Location of study and modeled area.

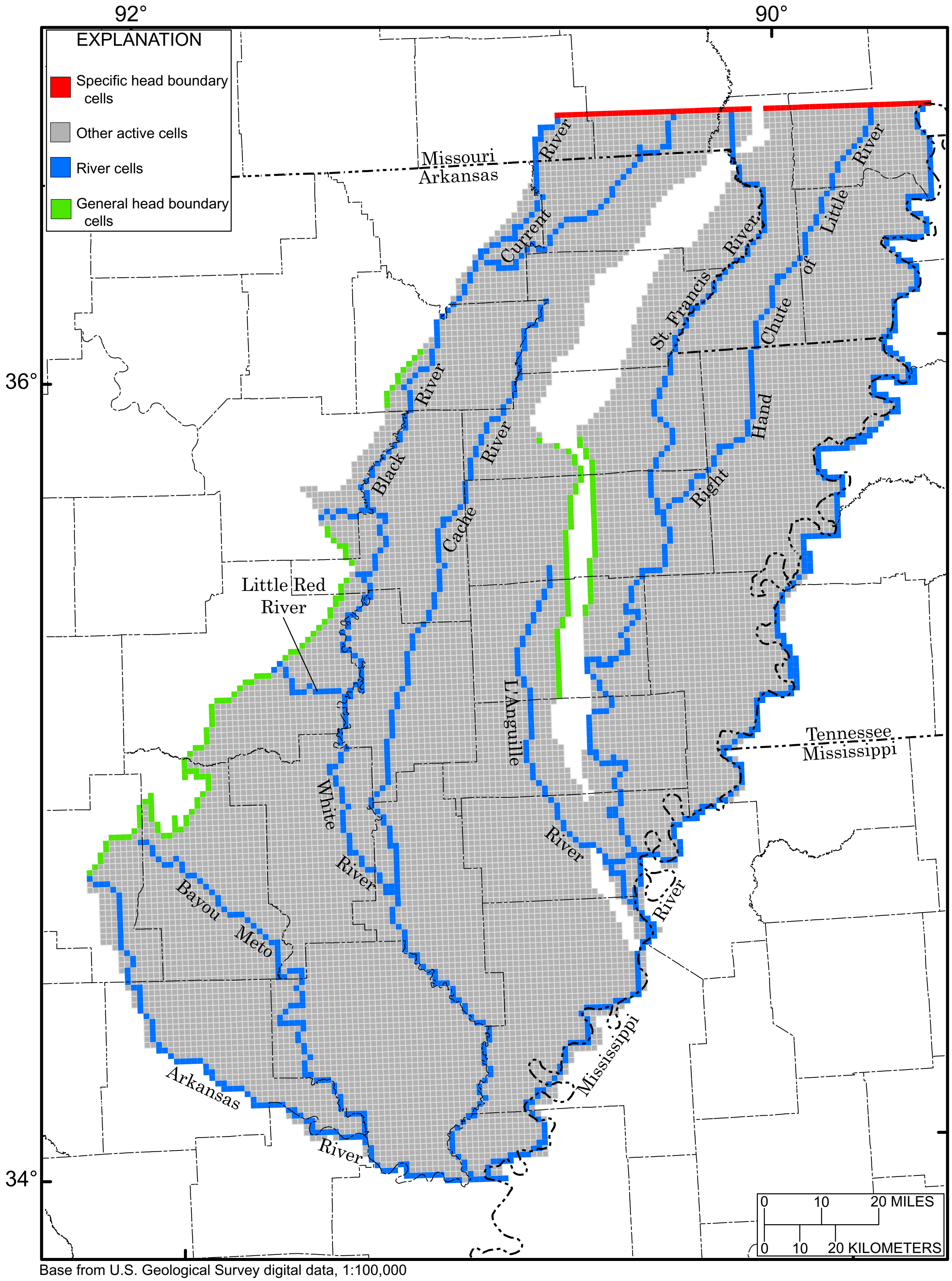


Figure 2. Finite-difference grid of model cells.

HYDROLOGIC SETTING

The alluvial aquifer, the uppermost aquifer system in eastern Arkansas, is part of a much larger sedimentary system known as the Mississippi Embayment. The Mississippi embayment extends southward in a fan shaped geosyncline, plunging southward from southern Illinois to the Gulf of Mexico, and covers about 160,000 mi^2 in parts of Alabama, Arkansas, Illinois, Kentucky, Louisiana, Mississippi, Missouri, and Tennessee (Cushing and others, 1964; Williamson and others, 1990). The ages of the embayment sediments range from Jurassic to Quaternary, but only units of Cretaceous age and younger crop out in Arkansas. The central axis of the Mississippi Embayment nearly parallels the Mississippi River, and the embayment surface drainage in Arkansas is ultimately to the Mississippi River.

The Mississippi Alluvial Plain is a broad, flat plain that lies within the Coastal Plain physiographic province (Fenneman, 1938) and is part of the Mississippi Embayment. The alluvial plain in Arkansas is bounded on the west by consolidated formations of Paleozoic age with very low hydraulic conductivity and by sediments of Tertiary age of the Mississippi Embayment that have a distinctly lower hydraulic conductivity than sediments of the alluvial aquifer (Ackerman, 1990).

Deposition of sediment from the Mississippi and Arkansas Rivers during Pleistocene and Holocene time has produced deposits of alluvium (fig. 3) consisting of a sequence of sands, silts, and clays that constitute the alluvial aquifer and semi-confining units in eastern Arkansas. From a regional perspective, this collection of sediment can be divided into two units. The upper unit consisting of clay, silt, and fine sand confines the alluvial aquifer in places and is often referred to as the “clay cap.” The lower unit, which contains the alluvial aquifer, is composed of coarse sand and gravel that grades upward to fine sand. Hydraulic conductivity of the aquifer will thus be higher towards the bottom. It is this confining unit and alluvial aquifer, along with its flow system, that has been defined and investigated previously (Broom and Lyford, 1981; Ackerman, 1989a, 1989b, 1990; Mahon and Ludwig, 1990; Mahon and Poynter, 1993). The alluvium is underlain by older water-bearing units including the Cockfield Formation and the Sparta Sand, and still older units known as the Carizzo Sand, Wilcox Group, Midway Group, and Nacatoch Sand (Renken, 1998). These units are shown in a hydrogeologic section shown in figure 3.

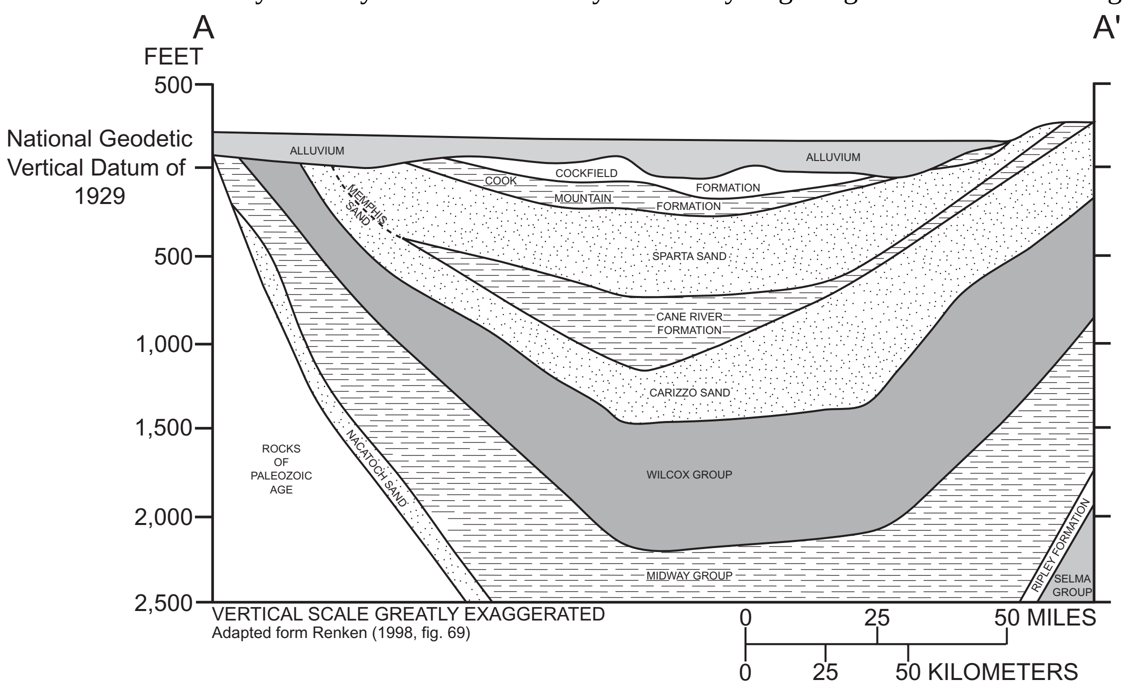


Figure 3. Idealized hydrogeologic section through the alluvium and underlying units. Trace of section shown in inset map on figure 1.

Both regional and local flow systems exist in the alluvial aquifer. Regional flow in the alluvial aquifer is considered to be water that has entered the deeper, coarse sand and gravel part of the aquifer and traveled for tens or hundreds of miles before discharging to major rivers such as the Mississippi, Arkansas, or White. Regionally, ground-water levels have been measured and mapped for several years (Ackerman, 1989b; Plafcan and Edds, 1986; Plafcan and Fugitt, 1987; Westerfield, 1990; Joseph, 1999). Ground water enters the modeled area laterally from the north and the west, flowing southward and eastward towards the Mississippi River or into cones of depression (fig. 4). Water-level altitudes range from nearly 300 ft in the north to less than 100 ft in the south over a distance of about 200 mi.

The alluvial aquifer underlies nearly all of eastern Arkansas, with the exception of Crowleys Ridge (fig. 1), which trends nearly north to south and divides the aquifer, north of the Arkansas River, into two hydraulically separate flow regimes. Crowleys Ridge averages about 10 mi in width and is an erosional remnant of strata of Tertiary age capped, in places, by several tens of feet of loess. In southern Craighead and northern Poinsett Counties the ridge is narrow, and the hydrologic units cropping out are the most transmissive of those comprising the ridge. Theoretically, this area would be the most likely to allow the transfer of ground water from one side of the ridge to the other. However, potentiometric maps of ground-water levels (Joseph, 1999; Schrader, 2001) indicate that there is about 20 to 30 ft of head difference from one side of the ridge to the other, which indicates that the ridge is a substantial barrier to flow. There is also a difference between water levels in the alluvial aquifer and those in the ridge with water levels in the ridge generally higher.

It is presumed that before development of the ground-water resource in the alluvial aquifer, flow was in a southward direction, with a general slope of about 1.2 ft/mi (Counts and Engler, 1954). Discharge from the aquifer was most likely to major rivers such as the Mississippi, Arkansas, and White. Prior to development, regional flow probably was southward beneath the Grand Prairie and the Arkansas River into the southern part of the alluvial aquifer and then into Louisiana.

Water levels in the aquifer have decreased with the increase in pumpage from the aquifer during development. Parts of the aquifer that once were confined are now (1998) unconfined (fig. 5).

The alluvial aquifer thickness as presented in the model presented by Mahon and Poynter (1993) varies substantially in the study area ranging from about 15 to 195 ft and averages about 100 ft (fig. 6). Pugh and others (1997, sheet 1) report thicknesses of 0 to 180 ft with an average of about 100 ft. The aquifer is thickest where the confining unit is thin or where depressions occur in the underlying Tertiary-age sediments.

The thickness of the confining unit or clay cap varies within the study area and as presented in the Mahon and Poynter (1993) model ranges from 0 to nearly 200 ft in the Grand Prairie (fig. 7). Thickness generally is 50 ft or less and the average thickness is about 25 ft. Gonthier and Mahon (1993, sheet 1) reported the clay thickness as 0 to 140 ft.

The thickness of the clay cap effects the integrity of the confining unit. The interconnection of laterally discontinuous but moderately transmissive sediments within the clay also affects the confining unit integrity.

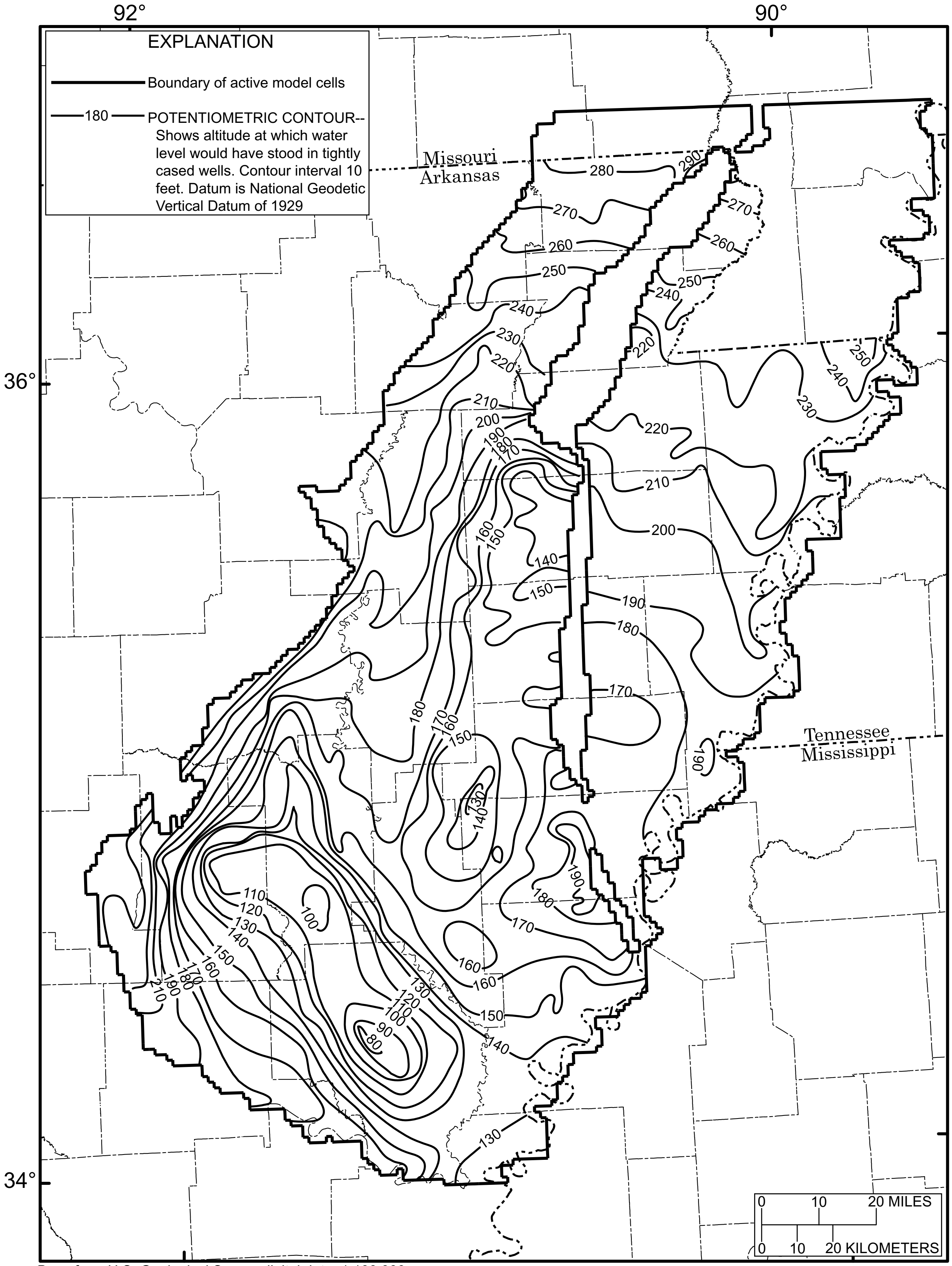
Sources and Sinks of Water

Upper Boundary—Areal Recharge

Annual precipitation within the study area averages about 49 in. annually (Freiwald, 1985), some of which seeps through the fine-grained material overlying the aquifer to the water table. Areal recharge to the aquifer is equal to precipitation minus (1) runoff into streams, (2) evaporation, and (3) evapotranspiration from plants in the soil zone. Infiltration of precipitation probably accounts for the largest amount of recharge to the aquifer.

Lower Boundary—Hydrologic Interchange Between Underlying Aquifers

Data related to the movement of ground water between the alluvial aquifer and the underlying hydrologic units during predevelopment time are sparse. Because outcrop areas of the older units are topographically higher than the aquifer, it is presumed that flow was upward from these older strata to the aquifer. However, water levels in the underlying aquifers have declined over time (Joseph, 2000) indicating that water levels in parts of the aquifer are now greater than those in the underlying sediments. In the southern part of the alluvial aquifer, including the Grand Prairie area, the alluvial aquifer water levels measured in 1998 were often 50 ft or more higher than the water levels measured in 1999 in the underlying Sparta aquifer (fig. 8).



Base from U.S. Geological Survey digital data, 1:100,000

Figure 4. Water-level altitudes in the alluvial aquifer in spring 1998 (modified from Joseph, 1999).

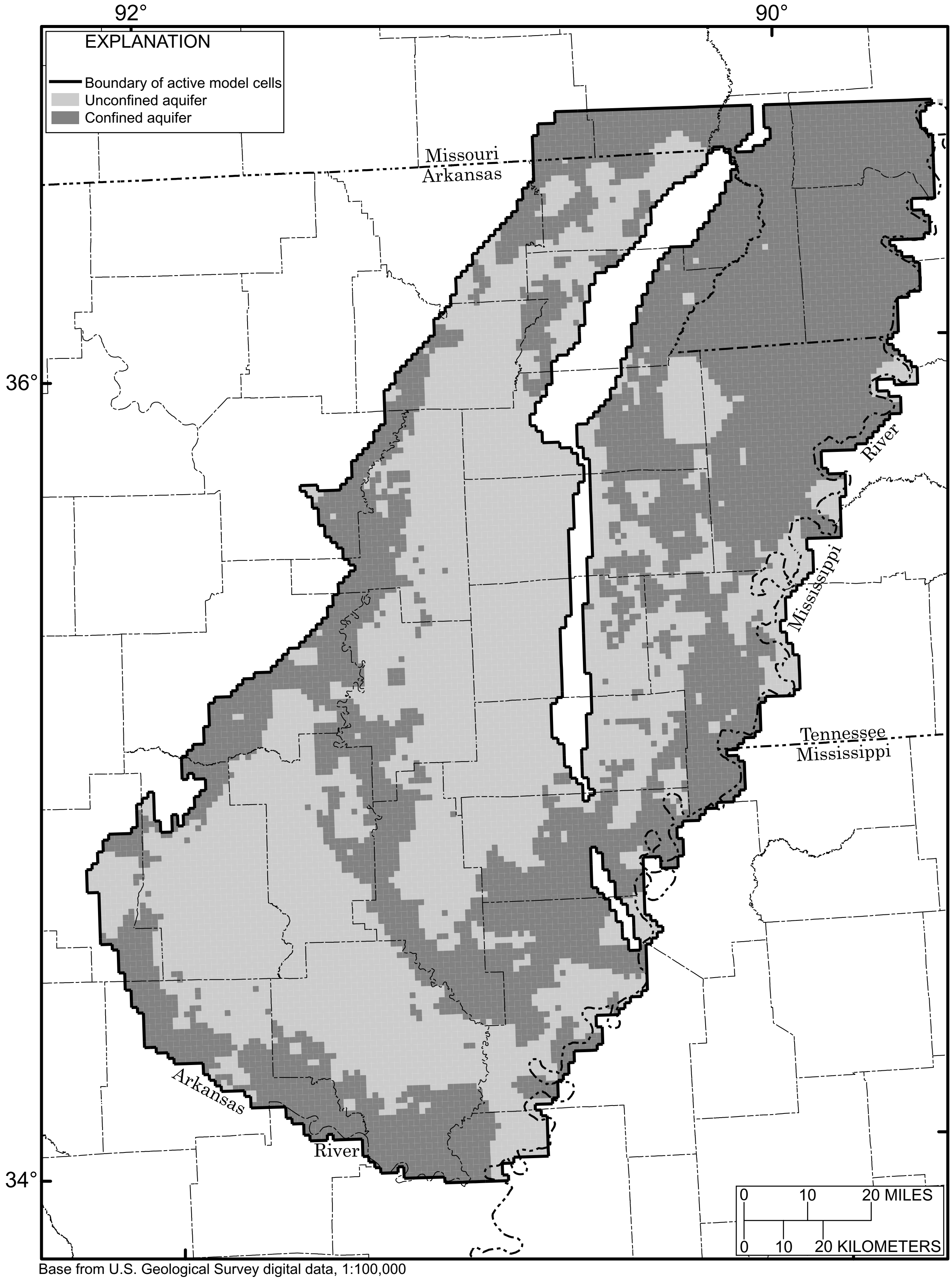


Figure 5. Areas of confined and unconfined conditions within the alluvial aquifer in 1998.

