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Scientific Investigations Report 2004–5102

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

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Vertical Datum

Vertical coordinate information is referenced to the National Geodetic Vertical Datum of 1929 (NGVD 29).

Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83).

Notes on units, hydraulic conductivity and transmissivity:

Hydraulic conductivity: The standard unit for hydraulic conductivity is cubic foot per day per square foot of aquifer cross-sectional area, $(ft^3/d)/ft^2$. In this report, the mathematically reduced form, foot per day (ft/d), is used for convenience.

Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness, $[(ft^3/d)/ft^2]ft$. In this report, the mathematically reduced form, foot squared per day (ft^2/d), is used for convenience.

Hydrogeology and Simulation of Ground-Water Flow and Land-Surface Subsidence in the Northern Part of the Gulf Coast Aquifer System, Texas

By Mark C. Kasmarek and James L. Robinson

Abstract

As a part of the Texas Water Development Board Ground-Water Availability Modeling program, the U.S. Geological Survey developed and tested a numerical finite-difference (MODFLOW) model to simulate ground-water flow and land-surface subsidence in the northern part of the Gulf Coast aquifer system in Texas from predevelopment (before 1891) through 2000. The model is intended to be a tool that water-resource managers can use to address future ground-water-availability issues.

From land surface downward, the Chicot aquifer, the Evangeline aquifer, the Burkeville confining unit, the Jasper aquifer, and the Catahoula confining unit are the hydrogeologic units of the Gulf Coast aquifer system. Withdrawals of large quantities of ground water have resulted in potentiometric-surface (head) declines in the Chicot, Evangeline, and Jasper aquifers and land-surface subsidence (primarily in the Houston area) from depressurization and compaction of clay layers interbedded in the aquifer sediments. In a generalized conceptual model of the aquifer system, water enters the ground-water-flow system in topographically high outcrops of the hydrogeologic units in the northwestern part of the approximately 25,000-square-mile model area. Water that does not discharge to streams flows to intermediate and deep zones of the system southeastward of the outcrop areas where it is discharged by wells and by upward leakage in topographically low areas near the coast. The uppermost parts of the aquifer system, which include outcrop areas, are under water-table conditions. As depth increases in the aquifer system and as interbedded sand and clay accumulate, water-table conditions evolve into confined conditions.

The model comprises four layers, one for each of the hydrogeologic units of the aquifer system except the Catahoula confining unit, the assumed no-flow base of the system. Each layer consists of 137 rows and 245 columns of uniformly spaced grid blocks, each block representing 1 square mile. Lateral no-flow boundaries were located on the basis of outcrop extent (northwestern), major streams (southwestern, northeastern), and downdip limit of freshwater (southeastern). The MODFLOW general-head boundary package was used to simulate recharge and discharge in the outcrops of the hydrogeologic units. Simulation of land-surface subsidence (actually, compaction of clays) and release of water from storage in the clays of the Chicot and Evangeline aquifers was accomplished using the Interbed-Storage Package designed for use with the MODFLOW model. The model was calibrated by trial-and-error adjustment of selected model input data in a series of transient simulations until the model output (potentiometric surfaces, land-surface subsidence, and selected water-budget components) reasonably reproduced field measured (or estimated) aquifer responses.

Model calibration comprised four elements: The first was qualitative comparison of simulated and measured heads in the aquifers for 1977 and 2000; and quantitative comparison by computation and areal distribution of the root-mean-square error between simulated and measured heads. The second calibration element was comparison of simulated and measured hydrographs from wells in the aquifers in a number of counties throughout the modeled area. The third calibration element was comparison of simulated water-budget components—primarily recharge and discharge—to estimates of physically reasonable ranges of actual water-budget components. The fourth calibration element was comparison of simulated land-surface subsidence from predevelopment to 2000 to measured land-surface subsidence from 1906 through 1995.

The simulated potentiometric surfaces of the Chicot, Evangeline, and Jasper aquifers for 1977 and 2000 show general agreement with measured potentiometric surfaces (or with measured point head data in areas where data are sparse). The root-mean-square errors for the aquifer potentiometric surfaces for 1977 were about 34 feet for the Chicot aquifer, about 43 feet for the Evangeline aquifer, and about 47 feet for the Jasper aquifer. The errors are about 7, 8, and 17 percent, respectively, of the total range in measured heads for the three aquifers. For

the 2000 potentiometric surfaces the root-mean-square errors were about 31 feet for the Chicot aquifer, about 40 feet for the Evangeline aquifer, and about 34 feet for the Jasper aquifer. The errors are about 8, 6, and 11 percent, respectively, of the total range in measured heads for the respective aquifers. Twenty-one pairs of simulated and measured hydrographs for the three aquifers match with varying degrees of closeness. For hydrographs in which the match between simulated and measured heads is less close than others, the trends in simulated and measured heads generally are similar.

For calibrated 1977 conditions, 757 cubic feet per second of recharge plus 742 cubic feet per second from depletion of sand storage plus 340 cubic feet per second from inelastic compaction of clays is approximately offset by 169 cubic feet per second of natural discharge and 1,670 cubic feet per second (1,080 million gallons per day) of withdrawals. Thus in 1977, net recharge supplied about 35 percent of withdrawals, depletion of sand storage about 45 percent, and inelastic compaction of clays about 20 percent. For calibrated 2000 conditions, 965 cubic feet per second of recharge plus 410 cubic feet per second from depletion of sand storage plus 106 cubic feet per second from inelastic compaction of clays is approximately offset by 161 cubic feet per second of natural discharge and 1,322 cubic feet per second (854 million gallons per day) of withdrawals. Thus in 2000, net recharge supplied 61 percent of withdrawals, depletion of sand storage 31 percent, and inelastic compaction of clays 8 percent. The most notable differences between the simulated water-budget components of 1977 and 2000, besides the fact withdrawals were about 21 percent less in 2000, are the increase in the percentage of withdrawals supplied by recharge and the decrease in the percentage of water supplied by depletion of storage and inelastic compaction of clays between 1977 and 2000.

The match between simulated and measured land-surface subsidence from predevelopment to near present day in the Harris-Galveston-Fort Bend County area, where compaction of subsurface material, and thus subsidence, has been monitored continuously since the 1970s, is close. As much as 10 feet of subsidence has occurred in southeastern Harris County near the northern end of Galveston Bay. Away from the Harris-Galveston-Fort Bend County area, subsidence of as much as 3 feet was simulated in the Evadale-Beaumont withdrawal area in southwestern Jasper County. No subsidence was simulated in the coastal irrigation area centered in southern Wharton County. No recent (near 2000) subsidence measurements are available for either area, although small amounts of subsidence (less than 2 feet) historically have been documented in both areas.

Several factors limit, or detract from, the ability of the model to reliably predict aquifer responses to future conditions. For example, associated with each of the input datasets is a level of uncertainty and a degree of bias, neither of which is quantitatively known. The result is that the optimum (but non-unique) distributions of input data arrived at through calibration, or history matching, are distributions of effective properties, not actual properties. In all likelihood, the property distributions reflect the order of magnitude of the real-system properties, but not the true distributions of the real-system properties. What can be said about the distributions of aquifer-system properties after calibration is that, collectively, they are one set of probably many sets of input data that allows the model to reasonably reproduce selected historical heads, subsidence, and flows. This implies that the reliability of the model for predictive simulation is uncertain.

Introduction

Ground water from the Gulf Coast aquifer system, which includes the Chicot aquifer in rocks of Holocene and Pleistocene age, the Evangeline aquifer in rocks of Pliocene and Miocene age, and the Jasper aquifer in rocks of Miocene age, is an important resource along the northeastern Gulf Coast of Texas. These aquifers supply most of the water used for industrial, municipal, agricultural, and commercial purposes for an approximately 25,000-square-mile (mi^2) area that includes the Beaumont and Houston metropolitan areas. The Houston metropolitan area, which is the 10th largest metropolitan area in the United States (U.S. Census Bureau, 2000), encompasses about 2,500 mi^2 and had an estimated population of 2.95 million in 1995. Water use in the Houston metropolitan area is projected to be about 1.2 billion gallons per day by 2030 (Turner Collie and Braden, Inc., 1996). As the population of Texas increases, appropriate management practices that lead to sustainable use of ground water will be critically important.

Historically, the Texas Gulf coastal area has relied almost entirely on ground water for its water supply. The area has an abundant amount of potable ground water, but withdrawals of large quantities of ground water have resulted in potentiometric-surface declines in the Chicot, Evangeline, and Jasper aquifers, land-surface subsidence from depressurization and compaction of clay layers interbedded in the aquifer sediments, and to a lesser extent, saline-water intrusion. The adverse effects of ground-water withdrawals led to the creation of the Harris-Galveston Coastal Subsidence District (HGCSD) in 1975, the Fort Bend Subsidence District (FBSD) in 1989, and more recently, several other ground-water conservation districts in southeastern Texas.

The primary purpose of the HGCSD and the FBSD is to control land-surface subsidence in Harris, Galveston, and Fort Bend Counties by regulating ground-water withdrawals. For example, the current (2004) HGCSD regulatory plan (Harris-Galveston Coastal Subsidence District, 1999) mandates restrictions on withdrawals in each of three jurisdictional areas. In the coastal and central areas, withdrawals for each permittee must be no more than 10 and 20 percent, respectively, of the permittee's total water demand. In the northwestern area, withdrawals for each permittee by 2010 must be no more than 70 percent of the permittee's total water demand, by 2020 no more than 30 percent, and by 2030 no more than 20 percent. Noncompliance incurs disincentive fees.

The Texas Water Development Board (TWDB) Ground-water Availability Modeling (or Model) (GAM) program was initially funded by the Texas Legislature in 1999. GAM studies by the TWDB, its contractors, and the U.S. Geological Survey (USGS) are developing publicly available computer models of the ground-water-flow systems in the major and minor aquifers of the State. The objective of the program, to be completed in 2004, is to provide reliable, timely data on ground-water availability to the citizens of Texas to ensure adequacy of water supplies or recognition of inadequacy of supplies throughout the 50-year planning period 2000 to 2050 (Texas Water Development Board, 2004). Results from the GAM program are intended to be a tool that water-resource managers can use to address future ground-water-availability issues.

The USGS has worked continually for years to increase the quantity and quality of hydrogeologic information available to water-resource managers. The most recent previous USGS study (Kasmarek and Strom, 2002), done in cooperation with the City of Houston, involved simulation of ground-water flow and land-surface subsidence in the Chicot and Evangeline aquifers in the Houston area. To meet GAM requirements for simulating the northern part of the Gulf Coast aquifer system, the Chicot and Evangeline aquifer flow model of Kasmarek and Strom (2002) was expanded to include the underlying Burkeville confining unit and Jasper aquifer in a study done in cooperation with the HGCSD and the TWDB as a part of the GAM program. Inclusion of the Burkeville confining unit and the Jasper aquifer necessitated extending the modeled area of Kasmarek and Strom (2002) northwestward to encompass the outcrops of these units. The modeled area of Kasmarek and Strom (2002) also was extended northeastward into the Sabine River Basin and southeastward into the Lavaca River Basin to encompass the prescribed GAM study area (GAM area) (Texas Water Development Board, 2004).

Purpose and Scope

The purpose of this report is to describe the hydrogeology and simulation of ground-water flow and land-surface subsidence in the northern part of the Gulf Coast aquifer system in the prescribed GAM area (fig. 1) and to document development and testing of the Gulf Coast (northern part) GAM (as TWDB calls the model). The hydrogeologic units, hydraulic properties, flow conditions, and development (ground-water withdrawals) in the system are summarized on the basis of available information. The hydrogeologic units from land surface downward are the Chicot aquifer, Evangeline aquifer, Burkeville confining unit, Jasper aquifer, and Catahoula confining unit. Little discussion of the Catahoula confining unit is included because only the uppermost four units are actively simulated layers of the GAM. Development and testing of the GAM consisted of designing the finite-difference grid, defining the boundary conditions and stresses, calibration or "history matching," and sensitivity analysis. Calibration involved making a series of transient simulations of historical flow conditions in which input data were iteratively adjusted between simulations on the basis of how closely simulated aquifer responses (potentiometric surfaces [hydraulic heads, or heads], land-surface subsidence, and selected water-budget components) matched measured or estimated responses for selected periods from 1891 through 2000. For this report, predevelopment refers to conditions prior to 1891, and the postdevelopment period is 1891–2000. Ground-water flow was simulated for parts of the hydrogeologic units that contain freshwater.

Previous Studies

Seven previous ground-water-flow-modeling studies, the more recent of which involved land-surface subsidence, have been done in all or parts of the GAM area by the USGS and others. The following information about the first three previous model studies is from Carr and others (1985). The first ground-water-flow model (Wood and Gabrysch, 1965) covered about 5,000 mi^2 in Austin, Brazoria, Chambers, Fort Bend, Galveston, Harris, Liberty, Montgomery, and Waller Counties and was an electric-analog model that used resistors and capacitors to simulate transmissivities and storativities, respectively. The aquifer system was conceptually represented as two layers defined as the "heavily pumped layer" (Evangeline aquifer) and the "Alta Loma Sand" (a permeable zone of the Chicot aquifer). One resistor-capacitor network was used for each layer, and each network was constructed over a base map of the area at a scale of 1 inch (in) equals 1 mile (mi). The model used five stress periods to approximate pumpage from 1890 through 1960 (1890–1930, 1931–40, 1941–47, 1948–53, and 1954–60) and was useful in predicting potentiometric-surface declines caused by various ground-water-withdrawal stresses. Transient simulations yielded reasonable results, but the model was limited by its inability to simultaneously stress both layers and its inability to simulate the effects of ground-water withdrawal in the western part of the modeled area, owing to insufficient historical ground-water-withdrawal data. Model simulations indicated that a more thorough understanding of the aquifer system hydrogeology was needed, and the transmissivity of the aquifers and vertical leakage between the aquifers needed further analysis.

The second ground-water-flow model (Jorgensen, 1975) was an updated electric-analog model that used updated and additional hydrologic data from 1890 to 1970. The two-layer conceptual model divided the aquifer system into the Chicot and Evangeline aquifers, and the electric-analog model made allowances for the vertical movement of water between the two aquifers. The model also accounted for water contributed to the system from storage in clay layers as withdrawals caused the clay layers to be depressurized and compacted, but the model did not simulate land-surface subsidence. The model used six stress periods to approximate pumpage from 1890 through 1970 (1890–1930, 1931–46, 1947–53, 1954–60, 1961–64, and 1965–70) and covered an expanded area of about 9,100 mi^2. Expanding the modeled area enabled the lateral boundaries to be farther

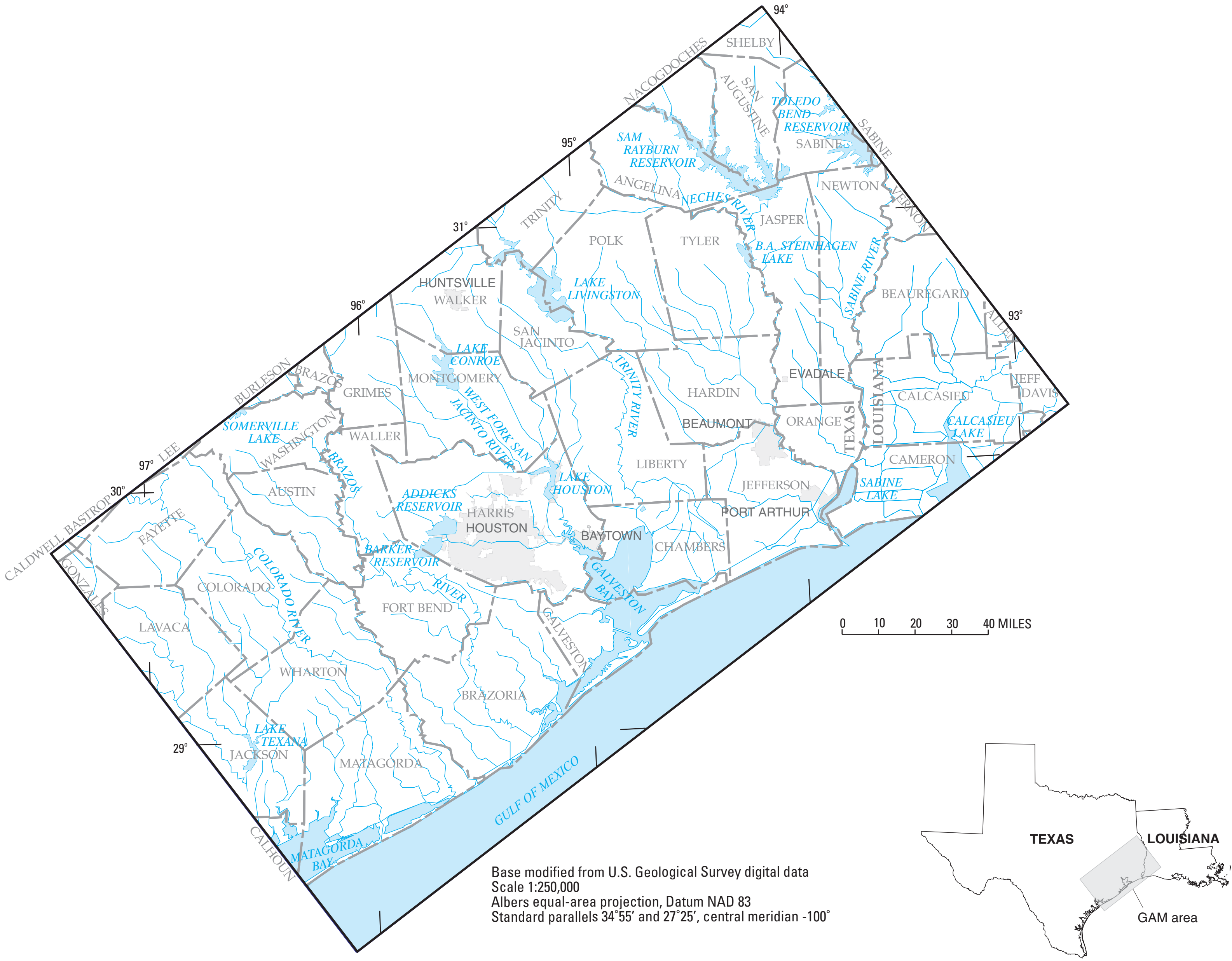


Figure 1. Location of Ground-Water Availability Model (GAM) area.

from areas of large ground-water withdrawals. The modeled area consisted of all of Fort Bend, Harris, and Waller Counties and parts of Brazoria, Chambers, Galveston, Liberty, and Montgomery Counties.