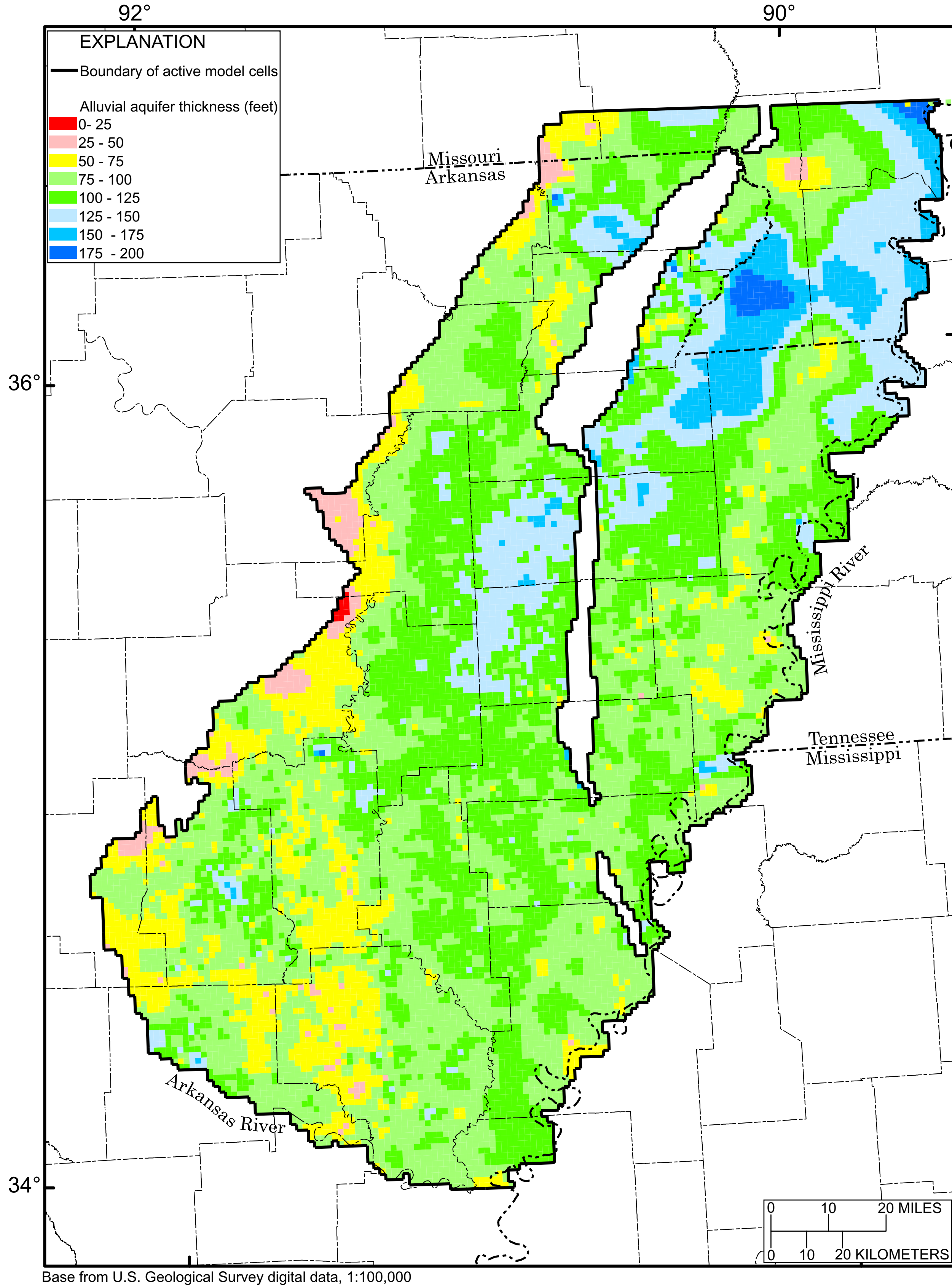
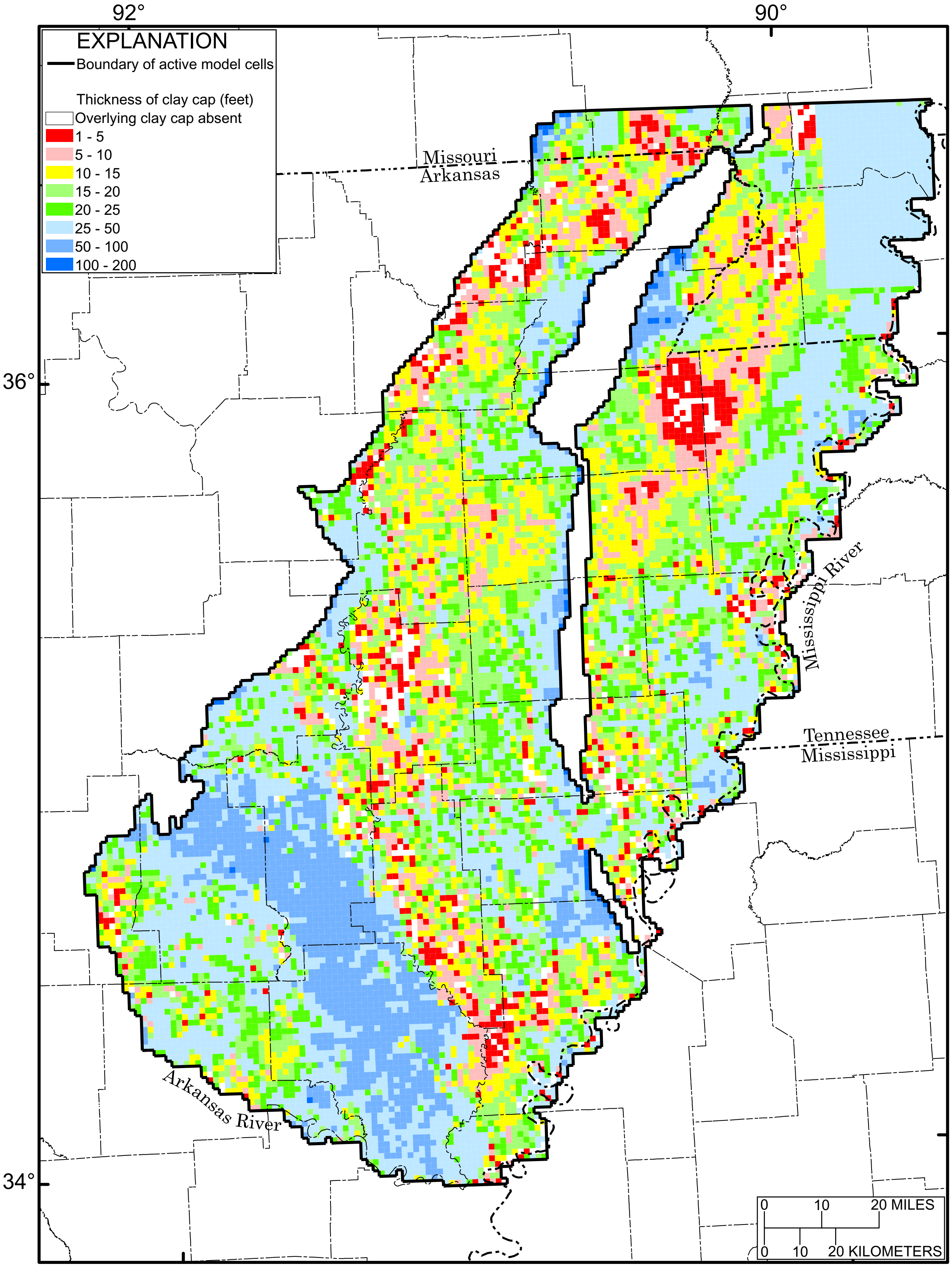


Figure 6. Thickness of the alluvial aquifer.



Base from U.S. Geological Survey digital data, 1:100,000

Figure 7. Thickness of the clay cap overlying the alluvial aquifer.

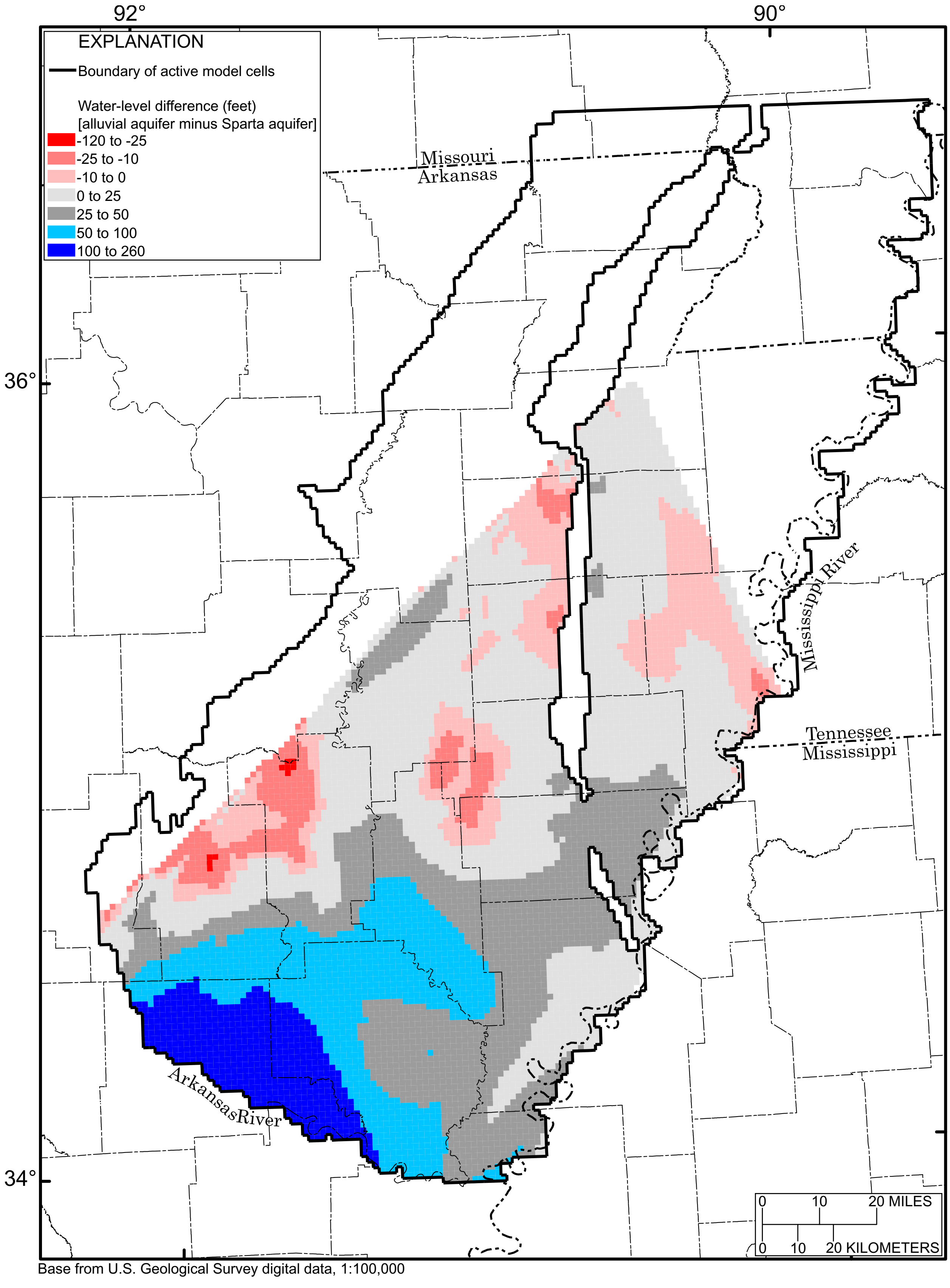


Figure 8. Difference between the alluvial aquifer water levels in 1998 and the underlying Sparta aquifer water levels in 1999.

This indicates that flow was from the aquifer into the underlying hydrologic units. Changing head relations between the alluvial aquifer and the underlying Sparta aquifer could over time change flows between the two aquifers. If water levels decline more rapidly in the alluvial aquifer than in the underlying units, flows downward into the Sparta aquifer will decline. Conversely, if water levels decline more rapidly in the Sparta aquifer, flows downward will increase.

Rivers

Many rivers flow across the alluvial plain and exchange water with the aquifer. The flow of water through riverbeds is dependent on the transmissive properties of the riverbed and the difference between the head in the aquifer and the river stage. The Mississippi and the Arkansas Rivers are presumed to have a good hydraulic connection with the aquifer because they are deeply incised into the aquifer, and the water level in the aquifer adjacent to the river is nearly identical to the river stage (Ackerman, 1989a, 1990; Mahon and Ludwig, 1990). The White and St. Francis Rivers may not be as well connected hydraulically with the aquifer although some hydrographs for wells near these rivers closely reflect changes in river stage (Ackerman, 1990, fig. 8). Field observations and water-level measurements indicate that other smaller streams such as the Cache River and L'Anguille River in the alluvial plain generally have less hydraulic connection with the aquifer (Ackerman, 1990, fig. 8). Model simulations and measured water-level altitudes (Joseph, 1999, plate 1) indicate that the Mississippi River remains a gaining stream throughout the study area from predevelopment times onwards. For other rivers, the general direction of water movement shifts from gaining to losing from predevelopment times to 1982 (Ackerman, 1990, p. 75).

Flow from the North

Recharge to the study area also includes the ground water that moves southward in the alluvial aquifer from Missouri. Because the hydrologic connection is continuous between the alluvial aquifer in Arkansas and the alluvial aquifer in Missouri, this flow from the north may be quantitatively important.

Lateral Flow from the West from Older Sediments

A small amount of the total recharge enters the aquifer from Tertiary- and Cretaceous-age sediments underlying the aquifer and from the consolidated formations of Paleozoic age flanking the western side of the study area (Ackerman, 1990). Potentiometric-surface maps (Joseph, 1999; Schrader, 2001) show flow eastward from the consolidated formations on the west.

Hydraulic Properties

The heterogeneity of the sediments within the aquifer is paralleled by that of the hydraulic properties of the ground-water system; that is, small stringer sand and silt beds are dispersed laterally and vertically, and represent local features of the aquifer and flow system. Ground-water flow related to these features may be hydraulically independent on a local scale. Although flow is made more complex by the multitude of heterogeneities in the aquifer and upper confining unit, the flow system can be generalized and conceptualized as water moving laterally in one or two zones or layers within the alluvial aquifer. This simplistic conceptualization of flow and hydraulic properties may be compatible with conditions observed in the field because of the coarse sand and gravel at the base of the aquifer that grades upward to fine sand.

A regional simplification of ground-water flow and hydraulic properties may be inappropriate when examined in detail at a local scale. Channel fill, point bar, and backswamp deposits associated with present or former channels of the major rivers locally can produce abrupt differences in lithology, resulting in spatial variations in the hydraulic properties of both the aquifer and confining unit within small distances. The local lithologic variations allow for small scale or localized flow systems that may have flow characteristics that differ somewhat from the regional flow system. Local flow systems may have recharge and discharge zones occurring within short distances, such as tens of feet to a few miles, from each other. Well yields in these shallow local aquifers are sufficient to provide for domestic sources of water but are much less than those from wells completed in the underlying regional system. For the purpose of this report in which the aquifer is considered at a regional scale, these local variations in lithology and flow characteristics are not considered.

Hydraulic Conductivity

Hydraulic conductivity of the alluvial aquifer ranges from 120 to 390 ft/d (Krinitzsky and Wire, 1964; Ackerman, 1989a). Geophysical and drillers' logs typically show an increasing particle size in the lower parts of the aquifer and the hydraulic conductivity is greatest in the coarse sand and gravel near the base of the aquifer. Because there are no laterally extensive confining units within the alluvial aquifer, the aquifer responds hydraulically as a single unit from a regional perspective.

Storage

Specific yield (applicable to unconfined conditions) and specific storage (applicable to confined conditions) are both required to characterize ground-water flow in the alluvial aquifer. Specific yield is the amount of water released per unit decline in hydraulic head; specific storage is the amount of water released from storage resulting from the compression of the aquifer matrix per unit decline in hydraulic head. Freeze and Cherry (1979, p. 60) present a range of 0.005 to 0.00005 for storage coefficient (dimensionless) in confined aquifers, which relates to specific storage values of 0.00005 to 0.0000005 (1/foot) with an average aquifer thickness of 100 ft, and a range of 0.01 to 0.30 for specific yield in unconfined aquifers. Driscoll (1986, p. 67) presents 0.10 to 0.30 as a representative range for specific yield for sand.

Water Use

Pumpage of water from the alluvial aquifer in eastern Arkansas varies annually, but has increased since the early years of development and is used mostly for irrigation. Ground-water pumpage in the study area ranges from an average of about 83,000,000 ft^3/d during 1918-1957 (Mahon and Ludwig, 1990) to about 636,000,000 ft^3/d in 1998 (Terry Holland, U.S. Geological Survey, written commun., 2002).

MODEL DEVELOPMENT AND CALIBRATION

The USGS finite-difference, three-dimensional, ground-water flow model, MODFLOW-2000 (Harbaugh and others, 2000), was used to develop and calibrate the ground-water flow model for the alluvial aquifer. The calibrated model was used to simulate ground-water flow in the aquifer and to evaluate the range of plausible values for hydrologic characteristics. MODFLOW-2000 was used to solve finite difference, ground-water flow equation approximations for spatial distributions of hydraulic head over time with certain simplifying assumptions. The Preconditioned-Conjugate-Gradient (PCG) solver, included in MODFLOW-2000, was invoked to solve the finite difference equation.

MODFLOW-2000 provides several additional processes (Hill and others, 2000) that were used to help estimate hydrologic properties and further evaluate the model. Cells considered to have a similar hydrologic properties were grouped together and assigned a common value for that property (referred to here as a parameter). This was done for a number of hydrologic properties. The sensitivity process and observation process were used together to calculate the sensitivity of hydraulic heads, called grid sensitivities, throughout the model with respect to each of these parameters using the accurate sensitivity-equation method. The observation process then uses these grid sensitivities to calculate sensitivities for the simulated water-level altitudes associated with the observed water-level altitudes. Thus, the sensitivity process and the observation process are used with observed and simulated water-level altitudes to evaluate the relative importance of the parameters to the modeling process and to calculate the objective function used by the parameter-estimation process. The parameter-estimation process uses a modified Gauss-Newton method to adjust values of user-selected input parameters in an iterative procedure to minimize the value of the weighted least-squares objective function. Thus, estimates of parameters were calculated to minimize difference between observed and simulated water-level altitudes.

The model simulated steady-state and transient conditions. For steady-state simulations, selected model parameters are constant with time during a single stress period. For transient simulations, selected model parameters may change with time during multiple stress periods.

Simplifying Assumptions

By definition, a model is a mathematical simplification of a process or a system. In that regard, two layers were used to represent the aquifer in the model described in this report to show vertical variations in horizontal hydraulic conductivity. Mahon and Poynter (1993) describe nearby wells in the alluvial aquifer open to both the upper and lower zones in the aquifer as having negligible head differences. Thus, vertical flow within the alluvial aquifer is likely to be low. In that regard, the model can be represented effectively with one or two layers. However, as vertical variations in hydraulic conductivity were believed to be important, two layers were used to represent the aquifer in this model. As water-level altitudes decrease due to pumpage, the arithmetic mean of the horizontal hydraulic conductivity will be higher if the lower layer has a higher value than the upper layer than if a single layer with the lower value is used.

The alluvial aquifer is modeled as two layers of equal thickness—the upper layer (layer 1) and lower layer (layer 2). The aquifer is modeled in this manner because of differences in hydraulic conductivities. Flow in both layers is modeled as confined or unconfined. Water-level altitude data indicate that pumping has drawn down these water levels below the upper confining bed throughout much of the alluvial aquifer. All lateral model boundaries are impermeable except for portions of the western boundary, portions of Crowleys Ridge, and a boundary of specified heads along the northern border.

Other simplifying assumptions in the model include: (1) the system is isotropic causing hydrologic properties to be spatially invariant; (2) all pumpage in a model cell can be simulated as coming from the cell center; (3) the pumpage throughout a stress period is applied equally throughout the stress period; (4) recharge is invariant over large periods of time; (5) small scale variations of hydraulic conductivity within cells are negligible; (6) all pumping and observation wells are completed in the lower layer; and (7) the head difference between the upper and lower layer is negligible.

Finite-Difference Grid

A finite-difference grid was used to subdivide the model area into a horizontally uniform cell network of 184 rows and 156 columns each 1 mi by 1 mi in size (fig. 2) as was done with the Mahon and Poynter (1993) model. The vertical dimensions of the cells also were derived from Mahon and Poynter (1993). The model code calculates a ground-water level at the center, or node, of each cell and a ground-water flux across each cell face based on water-level gradients between nodes. Of the 28,704 model cells, 14,104 are active. The model represents the alluvial aquifer in all or parts of 23 counties in Arkansas and 5 counties in southeastern Missouri.

Spatial and vertical variations in hydrologic characteristics in the aquifer framework were represented by discrete values in each of the model cells. Model cells extend vertically into the aquifer and divide the aquifer into discrete volumes of aquifer material that are assumed to have uniform hydrologic characteristics.

Initial Water-Level Altitudes

Initial water levels were specified at each model cell to represent conditions that probably existed before ground-water development began in the early 1900's (termed predevelopment conditions). Earliest potentiometric maps for the study area are limited to the Grand Prairie area (Engler and others, 1945), but even these water levels reflect that ground-water development had already begun. Consequently, a complete, accurate potentiometric-surface map for predevelopment conditions could not be constructed; however, previously developed flow models (Ackerman, 1989a; Broom and Lyford, 1981) have produced a distribution of water levels corresponding to approximate predevelopment conditions. It is these distributions, shown in figure 9, that were used as initial water-level altitude (predevelopment) conditions for the upper and lower model layers. The first stress period (stress period 1) was simulated so that steady-state conditions were attained. In this way, the hydraulic properties and boundary conditions were brought into hydrologic balance with the water-level altitudes used at the beginning of the transient time steps in 1918. Simulated initial water-level altitudes produced by the steady-state stress period 1 are shown in figure 10.

Stress Period Discretization

The stress periods for the model are shown in table 1. Stress period 1 was specified as steady-state to bring hydraulic heads into balance with the boundary conditions. Additional stress periods were all specified as transient. Stress periods 2 through 7 are the same as stress periods 1 through 6 used by Mahon and Poynter (1993) in their model. Stress period 8 is stress period 7 by Mahon and Poynter (1993) with the stress period extended through 1988 to extend the stress period 7 pumpage of the model of Mahon and Poynter (1993). Stress periods 5 and 7 stop at the ends of 1972 and 1982, respectively. Stress periods 9 and 11 are dis-

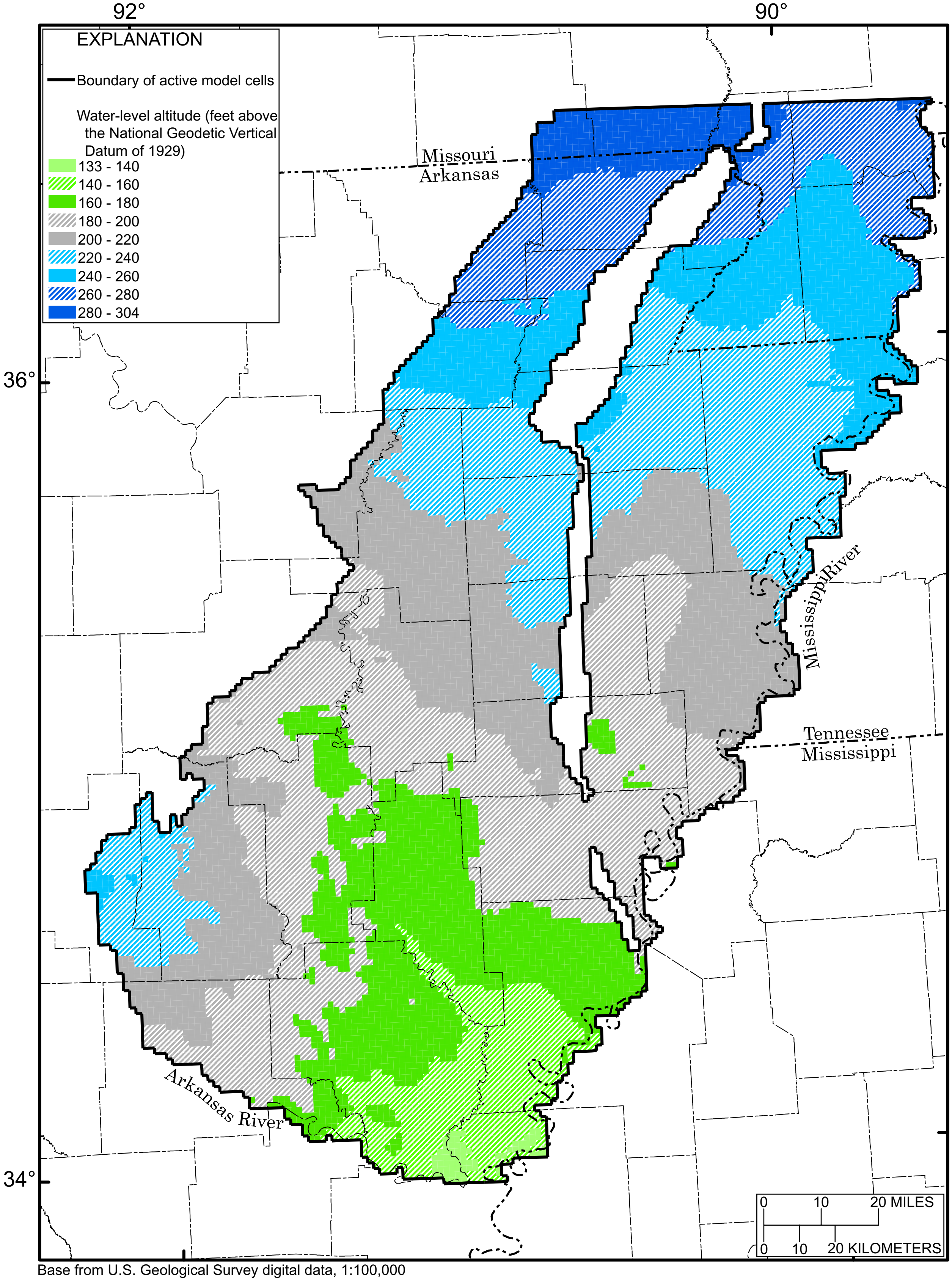


Figure 9. Initial water-level altitudes assigned to model cells in both layers.

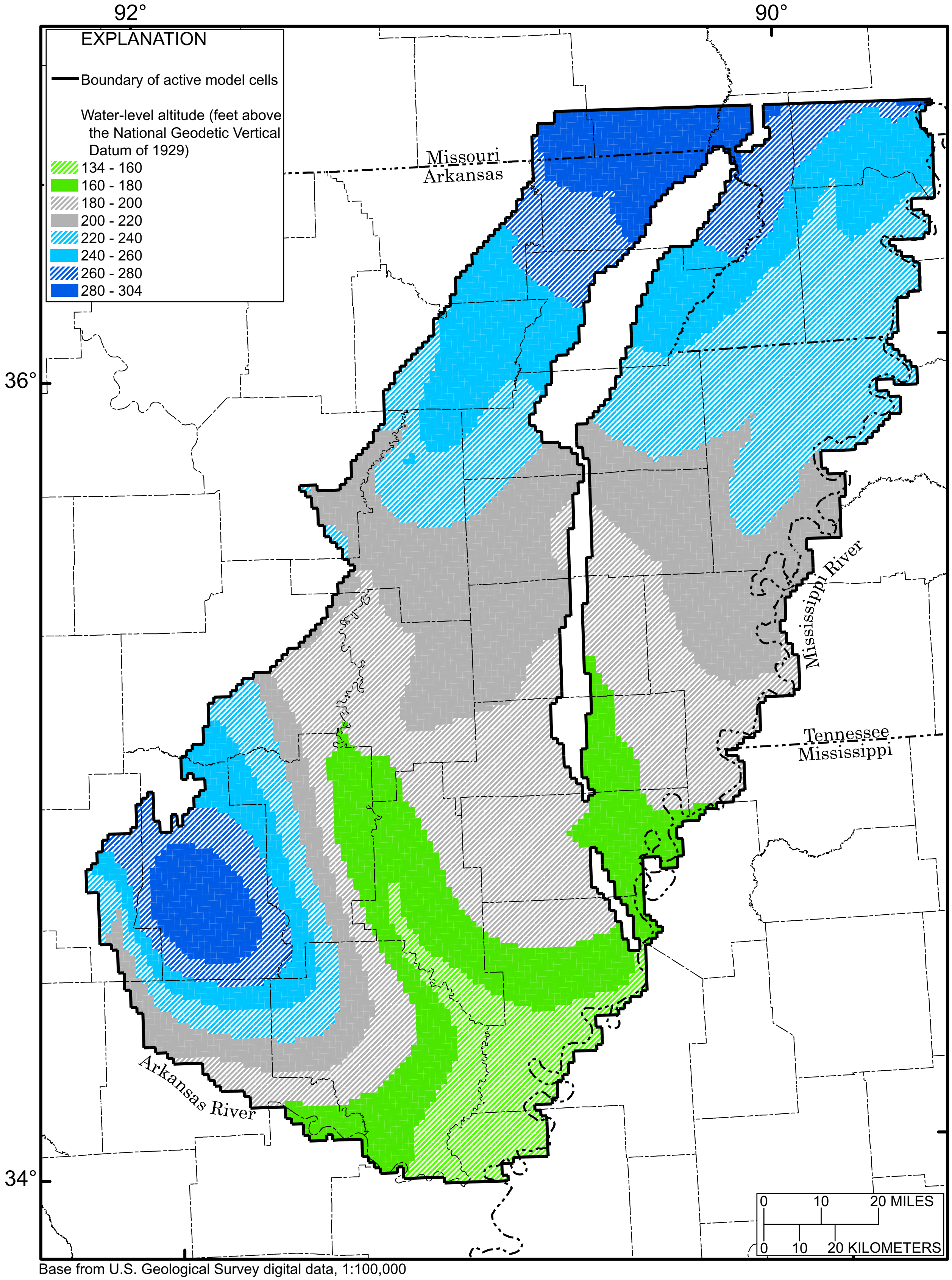


Figure 10. Simulated initial water-level altitudes in the lower layer produced by steady-state stress period 1 at the beginning of the transient simulations.

cretized to end in the spring (March) of 1992 and 1998, respectively, to allow the simulated water levels calculated at the ends of stress periods 9 and 11 to better match in time the water levels of observation wells measured during the springs of those years. Five additional stress periods (stress periods 12 to 16) were added in the projection simulations of future pumpage scenarios after 1998. These were roughly 10 years in length and stop at the end of 2049.

Flux Boundaries

Various flux-boundary conditions were specified in the model to represent different geohydrologic conditions in the flow system. The upper recharge boundary of the model represents the bottom of the confining bed or water table of the aquifer. The bottom of the aquifer is simulated as a no-flow boundary because data are insufficient to model the interaction between the alluvial and underlying aquifers. Flow to or from underlying aquifers is lumped with the areally distributed recharge.

Because of the relatively low hydraulic conductivity of the rocks flanking the aquifer, no-flow conditions were specified on much of the western side of the model. The potentiometric-surface map by Joseph (1999) indicates flow from the west in some areas, and general-head boundary cells were placed at selected locations (fig. 2). Values of hydraulic head for these cells were derived from the steady-state simulation of stress period 1 in the model. The conductance for cells defining this boundary were estimated during the parameter-estimation process.

Table 1. Stress periods and water use simulated

Stress period	Pumpage (cubic feet per day)	Length (days)	Length (years)	Beginning	End
1	0	Steady-state	Not applicable	Not applicable	Not applicable
2	82,806,688	14,610	40.00	1918	1957
3	92,879,304	1,826	5.00	1958	1962
4	127,464,032	1,826	5.00	1963	1967
5	142,161,888	1,826	5.00	1968	1972
6	249,573,216	1,826	5.00	1973	1977
7	434,081,856	1,826	5.00	1978	1982
8	391,937,728	2,192	6.00	1983	1988
9	621,882,112	1,184	3.24	1989	March 31, 1992
10	621,709,632	642	1.76	March 31, 1992	1993
11	631,324,480	1,549	4.24	1994	March 31, 1998
12	623,440,512-749,344,192	4,293	11.75	March 31, 1998	December 31, 2009
13	614,197,568-844,568,640	3,652	10.00	January 1, 2010	December 31, 2019
14	605,206,656-782,530,496	3,653	10.00	January 1, 2020	December 31, 2029
15	596,392,384-761,897,344	3,652	10.00	January 1, 2030	December 31, 2039
16	590,821,760-746,841,088	3,653	10.00	January 1, 2040	December 31, 2049

No natural geohydrologic boundary separates flow in the alluvial aquifer between Arkansas and southeastern Missouri. Accordingly, the alluvial aquifer in southeastern Missouri was included in the simulations because pumpage in that area may affect pumpage east of Crowleys Ridge in eastern Clay and Greene Counties, Arkansas. A specified-head boundary was placed in southeastern Missouri, even though this is not a natural boundary, because agricultural development of ground water in this area is minimal compared to that in Arkansas. This arbitrary boundary is about 10 mi (10 rows of cells) north of the Arkansas-Missouri border west of Crowleys Ridge and extends eastward to the Mississippi River. The placement of this boundary is sufficiently far from the areas of major pumping in Arkansas so as not to adversely affect simulation of water levels in these areas of interest. There are 63 specified-head nodes within the model governing, in part, the flow of water from southeastern Missouri into Arkansas. The heads for these specified-head nodes represent predevelopment conditions that were derived by Ackerman (1989a).

Some potential for flow exists between the aquifer and sediments underlying and composing Crowleys Ridge. In particular, the Memphis Sand, which is the coalescing of the Sparta Sand, Cane River Formation, and Carrizo Sand starting on the west of the hydrogeologic section shown in figure 3, extending out on both sides of the ridge in southern Craighead, northern Cross, and Poinsett Counties. Hydraulic connection with the Memphis Sand underlying the aquifer could account for some flow between the hydrologic units. In 1990, water levels in the Memphis Sand beneath western Cross and Poinsett Counties (Westerfield, 1990) were higher than water levels in the alluvial aquifer indicating possible ground-water flow from the Memphis Sand into the alluvial aquifer. For purposes of this report, parts of the ridge were simulated with a general-head boundary condition in the same manner as modeled by Ackerman (1989a) and Mahon and Poynter (1993). The conductances for the general-head boundaries were estimated during the parameter-estimation process.

River cells representing the Mississippi River were used to allow flux along the eastern boundary of the model as well as cells representing the Arkansas River along the southern boundary of the model (fig. 2). The hydrologic effect of these large rivers presents an effective hydrologic sink preventing ground-water flow to the east and south from the model area.

Eleven rivers were chosen to be actively simulated within the alluvial model. These rivers include the Arkansas River, Bayou Meto, Black River, Cache River, Current River, L'Anguille River, Little Red River, Mississippi River, Right Hand Chute of the Little River, St. Francis River, and the White River (fig. 2). Most of these rivers were classified as perennial by Hunrichs (1983). The river stages, which were modeled as constant throughout all stress periods, were derived from mean historic gage stages that were then interpolated to all river cells. The riverbed altitudes were set to zero in order to ensure underlying sediments never become unsaturated because these large, perennial streams would over time have saturated the underlying sediments down to the water-level altitude. The conductances for these 10 rivers were individually estimated during the parameter-estimation process and subsequently adjusted to better match the water-level altitude residuals. No attempt was made to relate these river conductances to stream lengths or widths within each model cell.

Hydraulic Conductivity

Horizontal hydraulic conductivities for model cells were estimated with the aid of the parameter-estimation process of MODFLOW-2000 and available aquifer test. The model was divided areally into 10 parameter zones (fig. 11) based on the surficial geology map of Arkansas (Haley and others, 1976). These surficial geologic units were considered to adequately represent the spatial variability of hydraulic properties from which to assign or estimate hydraulic conductivity and other hydrogeologic properties. The model was arbitrarily discretized into two equally thick vertical layers of aquifer material because of the vertical variability of horizontal hydraulic conductivity. This permitted a higher horizontal hydraulic conductivity to be used in the lower layer where coarser sands and gravels are present.

The model developed by Mahon and Poynter (1993) used a uniform value of 275 ft/d for hydraulic conductivity. Arthur (2001) specified a value of 425 ft/d for hydraulic conductivity in his calibration of a model of the alluvial aquifer east of the Mississippi River in Mississippi.

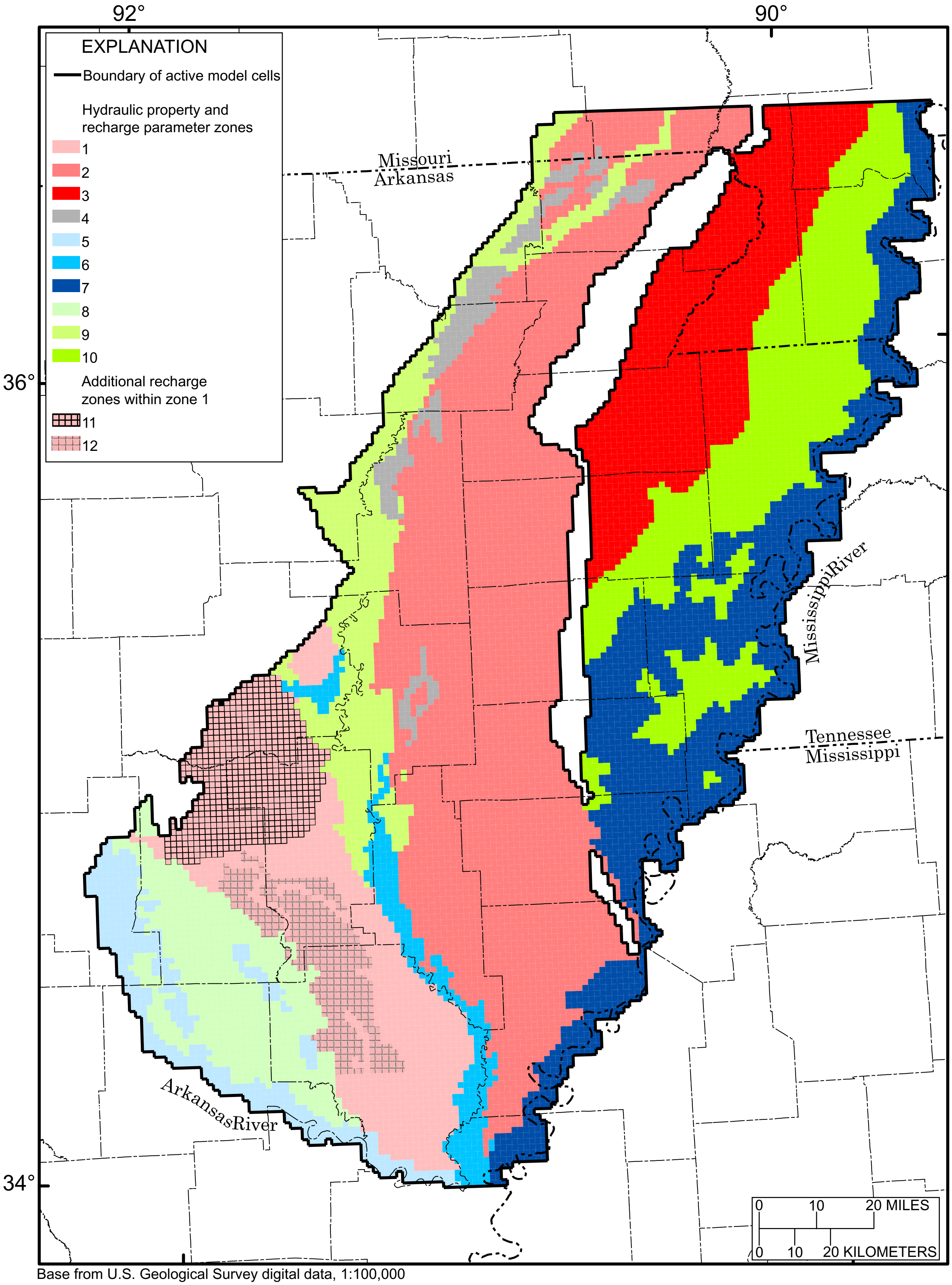


Figure 11. Parameter zones used to calibrate the hydraulic properties and recharge.

Rovey and Cherkauer (1995) describe the scale dependency of hydraulic conductivity obtained from hydraulic testing of wells. They showed that as the radius of investigation increases, so does the value of hydraulic conductivity. The hydraulic conductivities used in this model with model cells 1 mi in width may represent values larger than those calculated from aquifer tests with test radii much smaller than 1 mi. As such, the plausible upper limits for hydraulic conductivity used in this model may be 400 to 600 ft/d.

An examination of geophysical and drillers' logs indicates coarser alluvial material towards the bottom of the aquifer (Mahon and Ludwig, 1990; Mahon and Poynter, 1993). With such vertical variation in horizontal hydraulic conductivity, a vertical discretization of the aquifer may provide a better approximation of hydrologic properties. Earlier simulations of this model indicated that temporal variation in the arithmetic mean of horizontal hydraulic conductivity might provide a partial solution to some problems in calibration, particularly with simulated head variations over time north of the Arkansas River. The model was, therefore, divided vertically into two layers of equal thickness and provided with different parameter values for horizontal hydraulic conductivity. These parameter values were calibrated so the mean between them remained within reasonable values for overall horizontal hydrologic conductivity.

Recharge

Recharge zones for the model were based on the same parameter zones used for horizontal hydraulic conductivity (fig. 11). Recharge values were then modified locally to improve model fit. Although minimally constrained, recharge estimates are a function of vertical hydraulic conductivity which is a function of geology. Hence, surficial geologic units are likely to represent a reasonable initial distribution of recharge. The model developed by Mahon and Poynter (1993) simulated recharge using the MODFLOW River Package to simulate flow from the upper clay cap down into the aquifer. As simulated water-level altitudes in the aquifer declined, flows from this upper region into the aquifer increased by a factor of 2.3, in effect increasing recharge over time. Because the current model simulates recharge directly using the MODFLOW recharge package, it was necessary to simulate recharge differently for earlier and later stress periods in a manner that increases recharge over time, like the model of Mahon and Poynter (1993).

The recharge specified in the model represents the lumping of recharge due to precipitation, flow to or from some rivers, and ground-water flow to or from underlying aquifers. Fluxes from underlying aquifers such as the Sparta aquifer may either augment or counteract the recharge from precipitation, depending on the difference between the alluvial aquifer and Sparta aquifer water levels (fig. 8). Ground-water flux to and from underlying aquifers will depend on differences in water-level altitudes. Also, the surface of the aquifer is crossed with numerous perennial and ephemeral streams not accounted for directly in the model. These streams may saturate the aquifer material down to the ground water and, therefore, fluxes to and from the aquifer and these rivers will depend on differences in water-level altitudes. These different processes represented in the model recharge package may be strongly dependent on water-level altitudes and therefore will vary over time as these water-level altitudes change. Also, reducing water-level altitudes below the root zone will have the effect of reducing evapotranspiration and increasing recharge due to precipitation. Underestimates of water use also will have the effect of increasing apparent recharge. Thus, many different factors that are poorly understood and accounted for in the model may have the effect of increasing flows into the model from the recharge package over time. As such, recharge zones were modified locally during the calibration process to better match observed water levels. This was done first for recharge flows for the stress periods 1-5 (1918-1972) and then for stress period 11 (1994-March 1998). Recharge flows for the intervening stress periods 6-10 (1973-1993) were prorated on a cell by cell basis from the first recharge flows to the last.

Water Use

Water use for model cells in Arkansas was obtained from Mahon and Poynter (1993) for 1918 to 1988 (stress periods 2 through 8). Pumpage for 1918 to 1958 (stress period 2) was estimated based on results of previous models in eastern Arkansas (Mahon and Ludwig, 1990). Although pumpage has been documented as county totals, category totals, and aquifer totals in water-use reports, the documentation of the spatial distribution of pumpage within a county prior to 1997 is lacking. Computation of pumpage distributions for stress periods 3 through 8 was based on estimates of ground-water use for six, 5-year time periods beginning in 1958 (Stephens and Halberg, 1961; Halberg and Stephens, 1966; Halberg, 1972 and 1977; Holland and Ludwig, 1981; Holland, 1987). The total pumpage from the alluvial aquifer reported for each county was used to compute the distribution of pumpage within the county. The total pumpage simulated in each stress period is given for the model in table 1.