The third ground-water-flow model (Meyer and Carr, 1979) was the first finite-difference numerical model for simulation of three-dimensional ground-water flow. The model was modified from Trescott (1975). The model covered 27,000 mi^2 and consisted of five layers, each with 63 rows and 67 columns. The model grid was variably spaced with the smallest cells representing a 1-mi by 1-mi block; block size increased to nearly 397 mi^2 toward the lateral boundaries of the model. Layer 1 (lowermost) represented the total thickness of the sand beds in the Evangeline aquifer. Layer 2 represented the clay thickness between the centerline of the Chicot aquifer and the centerline of the Evangeline aquifer. Layer 3 represented the Alta Loma Sand where present; otherwise it represented the total sand thickness of the Chicot aquifer. Layer 4 represented the clay thickness between land surface and the centerline of the Chicot aquifer. Layer 5 represented an upper boundary that accounted for recharge from precipitation and return flow from irrigation and other agricultural sources.

Compared to the first and second models, the expanded area of the third model provided more distance from areas of large ground-water withdrawals to the lateral model boundaries. Ground-water withdrawals were compiled for seven historical periods from 1890 through 1975 (1890–1930, 1931–45, 1946–53, 1954–60, 1961–70, 1971–73, and 1974–75). The model was used to predict potentiometric-surface declines under different ground-water-withdrawal scenarios and included methods to increase or decrease the values of clay storage for heads equivalent to preconsolidation stress (Leake and Prudic, 1991), which allowed for the first time simulation of land-surface subsidence. Initial preconsolidated stress approximates the maximum effective stress to which deposits within the area have been subjected before ground-water development; it was estimated from model calibration to be 70 feet (ft) of head. Additionally, this model and the two previously mentioned models were designed to simulate well-field ground-water withdrawals using stress periods of 1 year or longer.

The fourth ground-water-flow model, developed by Espey, Huston and Associates, Inc. (1982) for the HGCSD, was a three-dimensional, finite-difference ground-water model. This model, also known as GWMOD, used the Trescott (1975) computer code subsequently modified by Meyer and Carr (1979). The model encompassed 27,000 mi^2, which included all of Galveston and Harris Counties and parts of Brazoria, Chambers, Fort Bend, Hardin, Jefferson, Liberty, Matagorda, Montgomery, Waller, and Wharton Counties. The vertical discretization of the aquifers was based on the previous modeling studies of the hydrogeology in the area by the USGS (Wood and Gabrysch, 1965; Jorgensen, 1975; and Meyer and Carr, 1979). The model used a uniformly spaced grid of 30 rows and 39 columns with a block size of 7.2 mi^2 and had the ability to simulate the release of water from sand and clay storage as water levels (hydraulic heads, or heads) declined. Model calibration was done using 1960–80 ground-water-withdrawal data collected by several agencies and primarily involved modifying transmissivity and vertical hydraulic conductance between the aquifers. Model calibration was tested by comparing the simulated potentiometric surfaces to measured hydraulic head data compiled and maintained by the USGS. Three main ground-water-withdrawal scenarios were selected to simulate projected water-level declines through 2020.

Modeling of land-surface subsidence was associated with, but not part of, the Espey, Huston ground-water-flow model. Land-surface subsidence was modeled using a modified version of the COMPAC code developed by Helm (1975; 1976a, b; 1978) known as the PRESS (Predictions Relating Effective Stress to Subsidence) model. The PRESS model solves the Terzaghi equations of consolidation on the basis of constant, one-dimensional total stress and transient changes of pore pressures for a given specific site in the aquifers. The Espey, Huston ground-water-flow model simulated water-level declines that were subsequently used as input data for 21 PRESS models, one for each of 21 different geographic locations. Calibration of each PRESS model and land-surface-subsidence simulation were done for the same time periods and water-level-decline data as those of the ground-water-flow model.

The fifth ground-water-flow model (Carr and others, 1985) actually was four separate modified Trescott (1975) finite-difference models that areally overlapped one another in places. The four models encompassed four subregions: Eastern, Houston, Central, and Southern. These subregions extended from Louisiana along the Texas Gulf Coast almost to Mexico. The model was conceptually equivalent to the Meyer and Carr (1979) model. The separate models were calibrated in areas having historical water-level data from 1890 through 1975 for the Houston subregion and from 1900 through 1970 for all other subregions. Notable findings of this study were that a large part of the updip section of the Chicot aquifer is under water-table conditions, vertical leakage from land surface to the Chicot aquifer is an important source of water to the aquifer system, and transmissivities derived from model calibration were about 70 to 80 percent of those obtained solely from aquifer tests. Additionally, initial preconsolidation stress as indicated by model calibration was 70 ft as used in the previous model.

The sixth ground-water-flow model, developed by LBG-Guyton Associates (1997), converted the HGCSD GWMOD model (Espey, Huston and Associates, Inc., 1982) code to a format that could be used with the USGS finite-difference model code MODFLOW (Harbaugh and McDonald, 1996). The model contained 5,850 grid blocks—five layers of 30 rows by 39 columns with blocks 2.5 minutes on a side (2.50 by 2.87 mi). The model area encompassed 8,400 mi^2, which included all of Fort Bend, Galveston, and Harris Counties and parts of Brazoria, Chambers, Grimes, Hardin, Liberty, Matagorda, Montgomery, Waller, and Wharton Counties. Transient calibration of the model was based on measured USGS potentiometric surfaces for 1980, 1988, and 1995.

Associated with but not part of the LBG-Guyton Associates (1997) ground-water-flow model was Fugro-McClelland

(Southwest), Inc. (1997), modeling of land-surface subsidence. Similar to Espey, Huston and Associates, Inc. (1982), Fugro-McClelland (Southwest), Inc., used the PRESS code to simulate land-surface subsidence. The simulated water-level declines from the LBG-Guyton Associates (1997) ground-water-flow model were used as input data for PRESS models at 22 separate sites. The land-surface subsidence modeling included recalibrating 20 of the 21 Espey, Huston PRESS models and calibrating two additional PRESS models. Recalibration of the 20 Espey, Huston PRESS models was necessary because the models had not been tested since their original 1982 calibrations, which were based on measured land-surface subsidence through 1978 and potentiometric-surface data through 1980. The 22 PRESS models were used to estimate land-surface subsidence from 1995 to 2030 for a ground-water-withdrawal scenario provided by the HGCSD, which was based on water-level declines for all post-1995 ground-water demand.

The seventh ground-water-flow model (Kasmarek and Strom, 2002), the precursor to the model described in this report, used the MODFLOW model code (McDonald and Harbaugh, 1988; Harbaugh and McDonald, 1996) to simulate ground-water flow in the Chicot and Evangeline aquifers. Coupled with MODFLOW, the Interbed-Storage Package (Leake and Prudic, 1991) was used to simulate clay compaction and storage in both aquifers. The finite-difference grid covered 18,100 mi^2 and encompassed all of Brazoria, Chambers, Fort Bend, Galveston, Harris, Liberty, and Waller Counties and parts of Austin, Colorado, Fayette, Grimes, Hardin, Jefferson, Matagorda, Montgomery, Polk, San Jacinto, Walker, Washington, and Wharton Counties. The focus of the study was Harris and Galveston Counties, but the modeled area was extended southwest into Colorado, Wharton, and Matagorda Counties and northeast into Hardin, Jefferson, and Polk Counties so that areas having concentrated high rates of ground-water withdrawal would have a minimal effect on potentiometric surfaces at the model boundaries. The model grid was oriented parallel to the Texas Gulf Coast to better coincide with the naturally occurring ground-water divides, boundaries, and predevelopment flow paths. The system was assumed to be horizontally isotropic. Each grid layer consisted of 103 rows and 109 columns. The model was vertically discretized into three layers that resulted in a total of 33,681 grid blocks. Layer 1 represented the water table using a specified constant head, layer 2 represented the Chicot aquifer, and layer 3 represented the Evangeline aquifer. The model grid-block areas varied from 0.95 mi^2 in the central part of the model in and around Harris County to 4.54 mi^2 at the distal model boundaries. The transient simulation period was from 1891 to 1996. Simulated potentiometric surfaces for the Chicot and Evangeline aquifers for 1977 and 1996, when compared to measured potentiometric surfaces for the same periods, showed strong similarities. Additionally, simulated land-surface subsidence for two periods, 1891–1995 and 1978–95, when compared with measured land-surface subsidence for about the same periods, also showed strong similarities.