From 1989 through March 1998 (stress periods 9 through 11), water-use estimates were compiled for the model described in this report (Terry Holland, U.S. Geological Survey, written commun., 2002) using site-specific water-use data augmented by estimates of total ground-water use. Reported pumpages for 1991 were used for 1989 through 1993 (stress periods 9 and 10) while reported pumpages for 1997, which included water-use data from about 33,000 wells, were used for 1994 through March 1998 (stress period 11).

Total water use simulated by the model ranged from about 83,000,000 ft^3/d for 1918-1957 to about 392,000,000 ft^3/d for 1983-1988, and about 631,000,000 ft^3/d for 1994-1998. Figure 12 shows the distribution of pumpages used for 1994 to 1998 (stress period 11), which is the latest period of actual water-use estimates used in the model.

The pumping rates for model cells in Missouri were obtained from Mahon and Poynter (1993) for the period 1918 through 1988 (stress periods 2 through 8). From 1989 through March 1998 (stress periods 9 through 11), pumping rates were based on data obtained from National Water-Use Information System (Terry Holland, U.S. Geological Survey, written commun., 2000; http://water.usgs.gov/watuse/spread95/moco95.txt, accessed May 17, 1997) for county totals for 1991 and 1997. These county totals were prorated over model cells by an area-weighted process.

Storage

Specific storage and specific-yield values for model cells were estimated using the parameter-estimation process of MODFLOW-2000. The parameter zones were based on the same delineations as the horizontal hydraulic conductivity and recharge (fig. 11).

Mahon and Poynter (1993) used values of 0.25 to 0.35 for specific yield in their model and 0.05 for storage coefficient, which relates to about 0.0005 for specific storage assuming an average aquifer thickness of 100 ft. Arthur (2001) used values of 0.32 for specific yield and 0.0016 for specific storage in his calibration of a model of the alluvial aquifer east of the Mississippi River in Mississippi. These values were used as a guide for plausible values during the calibration process.

Model Calibration Procedure

During calibration, various aquifer parameters were adjusted, either with the parameter-estimation process or manually, so that the observation well water-level altitudes located throughout the modeled area (fig. 13) matched as closely as possible the water levels simulated by the model. All simulated water levels used in the calibration were taken from the lower layer. At stress period 11, in the final simulation, the maximum difference between heads in the upper and lower layers was less than 0.06 ft. All observation wells penetrated the lower model layer. The observed water levels were measured in the springs of 1972, 1982, 1992, and 1998. The residuals associated with observed and simulated water-level altitudes were reduced to a minimum value during the calibration process. The observation process of MODFLOW-2000 allows for the weighting of observed water levels to reflect assumed differences in importance to the calibration process. Because water-use data in 1991 and 1997 are believed to be more accurate than those of earlier years, the observations for 1992 and 1998 were weighted more heavily than those of the earlier stress periods.

A parameter-estimation process of MODFLOW-2000 was used in the first phase of the calibration process. The parameters estimated in this phase were recharge, hydraulic conductivity, specific storage, specific yield, and conductances for cells with riverbeds and general-head boundaries. Parameter values were applied over broad parameter zones or over extended linear features such as rivers.

Not all parameters were estimated together in a single parameter-estimation process but were estimated iteratively during several successive processes. Nevertheless, all these parameters could be evaluated within the same sensitivity process, which produces scaled sensitivities for each parameter allowing the importance of each parameter to the calibration process to be compared. Although the parameter-estimation process can be used to estimate hydrologic parameters over large areas, local variations in these parameters may be important. During the calibration process, these parameters may be adjusted manually to better reflect local hydrologic conditions.

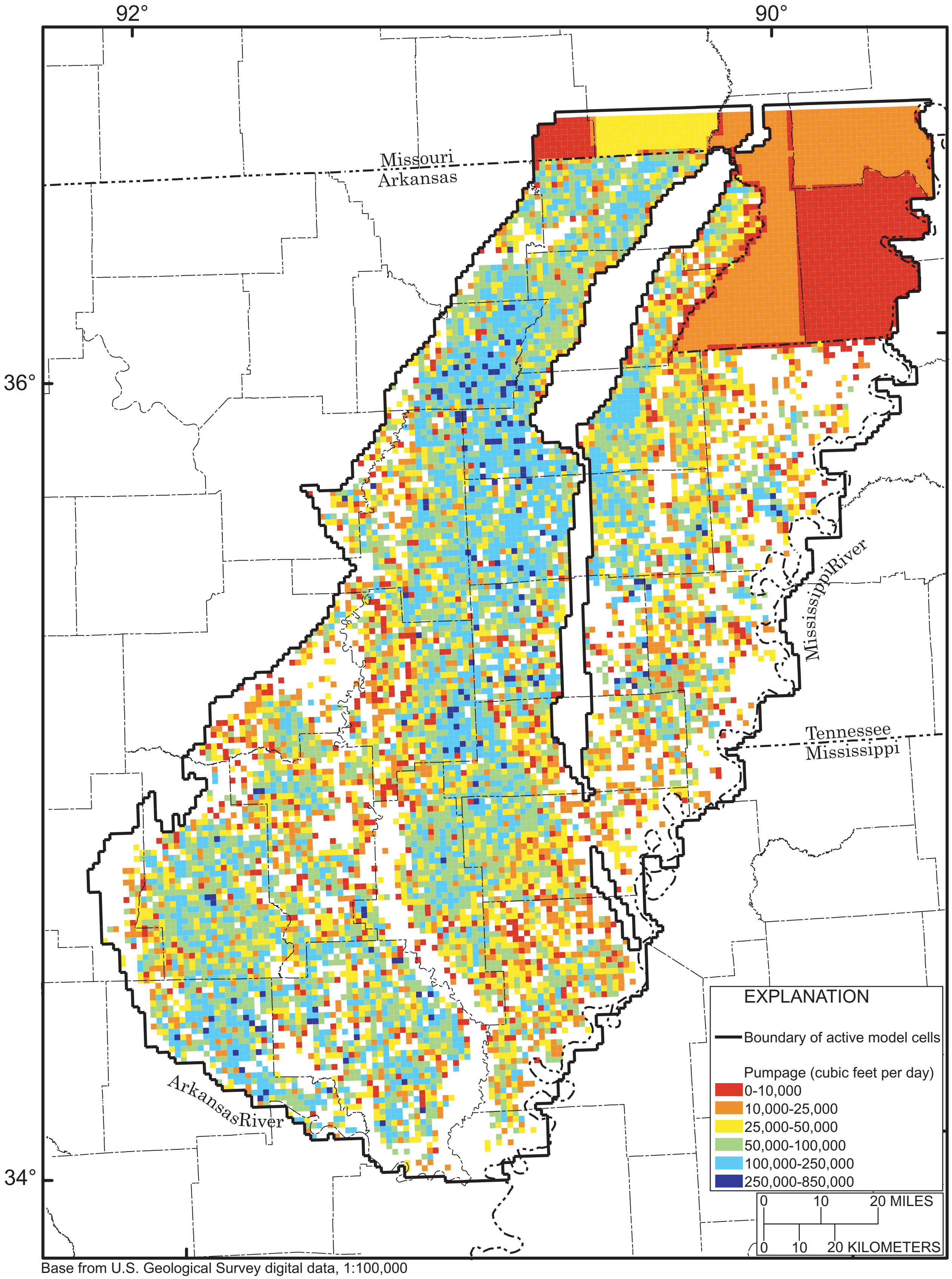
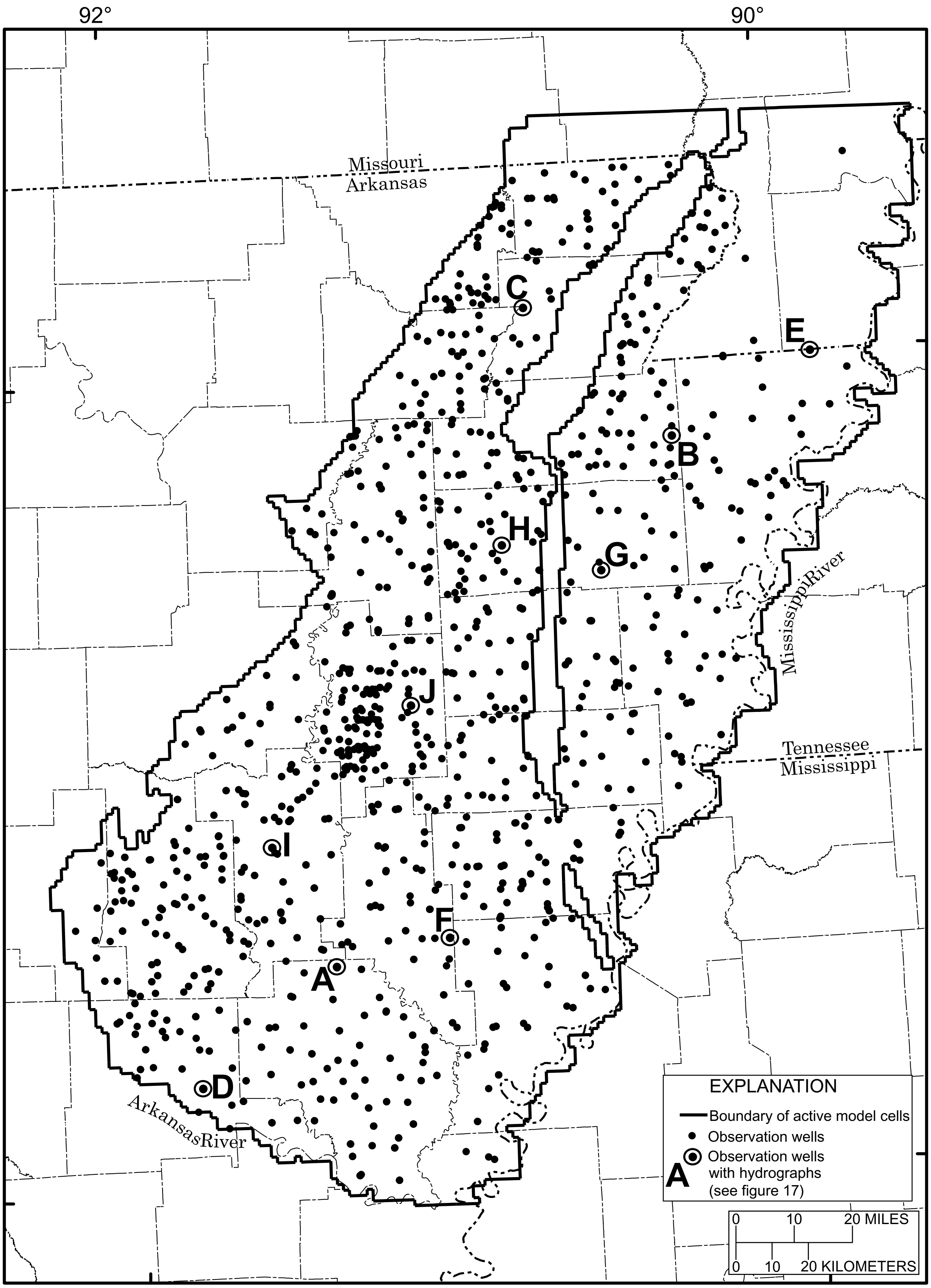


Figure 12. Distributions of pumpage used for 1994-1998 (stress period 11).



Base from U.S. Geological Survey digital data, 1:100,000

Figure 13. Observation wells used in the calibration process.

During the parameter-estimation process, the two hydrologic parameters most important to the model, hydraulic conductivity and specific storage, tended to increase to unrealistic levels in some areas. This may have been due to the underestimation of recharge in some of the earlier simulations. A decision was made to manually adjust the recharge parameter using small local zones based largely on areas with high residuals. Using professional judgment, recharge rates were modified within plausible limits in these zones to reduce these residuals. In these local areas, local multipliers were provided for the recharge parameter for the applicable zone. This was done first for recharge rates used for stress periods 1-5 and then for stress period 11 with the recharge rates for the intervening stress period prorated between these two recharge rates on a cell-by-cell basis.

Similarly, as river fluxes were believed to be locally important as sources of flow into the alluvial aquifer, professional judgment was used to alter riverbed conductances within plausible limits locally to reduce residuals. As with the recharge parameters, the riverbed conductance parameters were altered locally with multipliers to better reduce water-level altitude residuals.

Other hydrologic parameters of the model including horizontal and vertical hydraulic conductivity, specific yield, specific storage, and the conductances of the general-head boundaries were similarly estimated with the parameter-estimation process. The parameters were altered manually to better reduce water-level altitude residuals but without local multipliers to provide local variation as with recharge.

CALIBRATION RESULTS AND MODEL EVALUATION

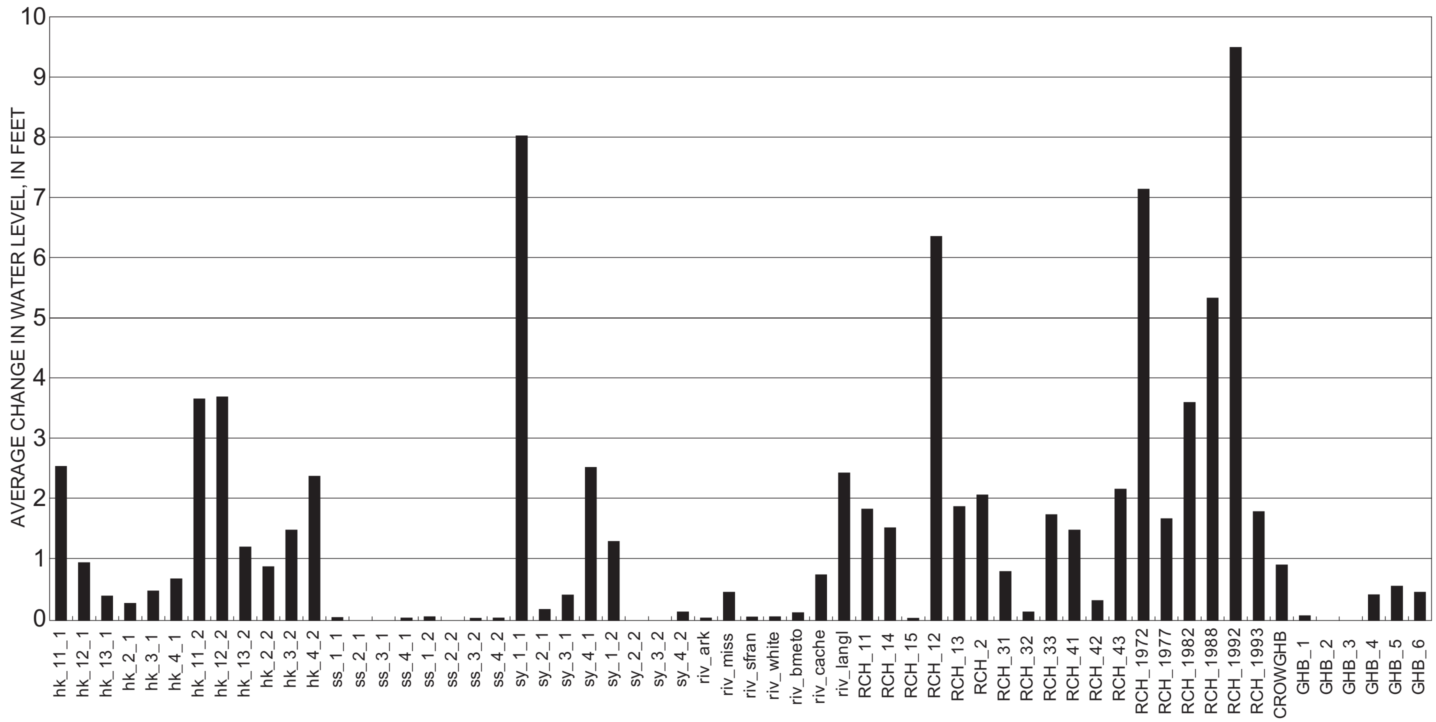
Through a combination of the parameter-estimation process and manual alterations of hydrologic parameters and the application of localized multipliers for the recharge and river conductance parameters, the difference between observed and simulated water-level altitudes was minimized. Table 2 shows the final calibrated values for the hydrologic parameters of horizontal hydraulic conductivity and specific storage. Hydraulic conductivity values were much higher for the lower layer than for the upper layer to be consistent with the occurrence of coarser, more transmissive sediments in the lower layer. Hydraulic conductivity in the upper layer was set at 230 ft/d throughout the model. In zone 2, hydraulic conductivity ranged from 230 ft/d to 730 ft/d, giving a maximum mean for the two model layers of 480 ft/d.

Table 2. Values of hydrologic parameters derived from the calibration process

Zone (fig. 12)	Hydraulic conductivity, upper layer (feet per day)	Hydraulic conductivity, lower layer (feet per day)	Mean of hydraulic conductivity (feet per day)	Specific storage for both layers
1	230	230	230	0.000001
2	230	730	480	0.000001
3	230	730	480	0.000001
4	230	730	480	0.000001
5-7	230	730	480	0.000001
8-10	230	730	480	0.000001

A uniform value of 0.30 was specified throughout the model for specific yield. Vertical hydraulic conductivity was set as a uniform value of 23 ft/d for layer 1 and 73 ft/d for layer 2, which represented values of 10 percent of horizontal hydraulic conductivity for most of the modeled area. For calculations of VKCB, the vertical conductance of the quasi-three dimensional layer dividing the two model layers as described in Harbaugh and others (2000, p. 29), the harmonic mean of these two vertical conductivities was subsequently used. The need to translate this model into a form usable by MODFLOW-96 to be used in connection with MODMAN optimization modeling (Greenwald, 1998) required this approach. As field observations (Mahon and Poynter, 1993, p. 11) show vertical flow within the aquifer vertical flow between the two layers is negligible and was not regarded to be of likely hydrologic significance and was not included in the parameterization process. These values produce a very low vertical hydraulic gradient between the two layers consistent with field observations (Mahon and Poynter, 1993, p. 11).

The average change in water level over the modeled area caused by a doubling of a given parameter, while others remain constant, is shown in figure 14. This demonstrates the sensitivity of each parameter to



EXPLANATION

hk_11_1...hk_4_2	Horizontal hydraulic conductivity, zone...zone 4 layer 2	RCH_11...RCH_43	Recharge stress periods 11-16 for various zones
ss_1_1_....ss_4_2	Specific storage, zone 1 layer 1..zone 4 layer 2	RCH_1972	Recharge stress period 5
sy_1_1....sy_4_2	Specific yield, zone 1 layer 1..zone 4 layer 2	RCH_1977	Recharge stress period 6
riv_ark	Arkansas River bed conductance	RCH_1982	Recharge stress period 7
riv_miss	Mississippi River bed conductance	RCH_1988	Recharge stress period 8
riv_sfran	St. Francis River bed conductance	RCH_1992	Recharge stress period 9
riv_white	White River bed conductance	RCH_1993	Recharge stress period 10
riv_bmeto	Bayou Meto bed conductance	CROWGHB	General head boundary conductance for boundary of Crowleys Ridge
riv_cache	Cache River bed conductance		
riv_langl	L'Anguille River bed conductance	GHB_1....GHB_6	General head boundary conductance for western boundary

Figure 14. Average change in water level from a doubling of the parameter.

the model results. The parameters with large sensitivity, such as sy_1_1, (specific yield in zone 1, layer 1) or recharge represent large areas within the model. The recharge parameters provided the largest sensitivity for the model. With the exception of L'Anguille River, the sensitivities for river conductance were small.

Volumetric budget component rates for the model representing 1968 through 1972 and 1994 to March 1998 (stress periods 5 and 11, respectively) are shown in figure 15. The increase in recharge from stress period 5 to stress period 11 is shown. Selected volumetric budget component rates for the model representing all stress periods are shown in figure 16. The increase in pumpage in stress period 7 (1978 to 1982) was due to drought conditions during this period.

Hydrographs for 10 representative observation wells used in the calibration process, are shown on figure 17 and well location are shown on figure 13. The hydrographs were constructed by taking the water-level altitudes for the period of record for these 10 observation wells and interpolating both spatially and temporally the simulated water-level altitude for each observation. The four observations used for calibration in stress periods 5, 7, 9, and 11 are circled on each hydrograph. These hydrographs generally show a reasonable fit between observed and simulated water-level altitudes. Further analysis of these hydrographs shows water-level changes common to most wells probably are caused by changes in precipitation, but differences in water-level changes among wells probably are caused by localized changes in pumpage.

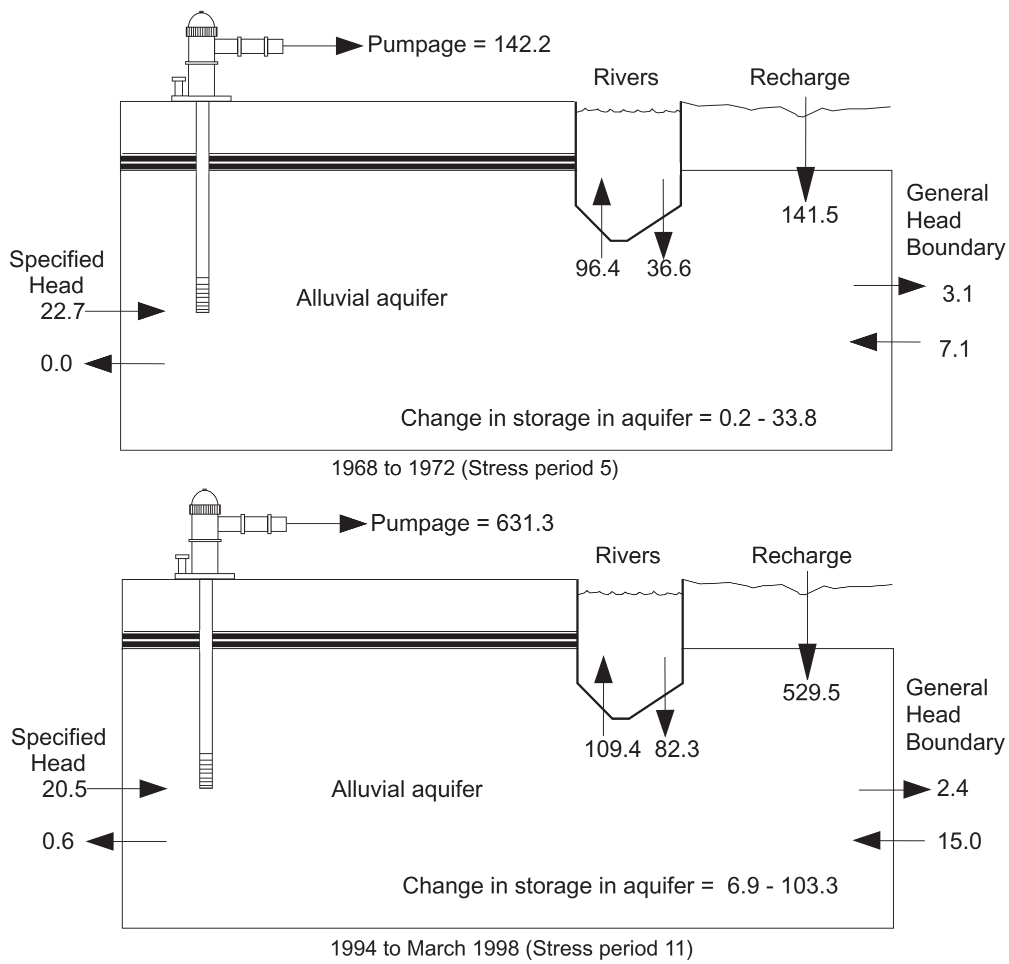


Figure 15. Volumetric budget component rates for 1968-1972 and 1994-March 1998 (stress periods 5 and 11). All numbers are in millions of cubic feet per day.

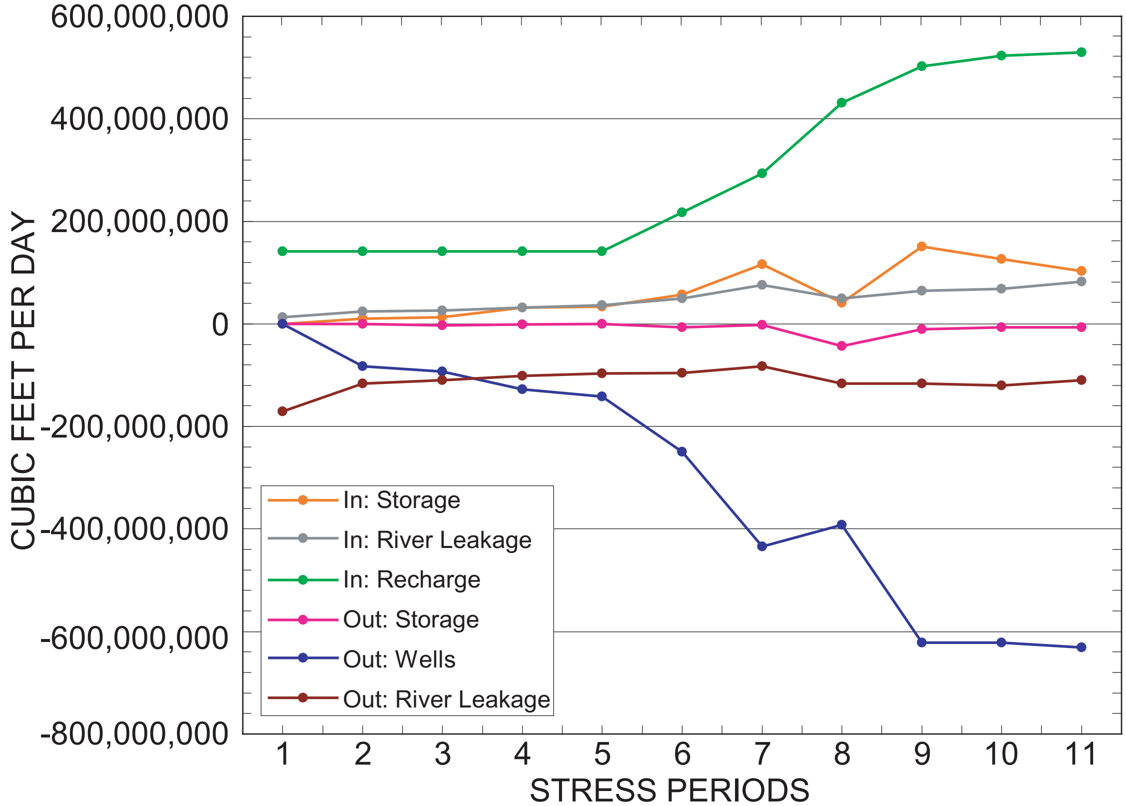


Figure 16. Selected volumetric budget component rates for the model by stress period.

Water level residuals for observation wells representing the stress periods ending in 1972, 1982, 1992, and 1998 (stress periods 5, 7, 9, and 11) are shown in figures 18, 19, 20, and 21. Negative residuals indicate observed water-level altitudes were lower than those simulated; positive residuals indicate observed water-level altitudes were higher than those simulated. Observed water levels generally were lower than simulated water levels in periods ending in 1972, 1982, and 1992 (stress periods 5,7, and 9), but were more random in distribution in the period ending in 1998 (stress period 11). This also is reflected in the histograms of residuals shown in figures 22 through 25. The simulated water levels in the Grand Prairie area during stress period 5 ending in 1972 were higher than observed water levels as shown by the negative residuals in figure 18. This may be due to underreported pumpage for this area during this period or to downward fluxes into underlying aquifers that are larger than surficial recharge during this period.

Table 3 shows a series of statistical measures for the sets of water-level residuals. The residuals for 1994 to March 1998 (stress period 11) have a mean for the absolute value of all residuals of 4.69 ft and a root-mean-square error of 6.04 ft. Histograms and statistical measures show minimal negative or positive bias in residuals for each of the four observation periods.

Figure 26 shows the simulated water-level altitudes for the stress period ending in March 1998 (stress period 11). These levels compare well with the contours of observed water-level altitudes for spring 1998 (fig. 4). Both simulated and observed water levels show ground-water flow southward into cones of depression, flow into the Mississippi River, and flow from the Arkansas River into the aquifer.

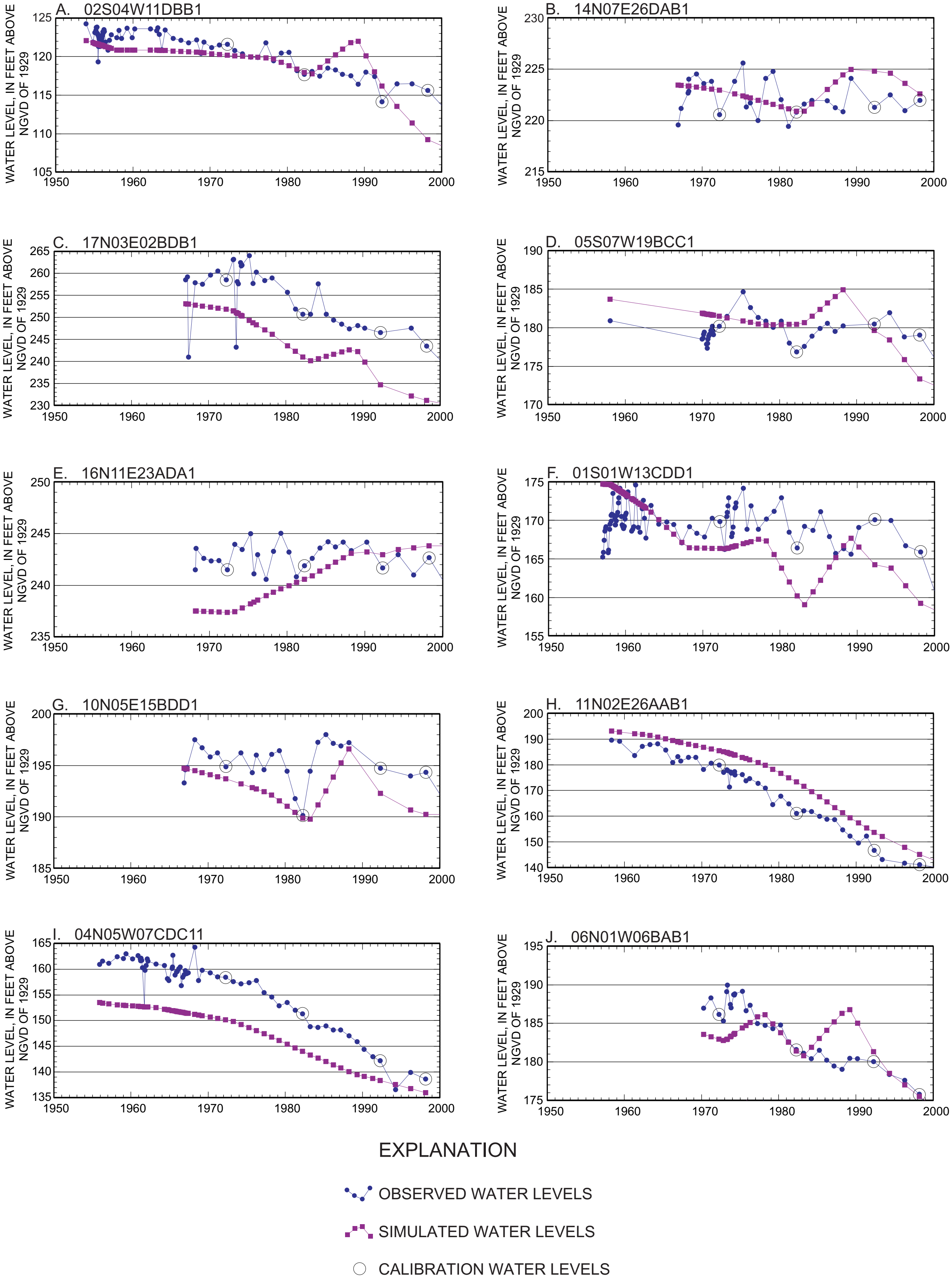


Figure 17. Selected hydrographs of observed and simulated water levels showing water levels used for calibration. See figure 13 for locations.

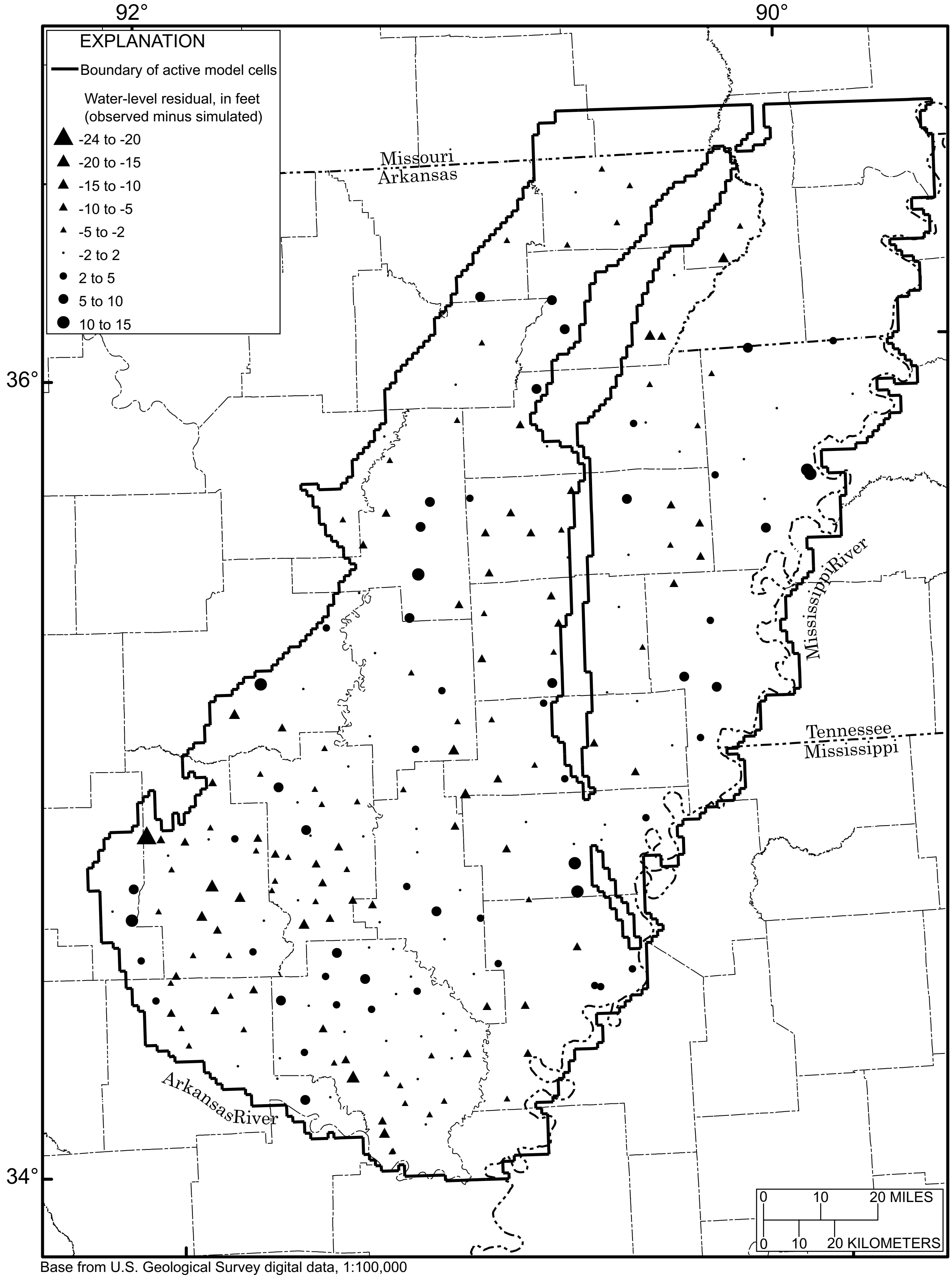


Figure 18. Water-level residuals for observation wells at the end of 1972 (stress period 5).

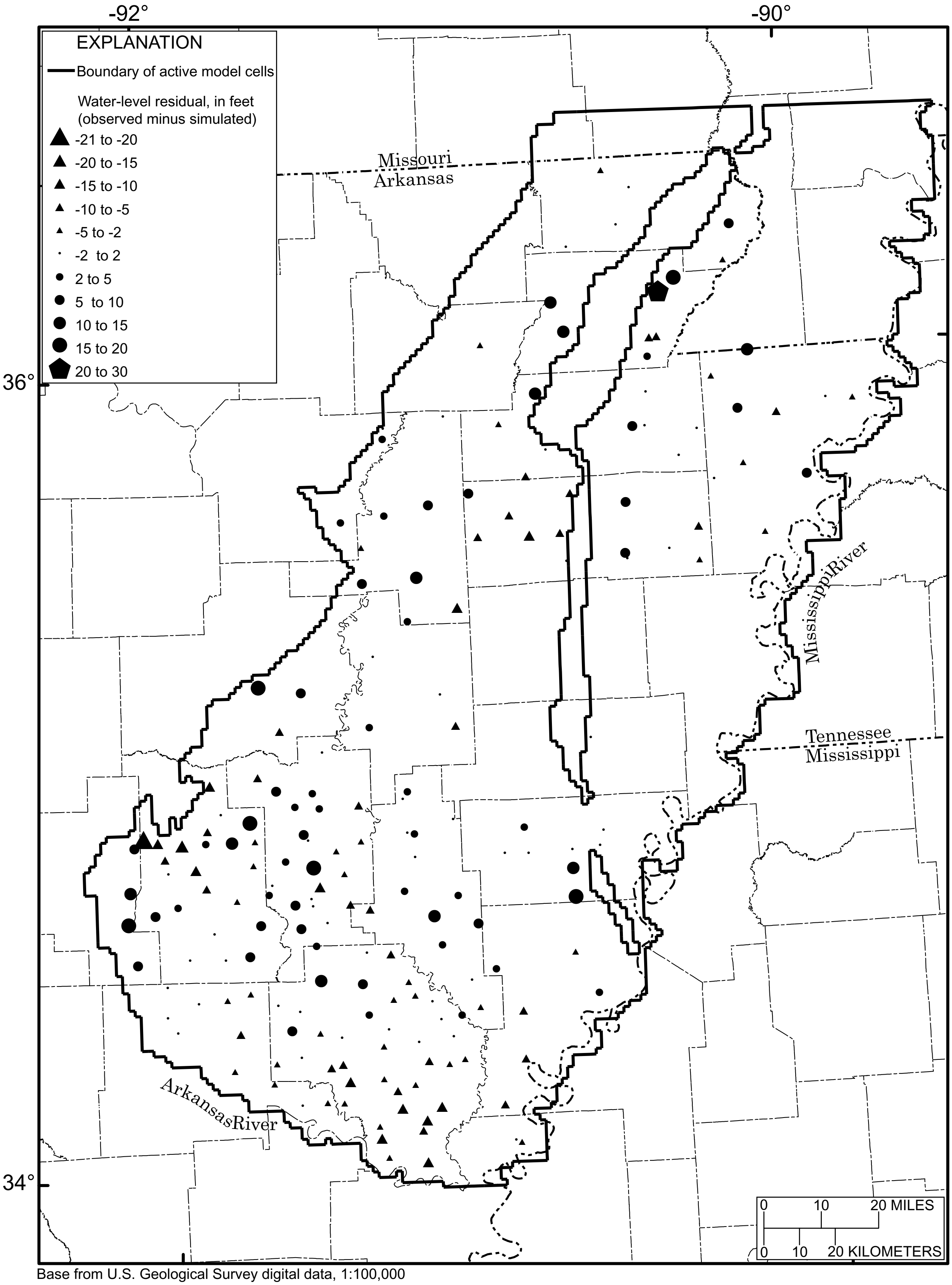


Figure 19. Water-level residuals for observation wells at the end of 1982 (stress period 7).