Description of Ground-Water Availability Model Area

The GAM area (fig. 1) includes all or parts of 38 counties in Texas. Within the GAM area are all or parts of six regional water planning groups: Regions G, H, I, K, L, and P (fig. 2); all or parts of 14 ground-water conservation districts (GCD): Bluebonnet GCD, Brazoria County GCD, Brazos Valley GCD, Coastal Bend GCD, Coastal Plains GCD, Fayette County GCD, Gonzales County Underground Water Conservation District, Lavaca County GCD, Lone Star GCD, Lost Pines GCD, Pineywoods GCD, Post Oak Savannah GCD, Southeast Texas GCD, and Texana GCD; and two subsidence districts: Fort Bend Subsidence District and Harris-Galveston Coastal Subsidence District (fig. 3). Parts of four natural subregions are in the GAM area: Blackland Prairie, Gulf Coast Prairies and Marshes, Oak Woods and Prairies, and Piney Woods (fig. 4); and all or parts of 14 river basins: Brazos, Brazos-Colorado, Colorado, Colorado-Lavaca, Guadalupe, Lavaca, Lavaca-Guadalupe, Neches, Neches-Trinity, Sabine, San Jacinto, San Jacinto-Brazos, Trinity, and Trinity-San Jacinto (fig. 4).

The GAM area is a gently sloping coastal plain, and land-surface altitudes are topographically highest along the northwestern boundary. The vegetation in the northern parts of the GAM area generally is composed of hardwood and pine forests, but as land-surface altitude decreases toward the coast, the vegetation becomes increasingly dominated by shrubs and grasses.

The major river basins in the GAM area are the Brazos, Colorado, Lavaca, Sabine, San Jacinto, and Trinity (fig. 4). Numerous constructed lakes and reservoirs are in the GAM area, but those water bodies generally only influence the water table on a local scale. The Gulf of Mexico and Galveston Bay have a large effect on the downdip ground-water-flow system and climate of the area.

Winters in the GAM area generally are short and mild with few days of freezing temperatures. During winter, moisture-laden Pacific and Canadian air masses produce regionally extensive bands of moderate rainfall. In contrast, summers generally are long and hot. The relative humidity is high, and prevailing winds are from the southwest. During summer, atmospheric convective cells can produce low to high rates of localized rainfall, and infrequently, moisture-laden tropical air masses produce moderate to extremely high rates of rainfall. The average annual rainfall over the GAM area is about 48 in, and the average annual temperature is about 68 degrees Fahrenheit (Larken and Bomar, 1983).

Acknowledgments

The authors acknowledge the HGCSD, the TWDB, and the numerous water-well owners in the GAM area for their help in supplying data. Additionally, the authors thank Robert K. Gabrysch (USGS [retired] and consultant to HGCSD) for his advice and insight on land-surface subsidence and the hydrogeology of the Houston area; Jeffery W. East, Natalie A. Houston, Jenny Lanning-Rush, and Barclay W. Shoemaker (USGS) for

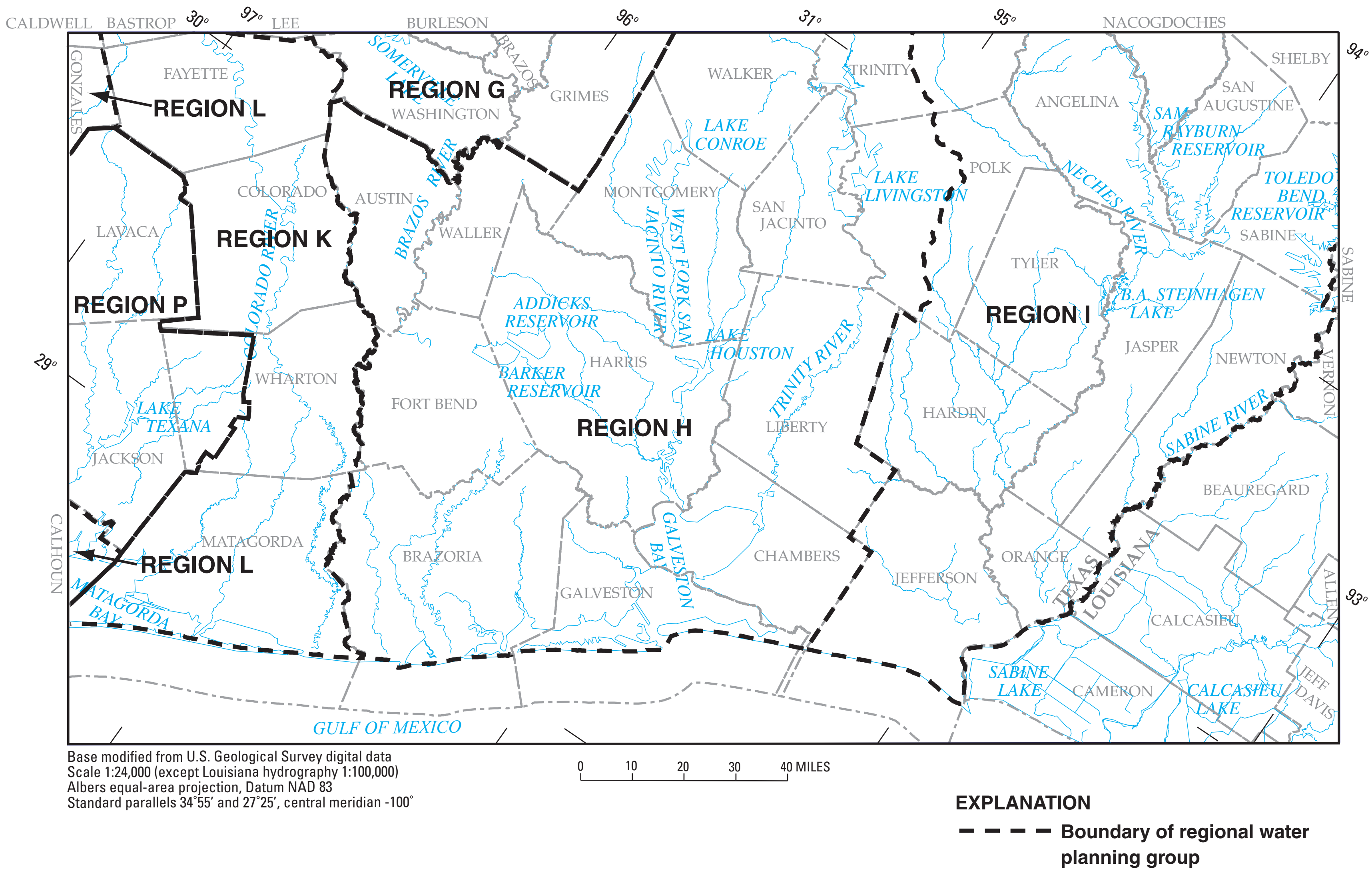


Figure 2. Location of regional water planning groups in the Ground-Water Availability Model area (Texas Water Development Board, 2003).