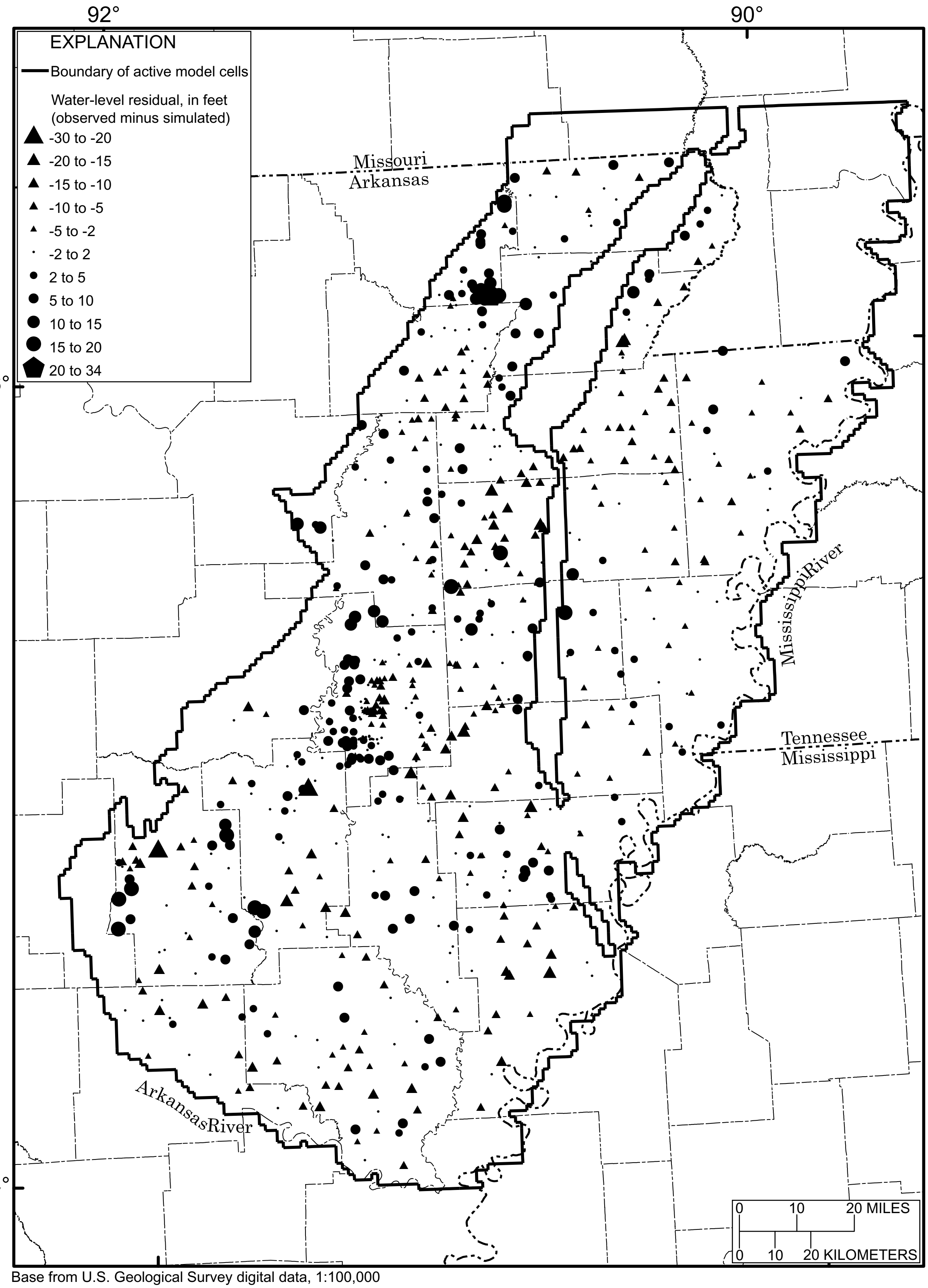
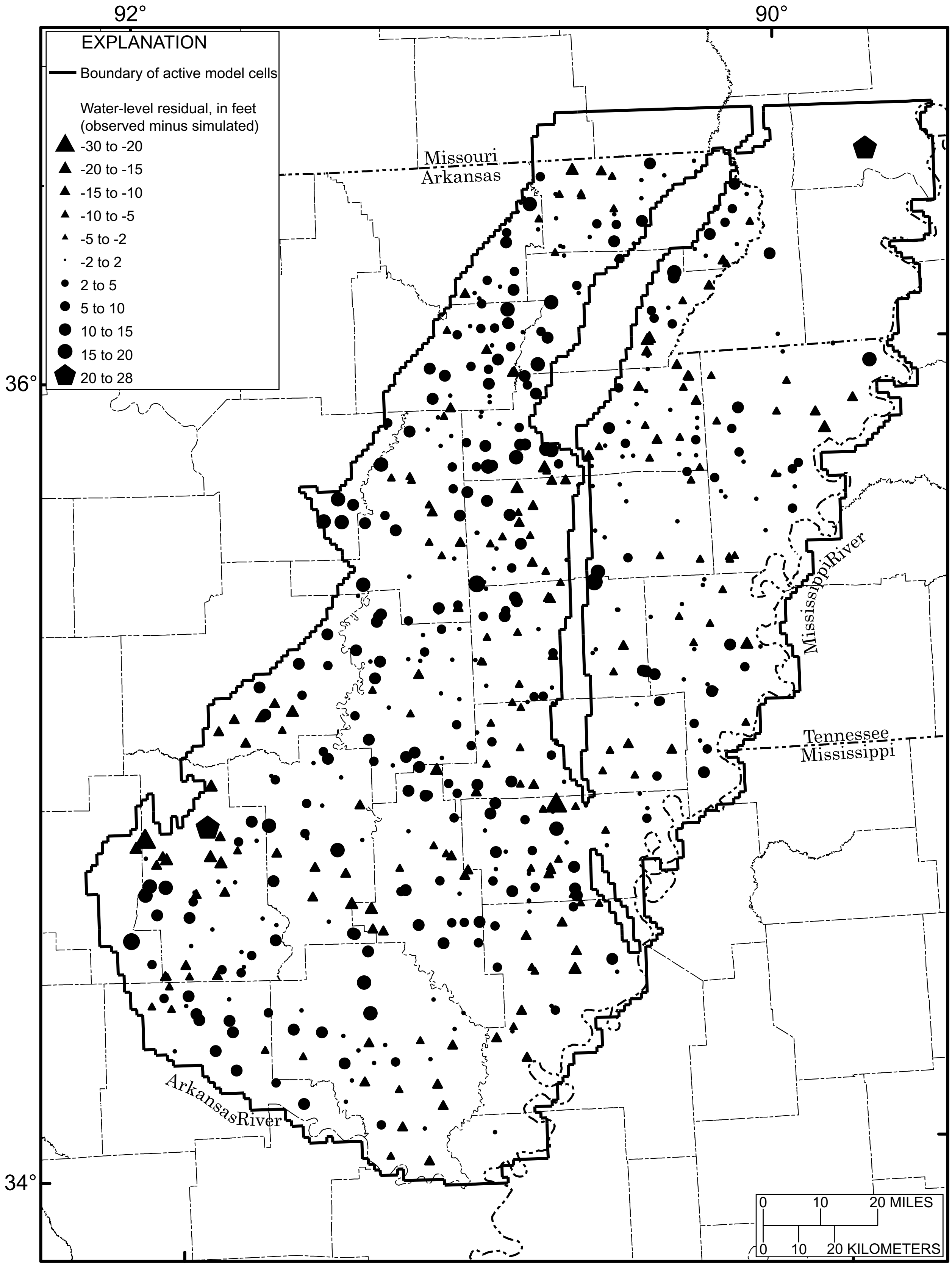


Figure 20. Water-level residuals for observation wells on March 31, 1992 (stress period 9).



Base from U.S. Geological Survey digital data, 1:100,000

Figure 21. Water-level residuals for observation wells on March 31, 1998 (stress period 11).

Table 3. Statistics for residuals for water-level observations

[Residuals are calculated as observed heads minus simulated heads]

										Mean absolute residuals					
Stress period number	Date range	Minimum (feet)	Maximum (feet)	Range (feet)	Mean (feet)	Median range (feet)	Root-mean square error (feet)	Standard error (feet)	Standard deviation (feet)	Residuals for all cells		Residuals for cells with rising water levels		Residuals for cells with falling water levels	
										Number of obser-vations	Mean (feet)	Number of obser-vations	Mean (feet)	Number of obser-vations	Mean (feet)
5	1968 to 1972	-23.77	12.74	36.51	-1.63	-2.28	5.79	0.37	5.56	223	4.69	150	4.70	73	4.68
7	1978 to 1982	-20.43	29.67	50.10	-0.05	-0.64	7.20	0.52	7.21	195	5.44	108	4.96	87	6.04
9	1989 to March 1992	-27.73	34.32	62.05	-0.72	-0.95	6.43	0.25	6.40	678	4.89	382	4.98	296	4.78
11	1994 to March 1998	-28.74	27.95	56.69	0.55	0.60	6.04	0.25	6.00	558	4.69	254	4.42	304	4.92

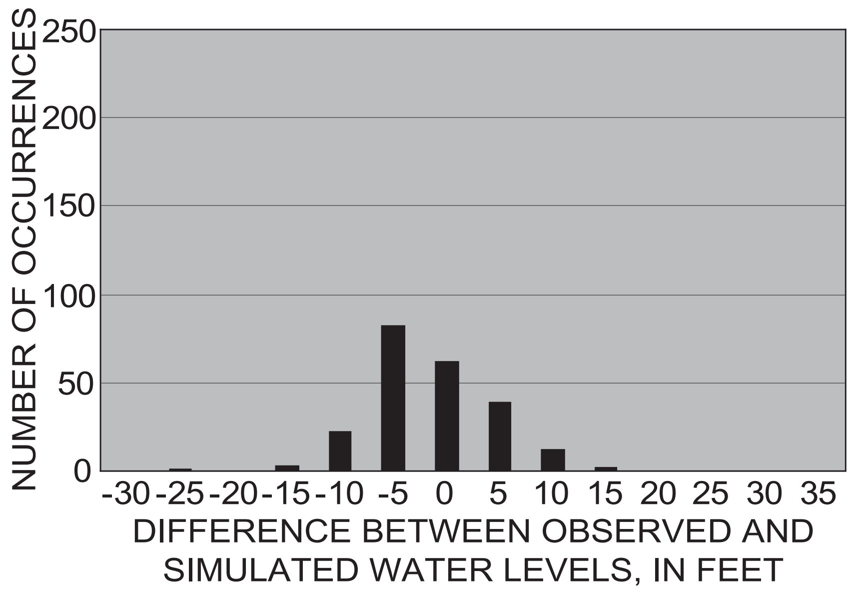


Figure 22. Difference between observed and simulated water levels from 1968 to 1972 (stress period 5).

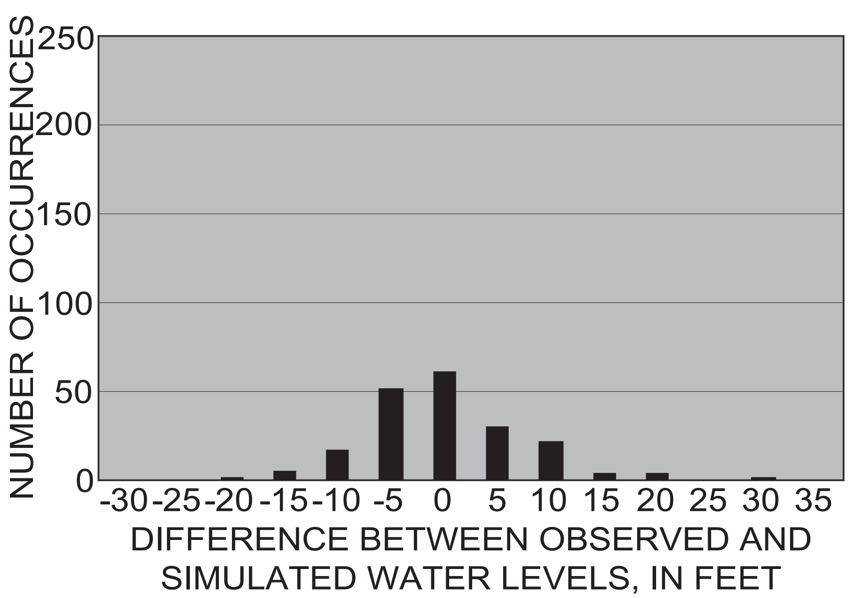


Figure 23. Difference between observed and simulated water levels from 1978 to 1982 (stress period 7).

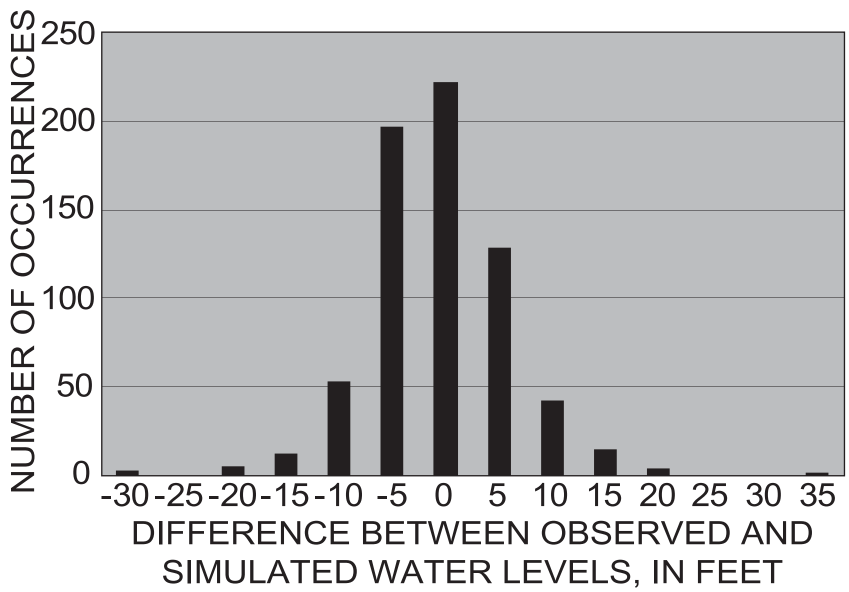


Figure 24. Difference between observed and simulated water levels from 1989 to March 1992 (stress period 9).

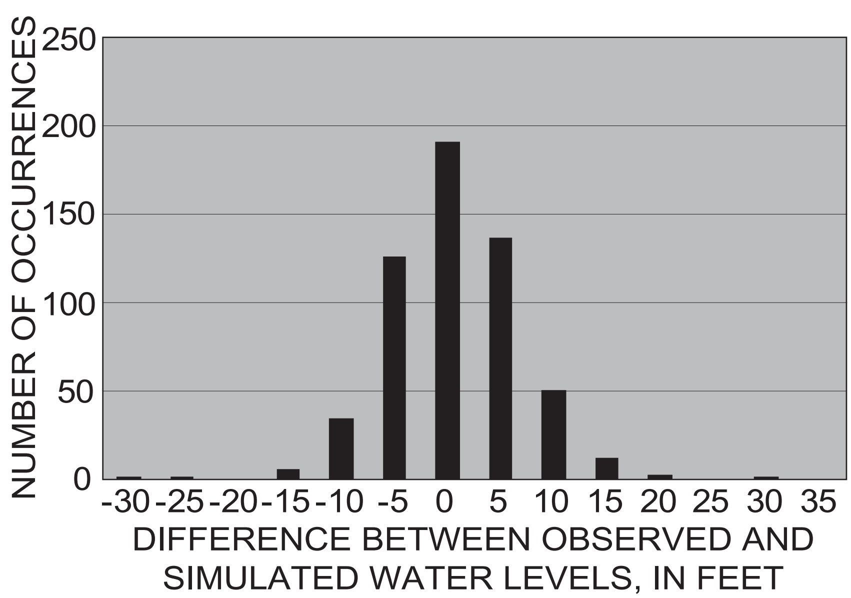


Figure 25. Difference between observed and simulated water levels from 1994 to March 1998 (stress period 11).

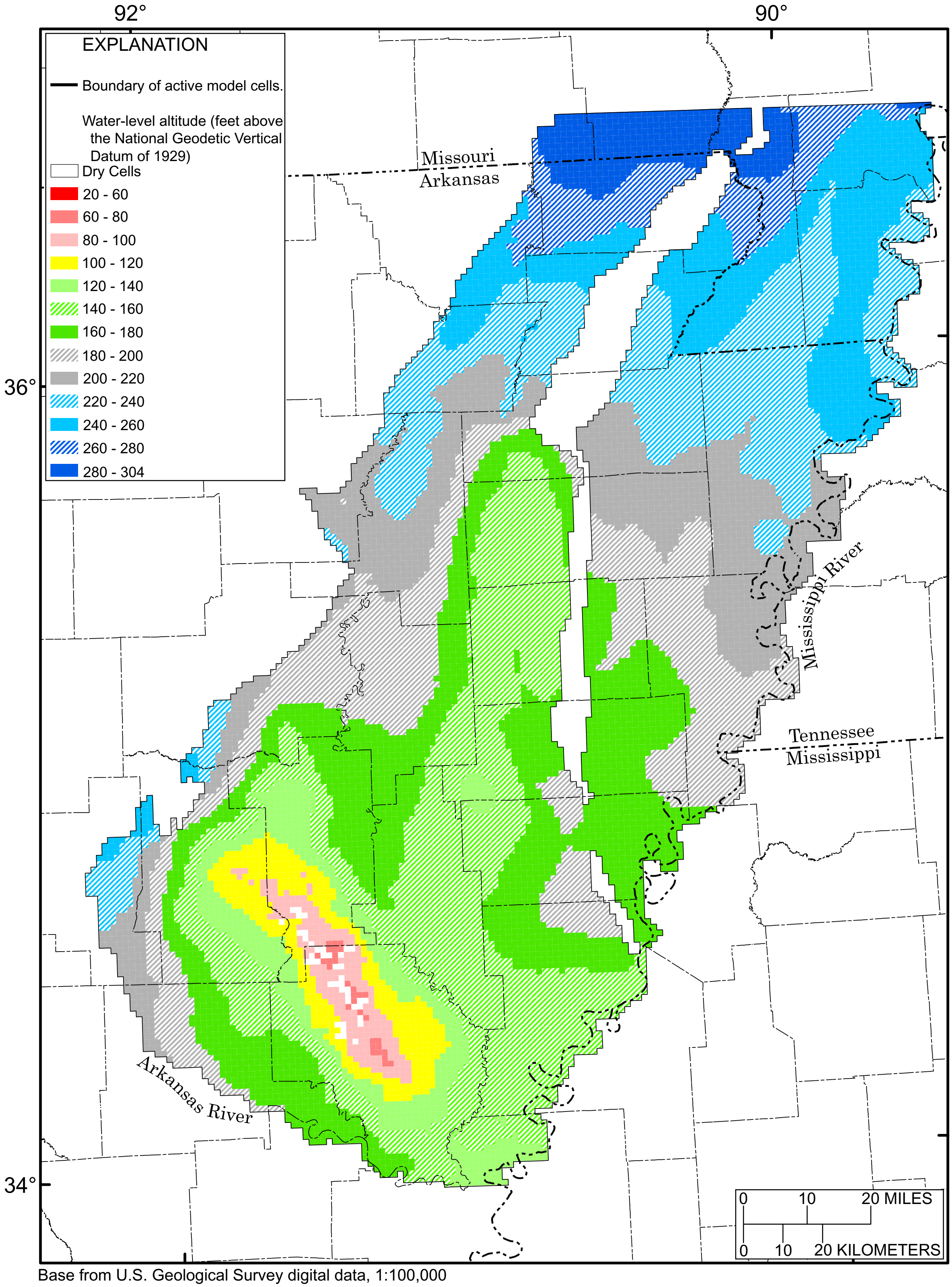


Figure 26. Simulated water-level altitudes for March 31, 1998 (stress period 11).

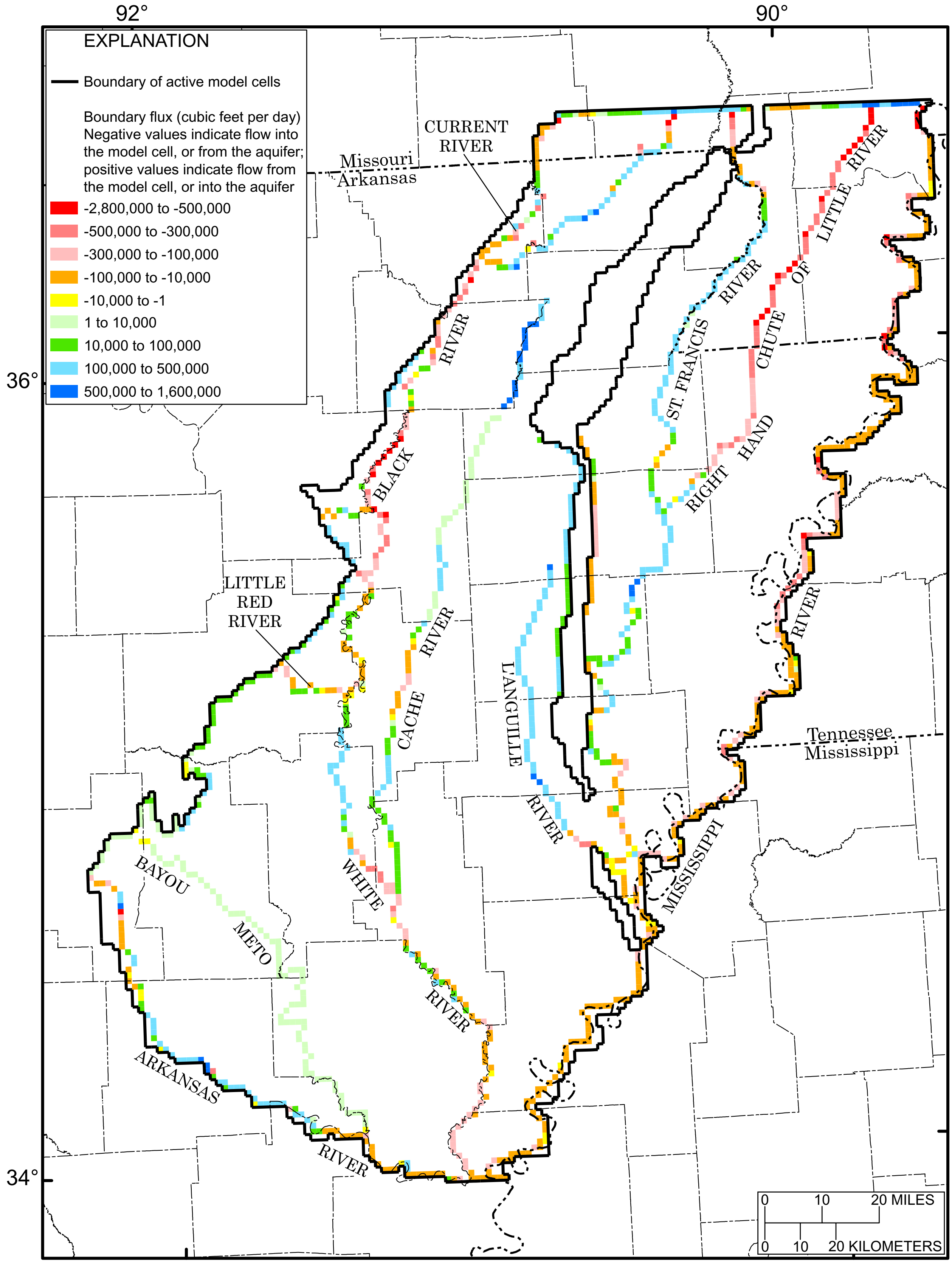
Flux values associated with river cells, specified-head cells, and general-head boundary cells are shown in figure 27. Red and yellow colors (negative values) indicate flows into the model cell from the aquifer; blue and green colors indicate flow from the model cell (into the aquifer). Except for the Mississippi River, the White River, and the Right Hand Chute of Little River, flow is generally from the rivers into the aquifer. Flow is generally eastward and southward from consolidated deposits to the west and southward from the aquifer in Missouri. The river fluxes shown in figure 27 are about 81,800,000 ft^3/d into the aquifer and 111,000,000 ft^3/d out of the aquifer. The Mississippi River accounts for about 37,600,000 ft^3/d, almost all out of the aquifer, and many of the other rivers account for similar amounts of river flux. Knowledge gained during the calibration process indicates that the amount of river flux may be less sensitive to riverbed conductance than to the properties of surrounding aquifer material and local head relations.

Recharge rates changed both quantitatively and spatially from stress period 5 to stress period 11. Figure 28 shows fluxes from recharge into the aquifer from 1918 through 1972 (stress periods 1 to 5), and figure 29 shows fluxes from recharge into the aquifer from 1994 to March 1998 (stress period to 11). Figure 15 shows the total rate of recharge into the aquifer throughout the model is 141,500,000 ft^3/d for 1918 to 1972 (stress periods 1 to 5), and 529,400,000 ft^3/d from 1994 to March 1998 (stress period 11). These model fluxes relate to areally specified recharge rates of 0.0 to 30.0 in/yr. The recharge rates for the intervening stress periods from 1973 to 1993 (stress periods 6 to 10) were prorated on a cell-by-cell basis from the rates for stress periods 1 to 5 and the rates for stress period 11. This proration was initially weighted for time but later altered to reduce residuals. Figure 16 shows the increase in recharge fluxes through time with a small increase between stress periods 10 and 11.

While the approach of this model to simulating recharge was by direct application using the MODFLOW recharge package (recharge package approach), as explained in the Model Development and Calibration section of this report, the model of Mahon and Poynter (1993) simulated areal recharge by applying river cells throughout the model using the MODFLOW river package (river package approach). In both approaches, recharge increases with time by either direct application as with the recharge package approach or by increased head gradients resulting from decreased water-level altitudes in the aquifer as with the river package approach. Both approaches attempted to incorporate a number of processes relating to recharge to the aquifer. This included both recharge from precipitation at the land surface, from leakage from the numerous perennial and ephemeral streams covering the area, and from interactions with underlying aquifers caused by increased drawdown from pumpage. The river package approach has the advantage over the recharge package approach because recharge increases are simulated continuously into the future, while the recharge package approach does not. However, during the calibration process the river package approach was unable to effectively simulate the additional stress periods with the additional water-use data and observation-well data. For the recharge package approach, the rates for stress periods 5 and 11 were manually adjusted for the residuals of those periods and the rates for stress periods 7 and 9 were prorated. Nevertheless, reasonable residuals were obtained for stress periods 7 and 9 as well as for stress periods 5 and 11 with the recharge package approach.

The inability of the river package approach to effectively account for the additional water-use data and observation well data reflect the complex processes relating to recharge. Basically, increases in recharge are related to drawdown through a variety of linear and non-linear processes whose mechanism and spatial distribution are poorly understood. Water-level gradients between the alluvial aquifer and underlying aquifers are complex and over time may involve net flows both into and out of the alluvial aquifer which cannot be accounted for effectively with the river package approach. Also, the hydrologic processes involved with recharge infiltrating downward through the overlying clay cap are not well understood and the river package approach may not always be effective in simulating them. In addition, other factors relating to recharge such as the application of large amounts of water to the surface during irrigation and the use of possibly inadequate water-use data may render the river package approach less effective.

The recharge package approach did not provide for increases in recharge after stress period 11. The increase in the recharge rates level off between stress periods 10 and 11 as shown in figure 16. This at least provides a basis for no change in recharge after stress period 11, which is reasonably consistent with trends in the observation-well data, and provides a generally conservative approach with regard to aquifer simula-



Base from U.S. Geological Survey digital data, 1:100,000

Figure 27. Simulated river, specified head, and general head boundary fluxes for model cells from 1994 to March 1998 (stress period 11).

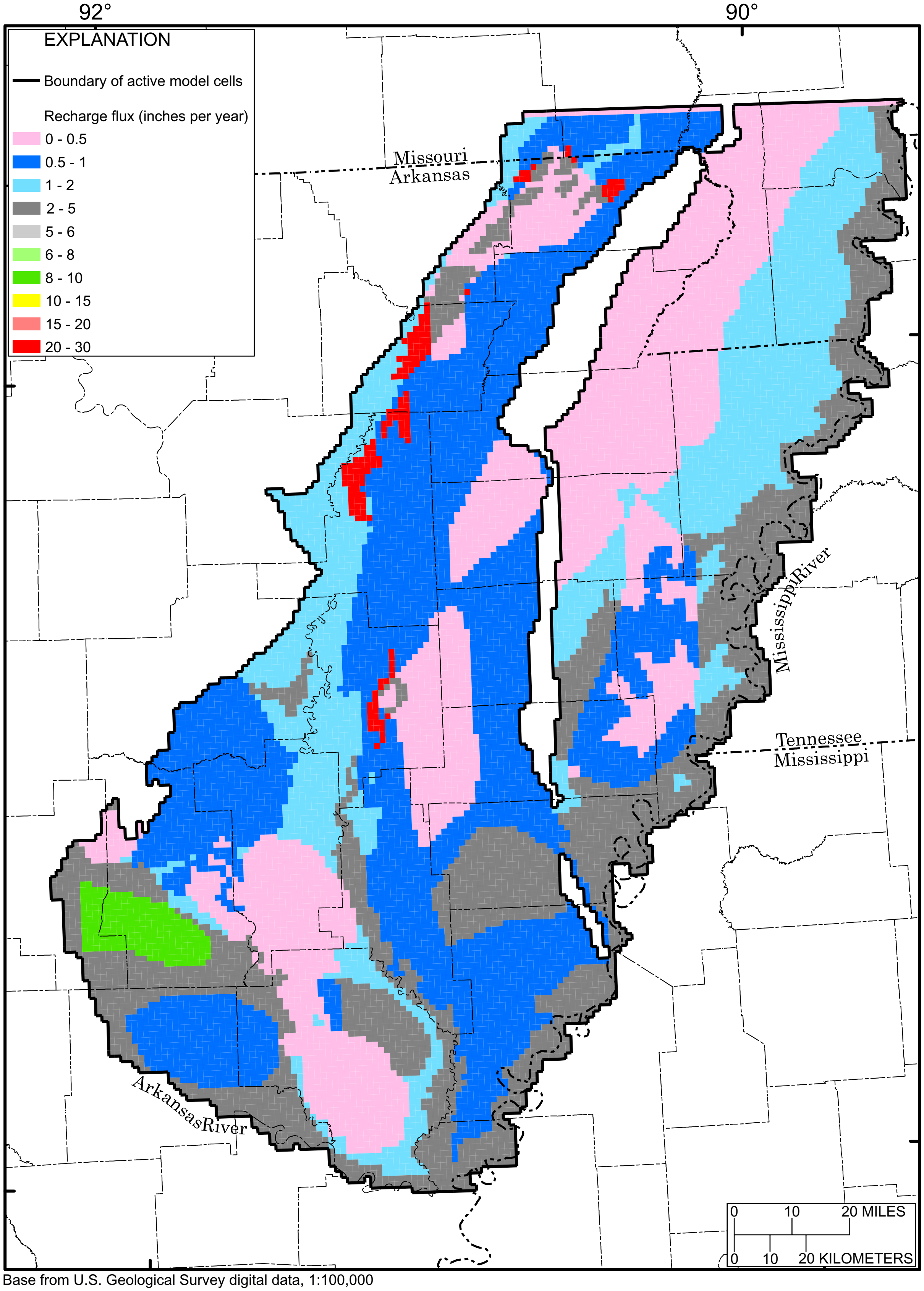


Figure 28. Simulated recharge fluxes for 1918 to 1972 (stress periods 1-5).

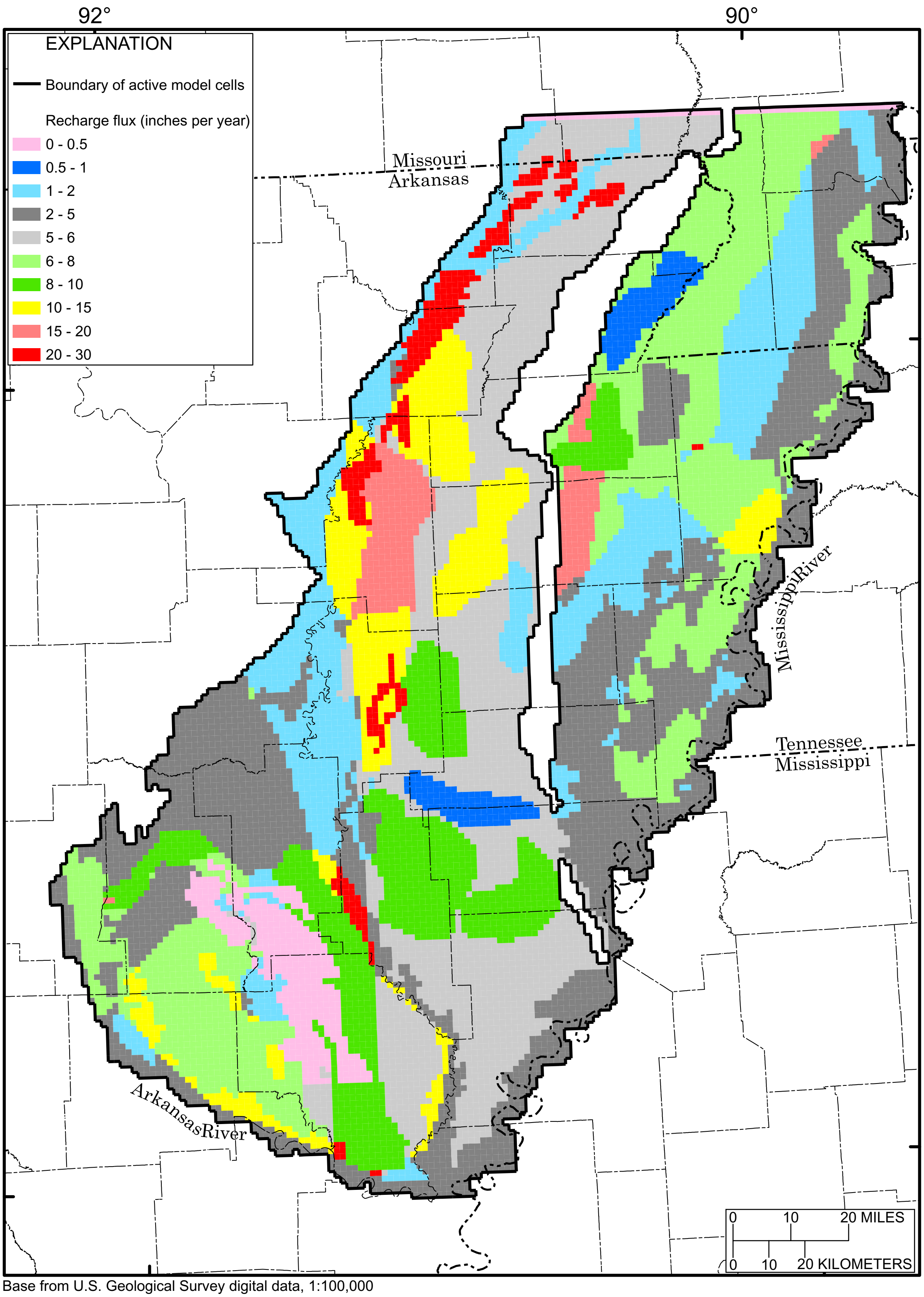


Figure 29. Simulated recharge fluxes for 1994 to March 1998 (stress period 11).

tion of hypothetical projected ground-water withdrawals after stress period 11 (discussed in the next section). As to whether recharge is actually stable after stress period 11, or whether it increases or perhaps even decreases, this model and the data based on it cannot provide an entirely reliable answer. Further studies and models may provide a better approach to characterizing recharge that can better account for recharge from the underlying aquifers and the hydrologic processes involved with recharge from the land surface and streams.

The ground-water flow model described in this report is useful in evaluating the flow system in the alluvial aquifer. However, model results are a simulation of the system and the following model limitations should be considered.

The model was discretized on a 1-mile grid and cannot be used to evaluate hydrologic conditions for a smaller scale setting such as those involving a single pumping well. Model input parameters are applied over entire cells and assumptions of uniformity for heterogeneous geologic materials and climatic conditions may introduce inaccuracies. The steady-state simulation used for initial head conditions for the assumed predevelopment conditions assumes that flows into and out of the alluvial aquifer were equal. If this were not so, the change in ground-water storage (that is, if water levels were actually rising or falling) would be a source of model error.

Further, the lumping of areal recharge from multiple sources (precipitation, flow to or from some rivers, and ground-water flow to or from underlying aquifers) did not allow each of these compounds to be evaluated independently during the modeling process. Also, estimates of ground-water use may not be accurate and as such may introduce error.

SIMULATIONS OF WATER LEVELS CAUSED BY PROJECTED GROUND-WATER WITHDRAWALS

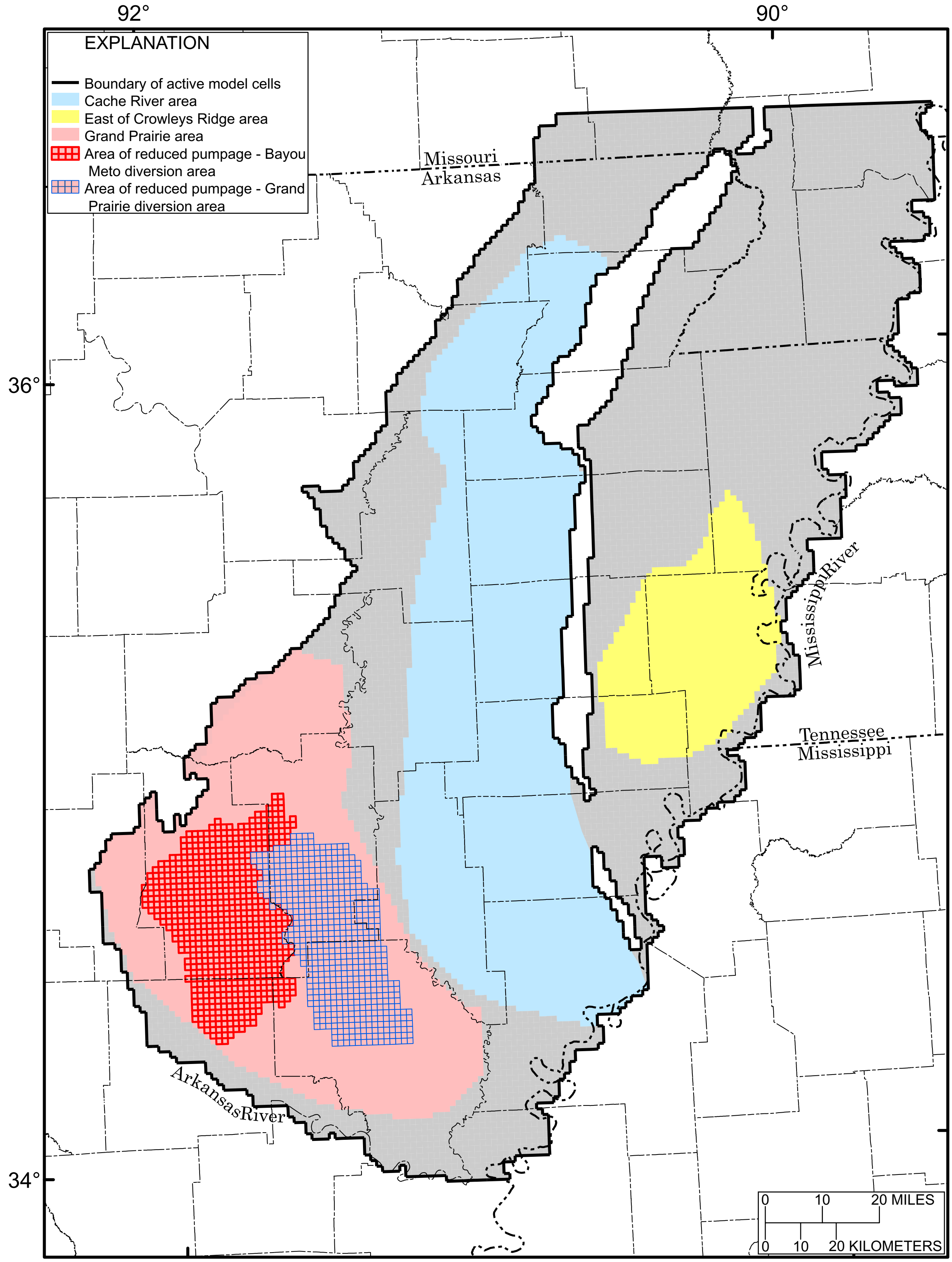
Water levels caused by projected ground-water withdrawals were simulated using the calibrated model. Simulations represented a period of 50 years into the future from 1998 through 2049 in three scenarios using different assumptions regarding future pumpage trends. As a visual aid, digital animations of the simulated scenarios are included on a compact disk in the back of the report. In Scenario 1, the 1997 pumpage rate is extended without change. For Scenario 2, water-use data from 1965 through 1995 (Halberg and Stephens, 1966; Halberg, 1972, 1977; Holland and Ludwig, 1981; Holland, 1987, 1993, 1999) were used with linear regression to establish historic trends in the rate of increase for each county and subsequently used to determine the estimated pumpage for each county for 2000. This trend was applied to each well in a county in proportion to its portion of the 1997 pumpage of that county to supply increased pumpage for successive stress periods. This was applied to successive stress periods to a maximum value of 1.25 times the estimated pumpage for each county in 2000. In Scenario 3, pumpage is decreased to 90 percent of that applied in Scenario 2 for each future stress period in two selected areas—the Bayou Meto diversion area (Natural Resources Conservation Service, 2003a) and in the Grand Prairie diversion area (Natural Resources Conservation Service, 2003b) (fig. 30)— to account for possible supplementation of ground water for irrigation by withdrawals from surface-water diversion projects by the U.S. Army Corps of Engineers and other parties in those areas. Table 4 shows the actual total pumpages used for these simulations. As cells went dry during these simulations the pumpage simulated within the cell was set to zero. Simulated water levels (figs. 31-33) and saturated thicknesses (figs. 34 and 35-36) were calculated for 10-year intervals to show the effect of pumpage on the aquifer during each scenario.

Table 4. Actual total pumpage for projection simulations

[Units are in cubic feet per day]

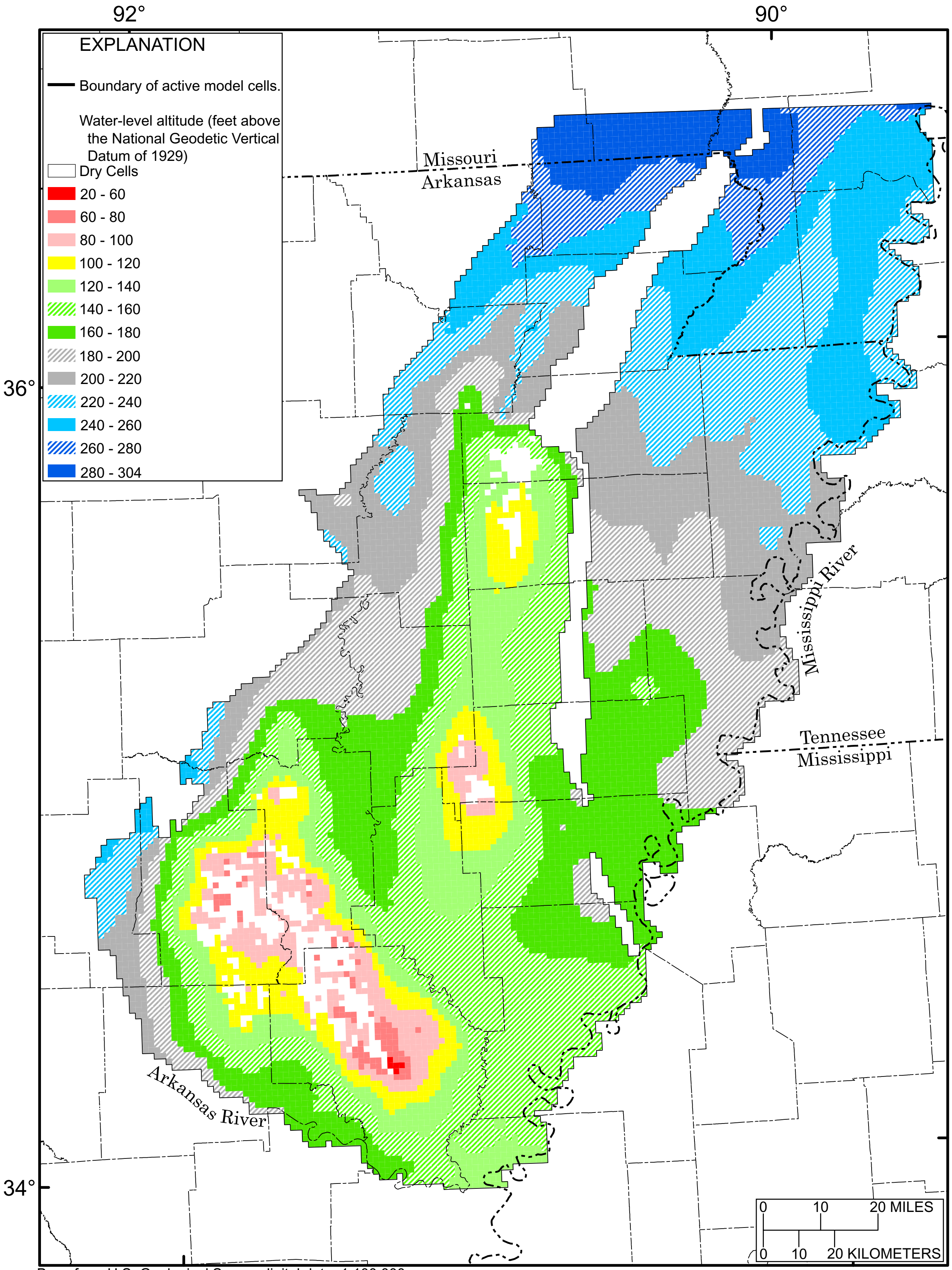
Stress period	With 1997 pumpage extended (Scenario 1)	With 1997 pumpage extended by county trends (Scenario 2)	Pumpage reduced to 90 percent of pumpage extended by county trends in selected zones (Scenario 3)
11	631,324,480	631,324,480	631,324,480
12	623,440,512	749,344,192	744,711,936
13	614,197,568	835,934,976	844,568,640
14	605,206,656	779,753,792	782,530,496
15	596,392,384	759,117,632	761,897,344
16	590,821,760	744,969,536	746,841,088

In each scenario water levels declined resulting in dry areas (aquifer depletion) by 2049 (figs. 34 and 36-37). These dry areas generally were distributed in the Grand Prairie area between the Arkansas River and White River and along the Cache River.