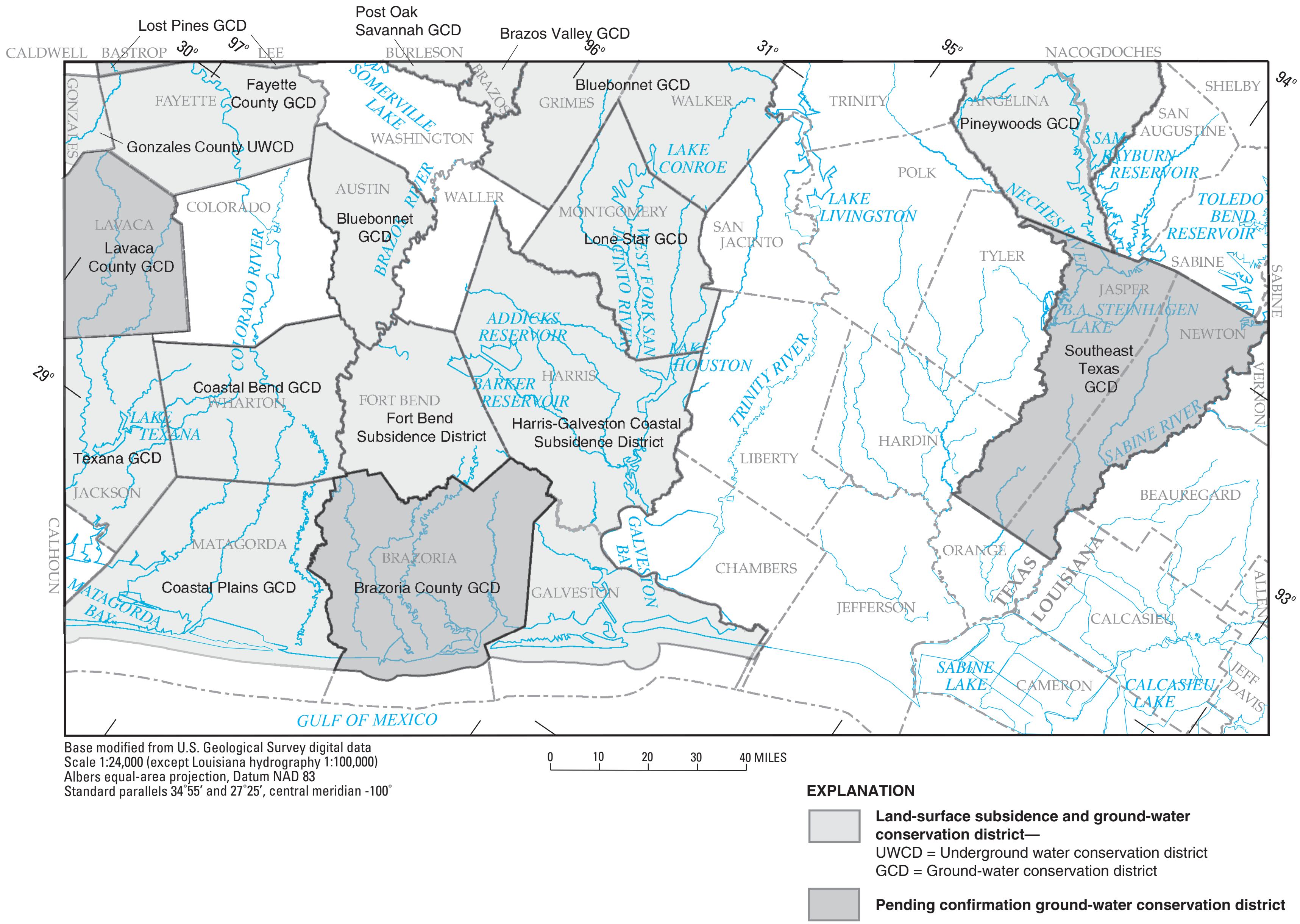


Figure 3. Location of land-surface subsidence and ground-water conservation districts in the Ground-Water Availability Model area (Texas Water Development Board, 2003).

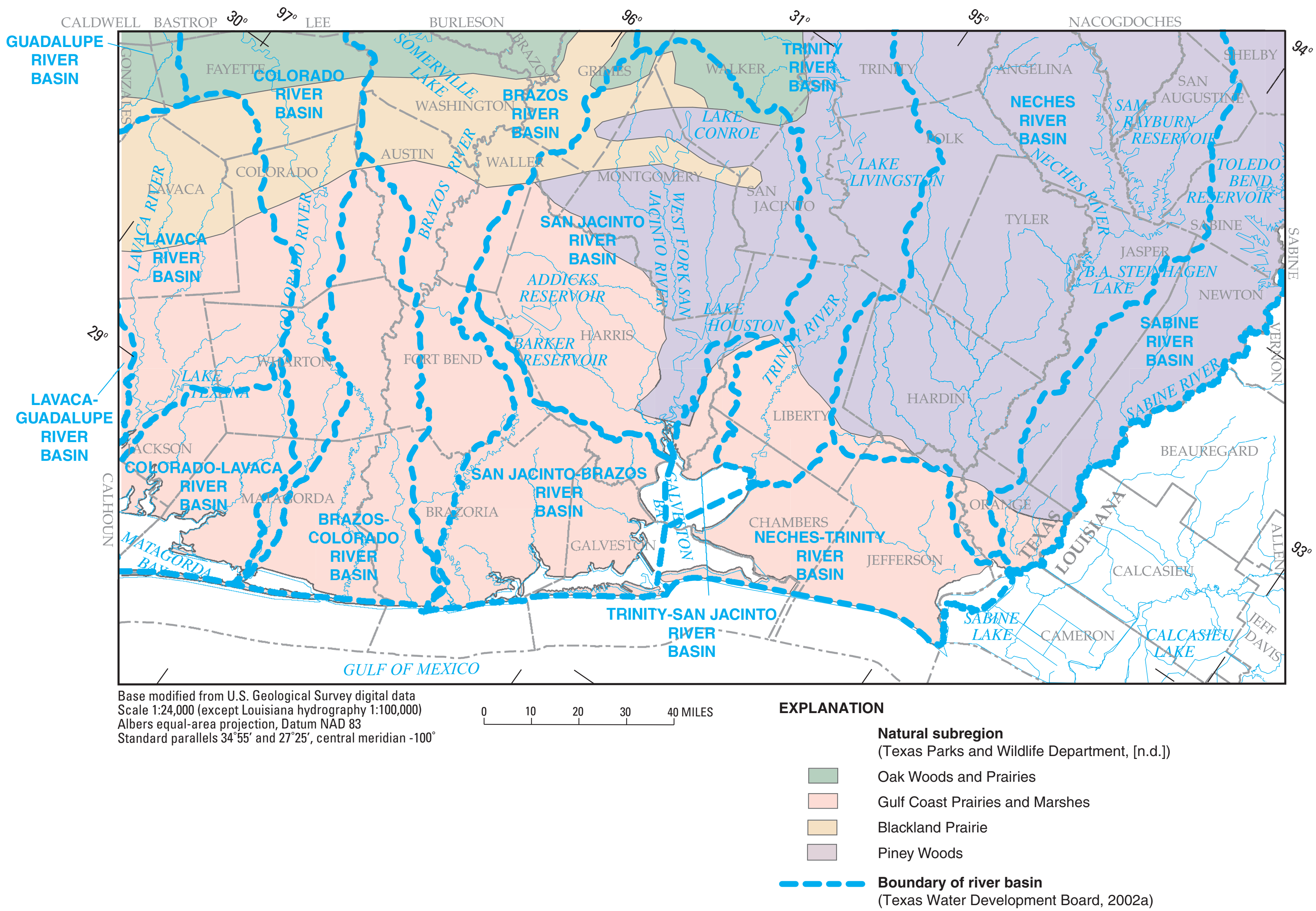


Figure 4. Location of natural subregions and river basins in the Ground-Water Availability Model area.

assistance with computer programming, modeling, and geographic information system (GIS) applications.

Hydrogeology of the Northern Part of the Gulf Coast Aquifer System

In a generalized conceptual model of the aquifer system, the fraction of precipitation that does not evaporate, transpire through plants, or run off the land surface to streams enters the ground-water-flow system in topographically high outcrops of the hydrogeologic units in the northwestern part of the system. Much of the water that infiltrates to the saturated zone flows relatively short distances through shallow zones and discharges to streams; the remainder of the water flows to intermediate and deep zones of the system southeastward of the outcrop areas where it is discharged by wells (in the developed system) and by upward leakage in topographically low areas near the coast (in both predevelopment and postdevelopment but much less in postdevelopment). Near the coast and at depth, saline water is present. The saline water causes less-dense freshwater that has not been captured and discharged by wells to be redirected upward as diffuse leakage to shallow zones of the aquifer system and ultimately to be discharged to coastal water bodies. Only parts of hydrogeologic units containing freshwater are described, because ground-water flow is simulated to the downdip limit of freshwater by the model described in this report.

Hydrogeologic Units and Geologic Setting

From land surface downward, the Chicot aquifer, the Evangeline aquifer, the Burkeville confining unit, the Jasper aquifer, and the Catahoula confining unit are the hydrogeologic units of the Gulf Coast aquifer system, as described by Baker (1979, 1986), and by Ashworth and Hopkins (1995). In general, where the hydrogeologic units crop out, they do so parallel to the coast and thicken downdip to the southeast with the older units having a greater dip angle (fig. 5). The surficial geology (stratigraphic units) in the part of the GAM area coincident with the aquifer system is shown in figure 6. The correlation of hydrogeologic units with stratigraphic units is shown in figure 7. The Chicot aquifer comprises (youngest to oldest) the alluvium, Beaumont Clay, Montgomery Formation, Bentley Formation, and Willis Sand. The Evangeline aquifer comprises (youngest to oldest) the Goliad Sand and the upper part of the Fleming Formation. The Burkeville confining unit consists entirely of the Fleming Formation. The Jasper aquifer comprises (youngest to oldest) the lower part of the Fleming Formation throughout its subsurface extent and the upper part of the Catahoula Sandstone in its outcrop and updip parts (fig. 7). The basal unit for this report is the Catahoula confining unit, which comprises the Catahoula Sandstone, and downdip, the Anahuac and Frio Formations also.

In the GAM area, the updip limit of the Chicot aquifer is an undulating boundary approximately parallel to the coast and extending as far north as Lavaca, Colorado, Austin, Waller, Grimes, Montgomery, San Jacinto, Polk, Tyler, Jasper, and Newton Counties (fig. 8). To the southeast, the freshwater part of the aquifer extends beneath the Gulf of Mexico. The altitude of the top of the Chicot aquifer in the GAM area approximates the land-surface altitude and ranges from sea level (NGVD 29) at the coast to as much as 445 ft above NGVD 29 at its updip limit (fig. 9). The altitude of the base of the Chicot aquifer in the GAM area (fig. 10) ranges from more than 1,500 ft below NGVD 29 southeast of the coast to more than 420 ft above NGVD 29 in the outcrop area and varies locally because of numerous salt domes. The altitude of the base of the Chicot aquifer was constructed from digital data of Strom and others (2003a). The original sources of base altitude data for Strom and others (2003a) were Baker (1979, fig. 2), Carr and others (1985, figs. 4, 5), and Kasmarek and Strom (2002, fig. 5), which included data from Jorgensen (1975, fig. 4). The thickness of the Chicot aquifer (fig. 11) also is from Strom and others (2003a). Cumulative clay thickness of the Chicot aquifer (fig. 12) was subtracted from aquifer thickness to construct cumulative sand thickness (fig. 13).

In the GAM area, the updip limit of the Evangeline aquifer is an undulating boundary approximately parallel to the coast and extending as far north as Lavaca, Fayette, Austin, Washington, Grimes, Montgomery, Walker, San Jacinto, Polk, Tyler, Jasper, and Newton Counties (fig. 14). The downdip extent of freshwater is approximately coincident with the coast. The altitude of the top of the Evangeline aquifer in the GAM area ranges from more than 1,440 ft below NGVD 29 to as much as 469 ft above NGVD 29 at its updip limit (fig. 15). The altitude of the base of the Evangeline aquifer in the GAM area (fig. 16) ranges from more than 5,300 ft below NGVD 29 at the coast near Galveston Bay to 430 ft above NGVD 29 in the outcrop area and varies locally because of numerous salt domes. The base of the Evangeline aquifer transgresses the stratigraphic boundary between the Goliad Sand and the Fleming Formation. (This transgression is not shown in the section of figure 5, as only outcropping stratigraphic units are shown.) The altitude of the base of the Evangeline aquifer, like the base of the Chicot aquifer, was constructed using digital data from Strom and others (2003b). The original sources of the base altitude data for Strom and others (2003b) were Baker (1979, figs. 6, 7; 1986, fig. 7), Carr and others (1985, figs. 6, 7), and Kasmarek and Strom (2002, fig. 7), which included data from Jorgensen (1975, fig. 7). The thickness of the Evangeline aquifer (fig. 17) also is from Strom and others (2003b). Cumulative clay thicknesses of the Evangeline aquifer (fig. 18) (Gabrysch, 1982, fig. 37) was subtracted from aquifer thickness to construct cumulative sand thickness (fig. 19).

In the GAM area, the updip limit of the Burkeville confining unit is an undulating boundary approximately parallel to the coast and extending as far north as Lavaca, Fayette, Austin, Washington, Grimes, Montgomery, Walker, San Jacinto, Polk, Tyler, Jasper, and Newton Counties (fig. 20). The Burkeville

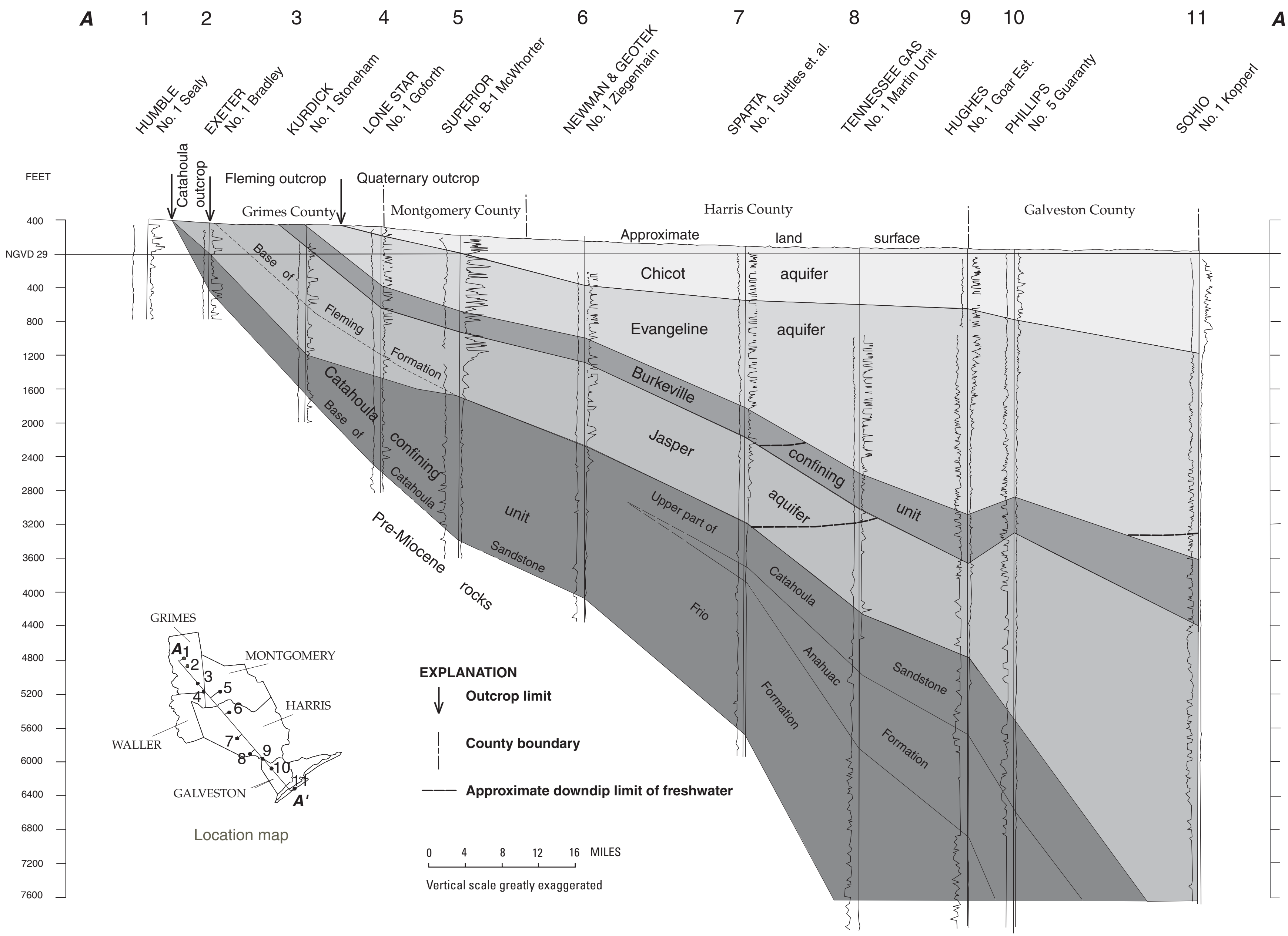
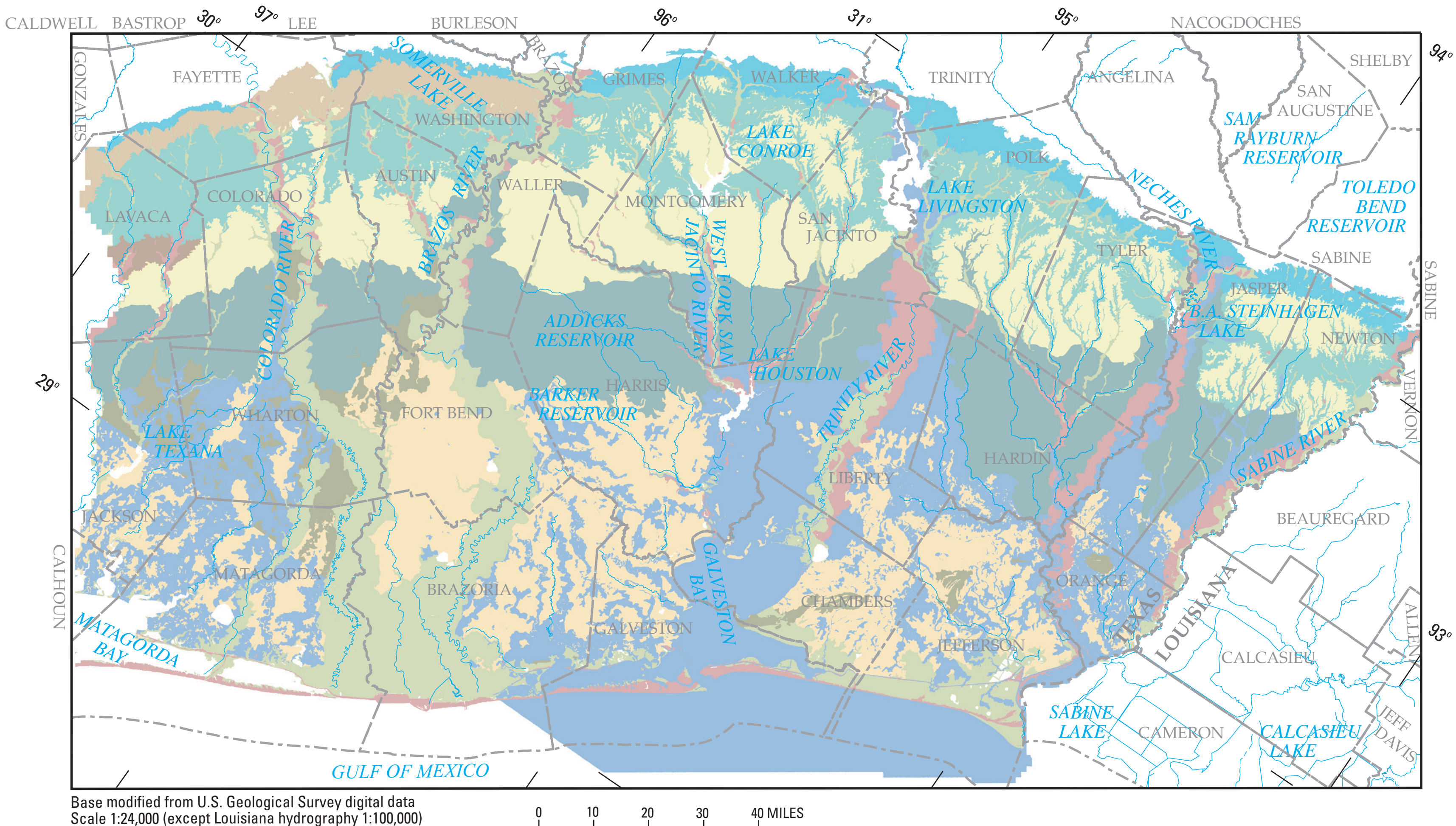


Figure 5. Section showing stratigraphic units that crop out and hydrogeologic units in the Grimes-Harris-Galveston Counties area of the Ground-Water Availability Model area (modified from Baker, 1979, fig. 4; 1986, fig. 5).