Base from U.S. Geological Survey digital data, 1:100,000

Figure 30. Selected zones used for pumping changes in model projection simulations.



Base from U.S. Geological Survey digital data, 1:100,000

Figure 31. Simulated water levels at the end of 2049 (stress period 16) with 1997 pumpage extended unchanged (scenario 1).

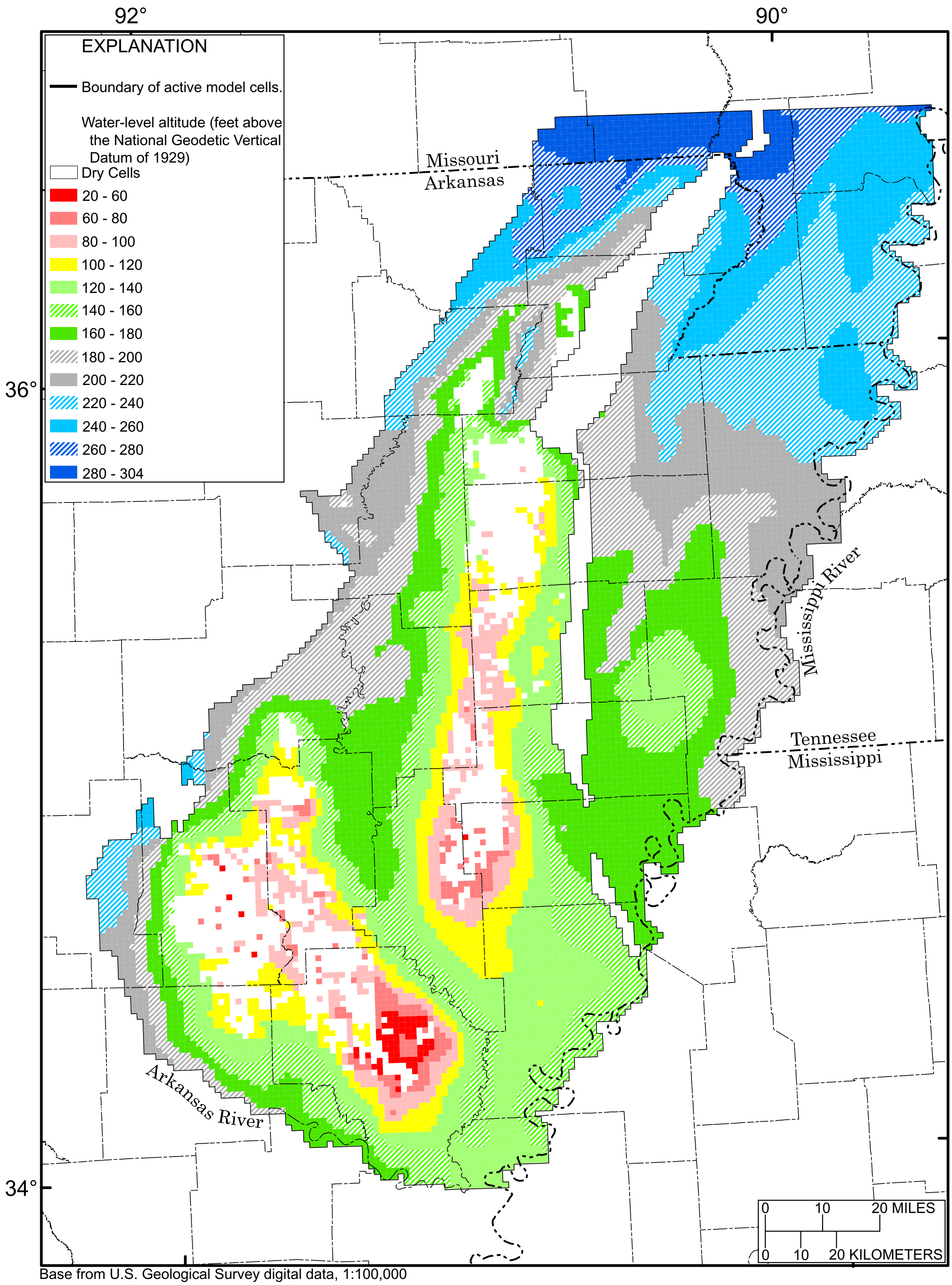


Figure 32. Simulated water levels at the end of 2049 (stress period 16) with 1997 pumpage extended by county trends (scenario 2).

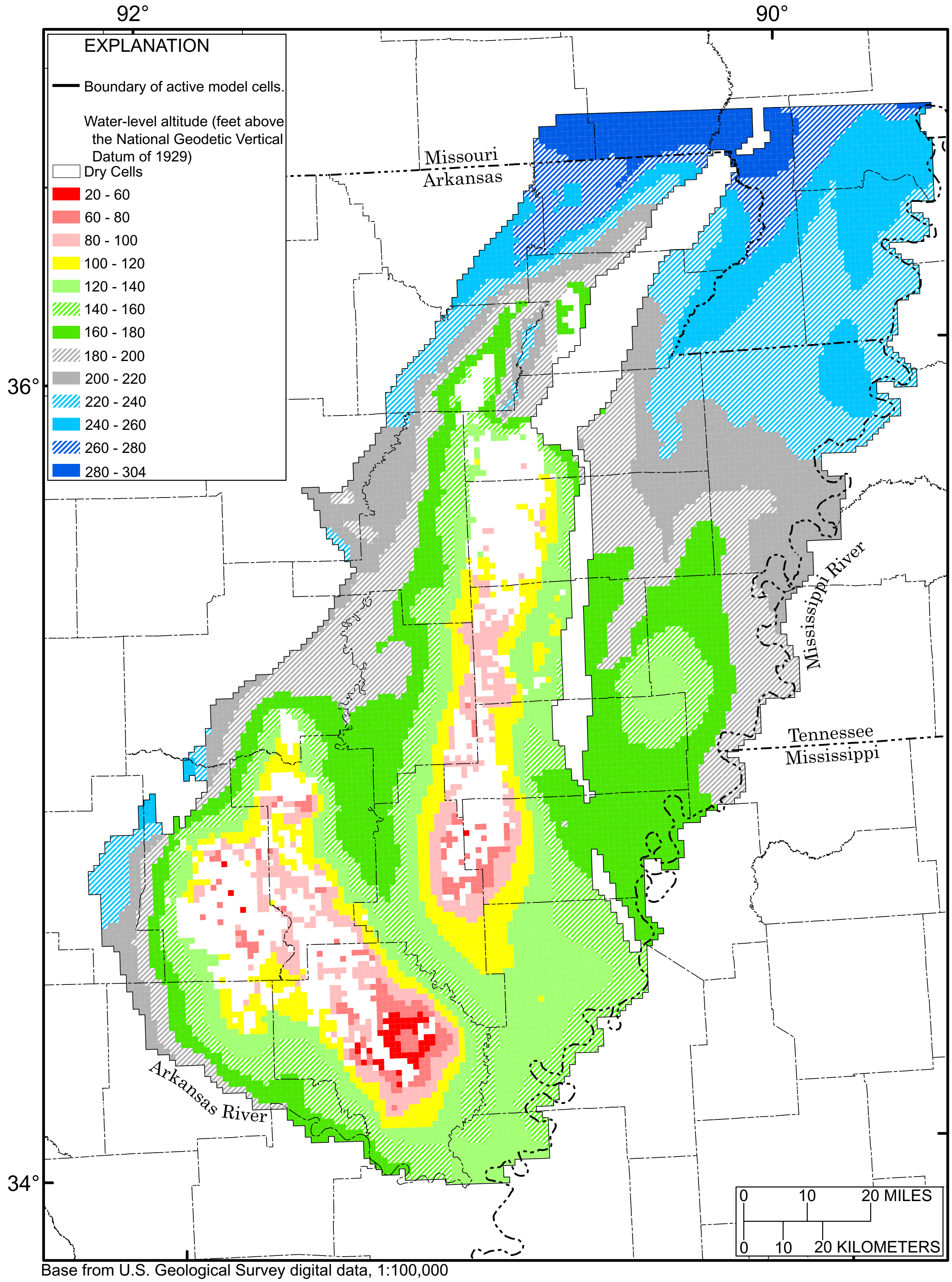


Figure 33. Simulated water levels at the end of 2049 (stress period 16) with pumpage reduced to 90 percent of county trends in selected zones (scenario 3).

Click here to open figure 34.

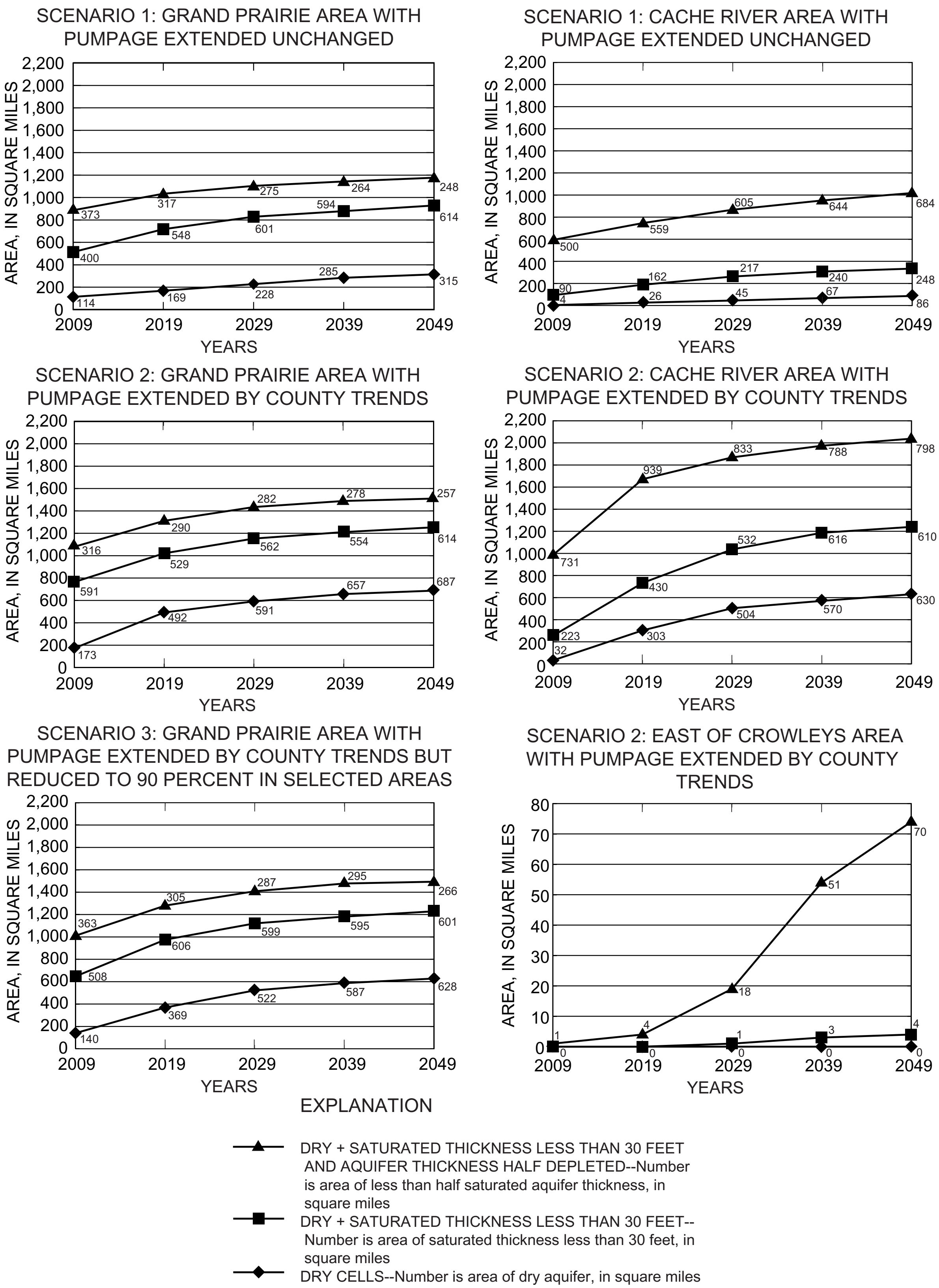


Figure 35. Areas of extreme water-level declines produced by model projection scenarios.

Continuation of pumpage at 1997 rates through 2049 (Scenario 1) results in simulated cones of depression between the Arkansas River and the White River in the Grand Prairie area and in the Cache River area (figs. 31 and 34). The simulated dewatering results in model cells covering about 300 mi^2 going dry in the Grand Prairie area and about 100 mi^2 going dry in the Cache River area on the west side of Crowleys Ridge by 2049. Areas of extreme water-level decline include areas where the model cells have gone dry and areas where the water level in the aquifer is equal to or less than 50 percent of the original saturated thickness, assuming confined conditions everywhere in the aquifer in predevelopment times. The latter areas are hereafter referred to as “half-saturated” or “areas of half-saturation.” Areas of extreme water-level decline have increased over time in the Grand Prairie area and in the Cache River area (figs. 34 and 35). Dry cells and areas of half-saturation, shown in brighter colors in figures 33 and 36-37 increase in area most rapidly between 2009 and 2019 (fig. 35) and the areas of dry cells and cells with 30 or less feet of saturated thickness extend westward in the Grand Prairie area and also extend southward from the northern part of the Cache River area (west sides of Poinsett and Craighead Counties). The rate of increase of extreme water-level declines decreases slightly between 2019 and 2049 (fig. 35), but these affected areas continue to extend westward in the Grand Prairie area and extend southward from the northern part of the Cache River area and northward from the southern part of the Cache River area (intersections of Monroe, Lee, and St. Francis Counties). The increase in dry cells is greater in the Grand Prairie area than in the Cache River area, while there is a greater increase in half-saturation in the Cache River area than in the Grand Prairie area (fig. 35).

Increasing the pumping rates to that which would be projected using historic water-use data (Scenario 2) results in simulated cones of depression between the Arkansas River and the White River in the Grand Prairie area and in the Cache River area that are substantially larger than in Scenario 1 (figs. 32 and 36). The extreme water-level declines in the Grand Prairie area are very extensive with cells covering nearly 700 mi^2 going dry by 2049; this is more than twice the number of dry cells resulting from pumpage continuing at the 1997 rate (Scenario 1). In the Cache River area, extreme water-level declines are more extensive than in Scenario 1, with cells covering over 600 mi^2 dewatered by 2049; this is approximately seven times the number of dry cells resulting from Scenario 1 (fig. 35). The areas of dry cells in the northern and southern parts of the Cache River area nearly coalesce into a single area of dry cells (fig. 36). An area of half-saturation east of Crowleys Ridge at the intersection of Crittenden, Cross, and St. Francis Counties also grows substantially to about 70 mi^2 by 2049 (fig. 36) in Scenario 2, although no cells go dry.

By as early as 2019, simulated conditions resulting from Scenario 2 result in more aquifer dewatering than conditions in 2049 resulting from Scenario 1. Most of the Grand Prairie area is half-saturated and approximately 500 mi^2 of dry cells are simulated (fig. 36). The northern and southern parts of the Cache River area cover approximately 300 mi^2 of dry cells and are connected by half-saturated cells (fig. 36).

As in Scenario 1, half-saturated cells increase in area most rapidly between 2009 and 2019 (fig. 37). These rates of increases slow somewhat between 2019 and 2049.

Pumpage in Scenario 3 is increased at the rate of historic trends, except in two selected areas (fig. 30) in the Grand Prairie area where pumpages were reduced to 90 percent of that used in Scenario 2 owing to surface-water diversion projects by the U.S. Army Corps of Engineers and other parties (Scenario 3), and simulates cones of depression between the Arkansas River and the White River in the Grand Prairie area and in the Cache River area that are substantially larger than in Scenario 1 (figs. 33 and 37). Water levels and half-saturation areas outside of the Grand Prairie area containing the two selected areas differ little from those in Scenario 2. However, water levels and areas of half-saturation within the Grand Prairie area also are little different from water levels in Scenario 2 (figs. 32 and 36). Also, areas of extreme water-level decline also decrease little in the Grand Prairie area compared to Scenario 2 (fig. 35).

The extreme water-level declines of the Grand Prairie area again is very extensive with cells covering more than 600 mi^2 going dry by 2049 (fig. 37); this is approximately twice the area resulting from pumpage continuing at the 1997 rate (Scenario 1), but is approximately 60 mi^2 less than in Scenario 2 (fig. 35). In the Cache River area, extreme water-level declines are similar to Scenario 2 and much more extensive than in Scenario 1, with cells covering over 600 mi^2 dry by 2049 (figs. 35 and 37) and with the areas of dry cells in the northern and southern parts nearly coalescing into a single area of nearly continuous dry cells (fig. 37). As

Click here to open figure 36.

Click here to open figure 37.

in Scenario 2 the area of half-saturation east of Crowleys Ridge at the intersection of Crittenden, Cross, and St. Francis Counties also grows substantially to about 70 mi^2 by 2049 (figs. 35 and 37) although no cells go dry.

The relatively small differences between the Scenario 2 and Scenario 3 measures of water-level decline in 2049 indicate that a 10 percent reduction of pumpage from the historic rate within the Bayou Meto and Grand Prairie diversion areas would do little to improve water levels in those areas. However, these reductions in pumpage (which in 2049 would be about 12 million ft^3/d) are small, about 16 percent, relative to the reductions that could result (about 76 million ft^3/d; Ken Bright, U.S. Army Corps of Engineers, written. commun., 2003) from implementation of these diversion projects.

Model results from the scenarios indicate that at 1997 pumpage rates water is being withdrawn from the aquifer at rates greater than can be sustained for the long term.

SUMMARY

The Mississippi River Valley alluvial aquifer lies in a broad alluvial plain that encompasses an area of about 32,000 mi^2 in parts of six states, more than 54 percent occurring in eastern Arkansas. The alluvial aquifer in Arkansas is bounded on the west by consolidated formations of Paleozoic age that have a distinctly smaller hydraulic conductivity than unconsolidated deposits of the alluvial aquifer, and on the east by the Mississippi River. The alluvial sediments of eastern Arkansas are composed of a sequence of sands, silts, and clays that can be divided vertically into two units from a regional perspective: the upper unit consisting of a confining unit of clay, silt, and fine sand, and the lower unit containing the alluvial aquifer, which contains coarse sand and gravel. In northeastern Arkansas, the alluvial aquifer is divided by Crowleys Ridge into two hydraulically separate flow regimes. Ground water flows mainly from the north and the west southward and eastward towards the Mississippi River or into local cones of depression.

The alluvial aquifer is the most prolific aquifer supplying water to eastern Arkansas. Surface water is abundant in the region, and conjunctive use of surface water with ground water is a key element to assuring sustainable use of the alluvial aquifer. The effects of current and future ground water on water availability concern water managers and users; a full understanding of the behavior of the aquifer under various water-use scenarios is critical to development of viable water-management and alternative source plans.

A finite-difference, three-dimensional, ground-water flow model modified from an earlier model (Mahon and Poynter, 1993) was used to simulate ground-water flow in the alluvial aquifer for the period 1918 to 2049. The aquifer was divided vertically into two layers and horizontally into a grid of 184 by 156 cells each measuring 1 mi on a side. The lateral boundaries consist of the major rivers on the east and south. Specified heads were applied on the north, and general-head boundaries on portions of the west and along parts of Crowleys Ridge along with no flow boundaries.

Hydraulic conductivity values obtained during the calibration process were 230 ft/d for the upper layer and ranged from 230 to 730 ft/d for the lower layer with the maximum mean for the combined aquifer of about 480 ft/d. Specific yield values were 0.30 throughout the model, and specific storage values were 0.000001 inverse-feet throughout the model. Total recharge rates for 1973 to 1977 were about 141,500,000 ft^3/d, and from 1994 to March 1998 were about 529,500,000 ft^3/d. These total recharge rates relate areally distributed recharges from 0 to 30 in/yr, and total recharge increased from 1972 to 1998 by a factor of about four.

Residuals of simulated versus observed water levels had a root-mean-square error of 6.04 ft and an absolute mean value of 4.69 ft for the period ending in 1998 (stress period 11) and an absolute mean value of less than 6 ft for three earlier stress periods. The overall distribution of these residuals resembled a normal distribution for all stress periods. Spatial distribution of residuals generally was random for stress period 11 for simulated water levels higher or lower than observed water levels. For three earlier stress periods, observed water levels tended to be lower than simulated water levels.

Water levels caused by projected ground-water withdrawals were simulated using the calibrated model. Simulations represented a period of 50 years into the future in three scenarios with either unchanged 1997 pumpage, pumpage increased by historic trends, or pumpage increased by historic trends except in two areas of the Grand Prairie. All scenarios produced extreme water-level declines (areas where model cells have gone dry or where the (aquifer is half-saturated). Two areas of the aquifer covering a total of about 400 mi^2 go dry if pumping remains at 1997 rates. One of these areas is located between the Arkansas and White Rivers and the other west of Crowleys Ridge along the Cache River. Increasing the pumping rates to that which would be projected using historic data led to increased extreme water-level declines in both areas with about 1,300 mi^2 going dry. Water-level declines in both scenarios generally occurred most rapidly between 2009 and 2019. Reducing the pumping rates to 90 percent of that used for projected historic rates in areas of the Grand Prairie relating to two diversion projects by the U.S. Army Corps of Engineers and other parties did little to decrease the extreme water-level declines. However, these pumpage reductions are small (amounting to about 16 percent of the reductions that could result from implementation of these diversion projects). Model results from the scenarios indicate that at 1997 pumpage rates water is being withdrawn from the aquifer at rates greater than can be sustained for the long term.

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