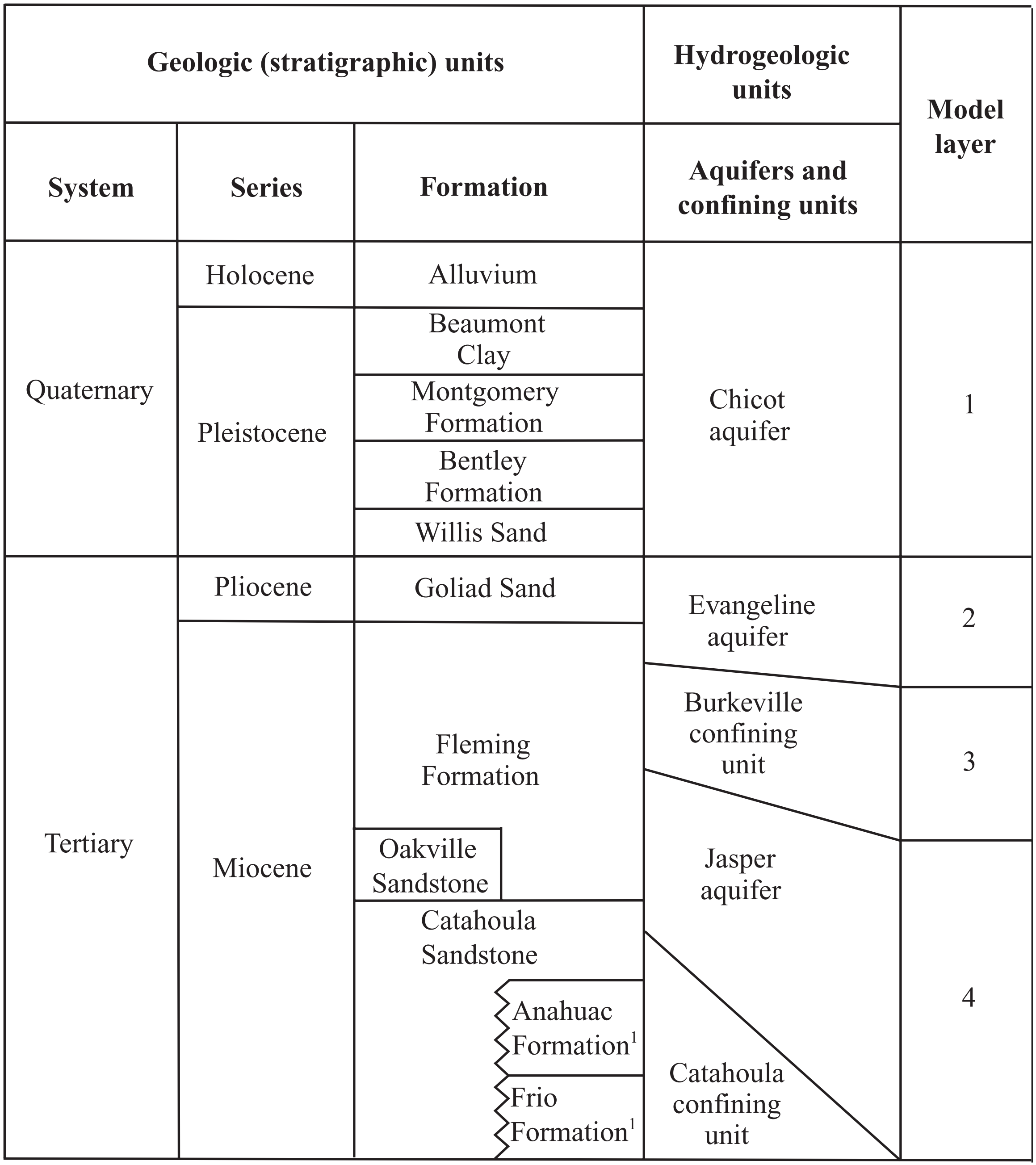
Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

Stratigraphic units

- Quaternary, undifferentiated
- Alluvium
- Beaumont Clay, undifferentiated
  - Sand
  - Clay
- Montgomery and Bentley Formations, undifferentiated
- Willis Sand
- Goliad Sand
- Fleming Formation
- Oakville Sandstone
- Catahoula Sandstone

Figure 6. Surficial geology in the part of the Ground-Water Availability Model area coincident with the aquifer system (modified from University of Texas, Bureau of Economic Geology, 2004).



[1]Present only in subsurface.

Figure 7. Correlation of stratigraphic and hydrogeologic units (modified from Baker, 1979, table 1; 1986, table 1).

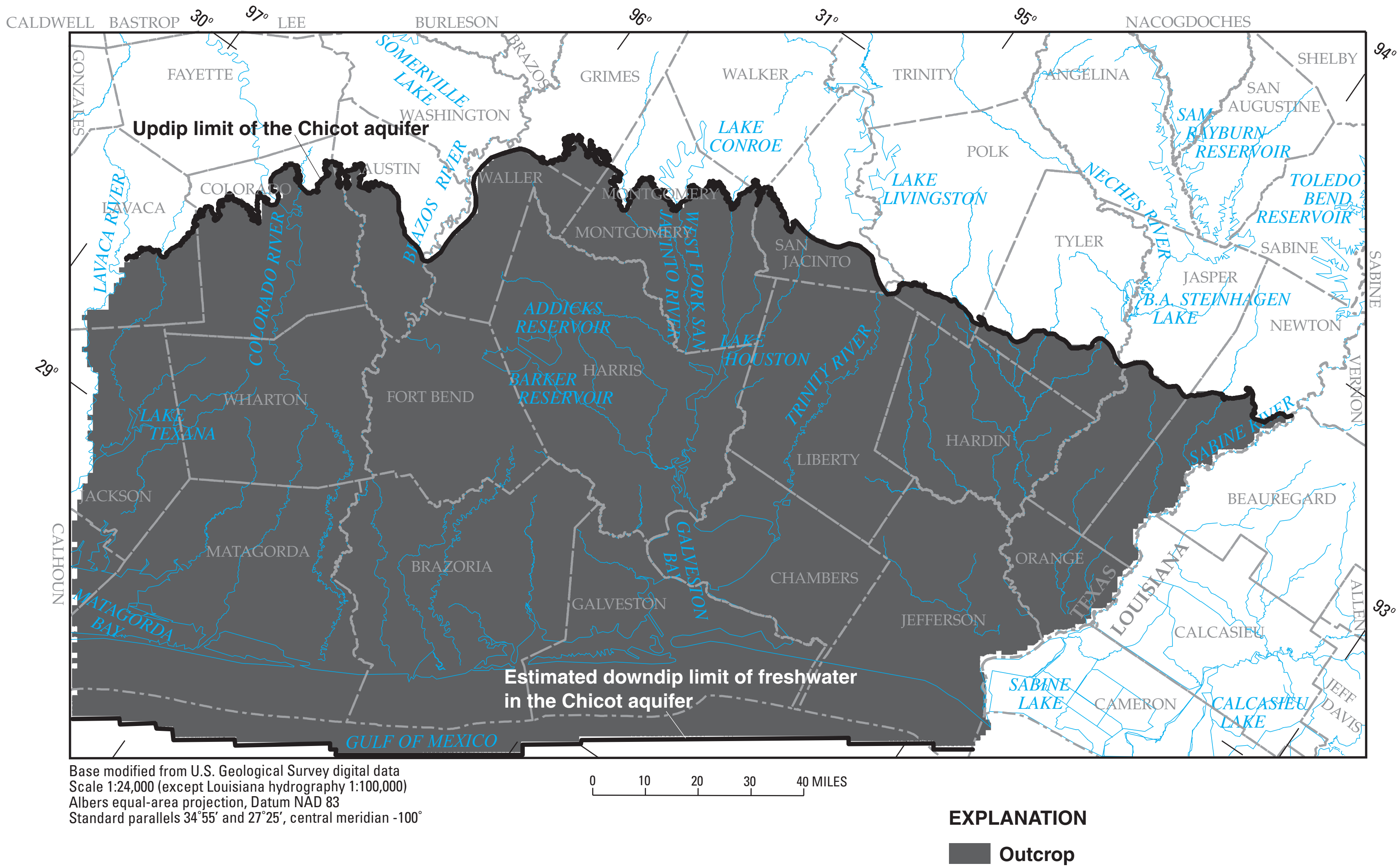
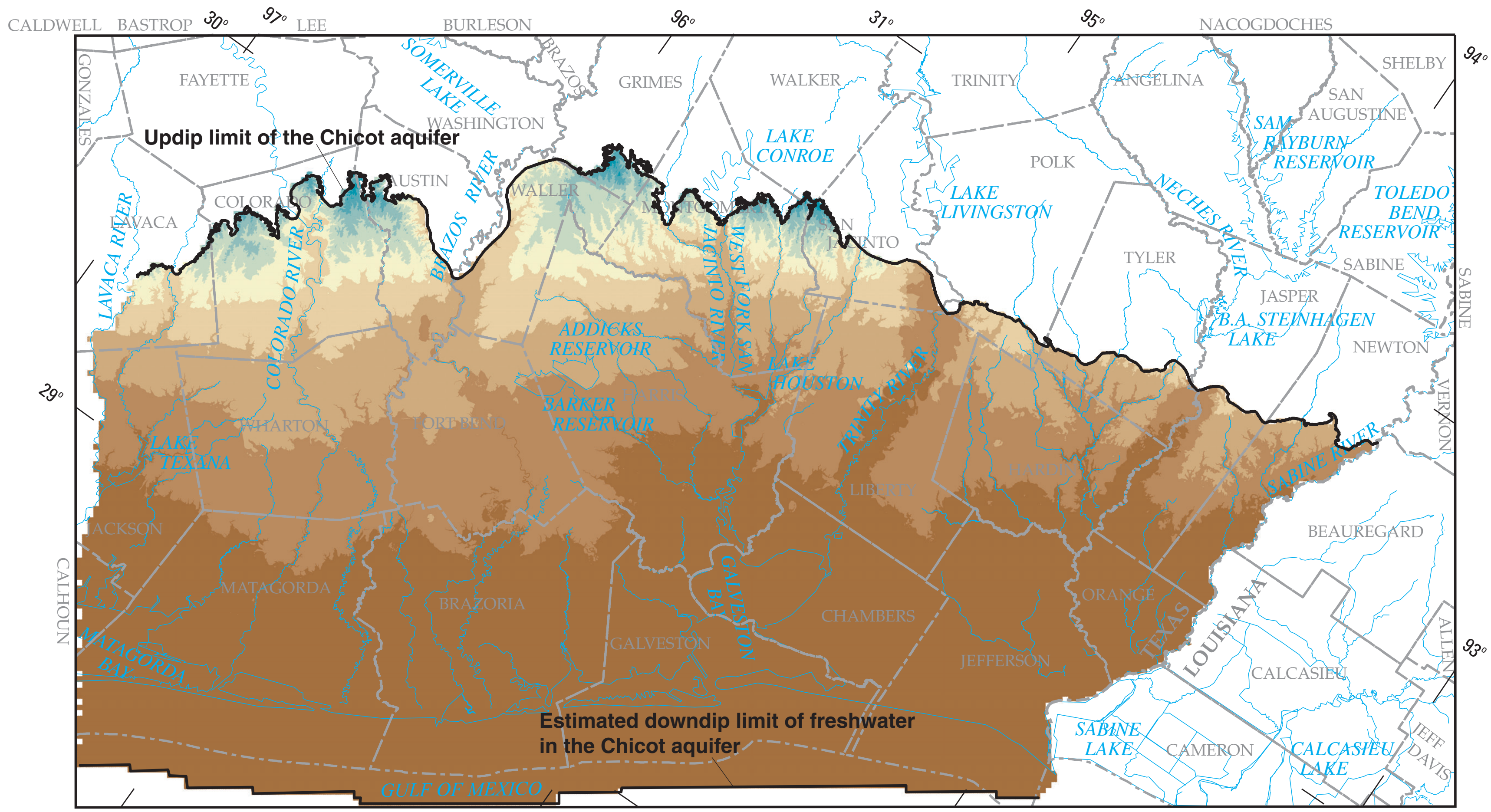


Figure 8. Extent and outcrop area of the Chicot aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003a).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

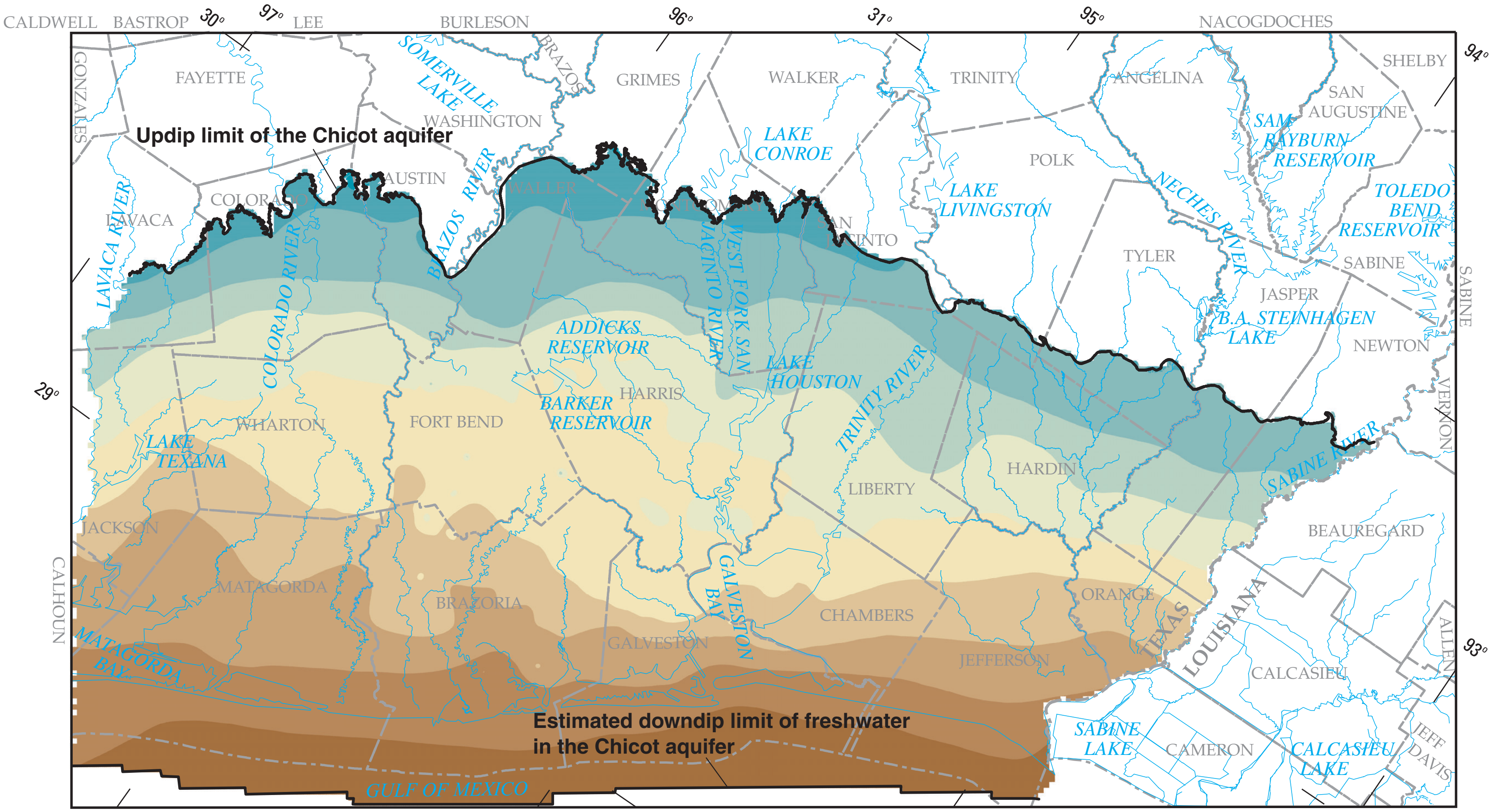
EXPLANATION

Altitude, in feet above NGVD 29

- 0 to 50
- 50 to 100
- 100 to 150
- 150 to 200
- 200 to 250
- 250 to 300
- 300 to 350
- 350 to 400
- 400 to 450

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 9. Altitude of the top of the Chicot aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003a).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

EXPLANATION

Altitude, in feet above NGVD 29

- -1,550 to -1,350
- -1,350 to -1,150
- -1,150 to -950
- -950 to -750
- -750 to -550
- -550 to -350
- -350 to -150
- -150 to 150
- 150 to 350
- 350 to 450

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 10. Altitude of the base of the Chicot aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003a).

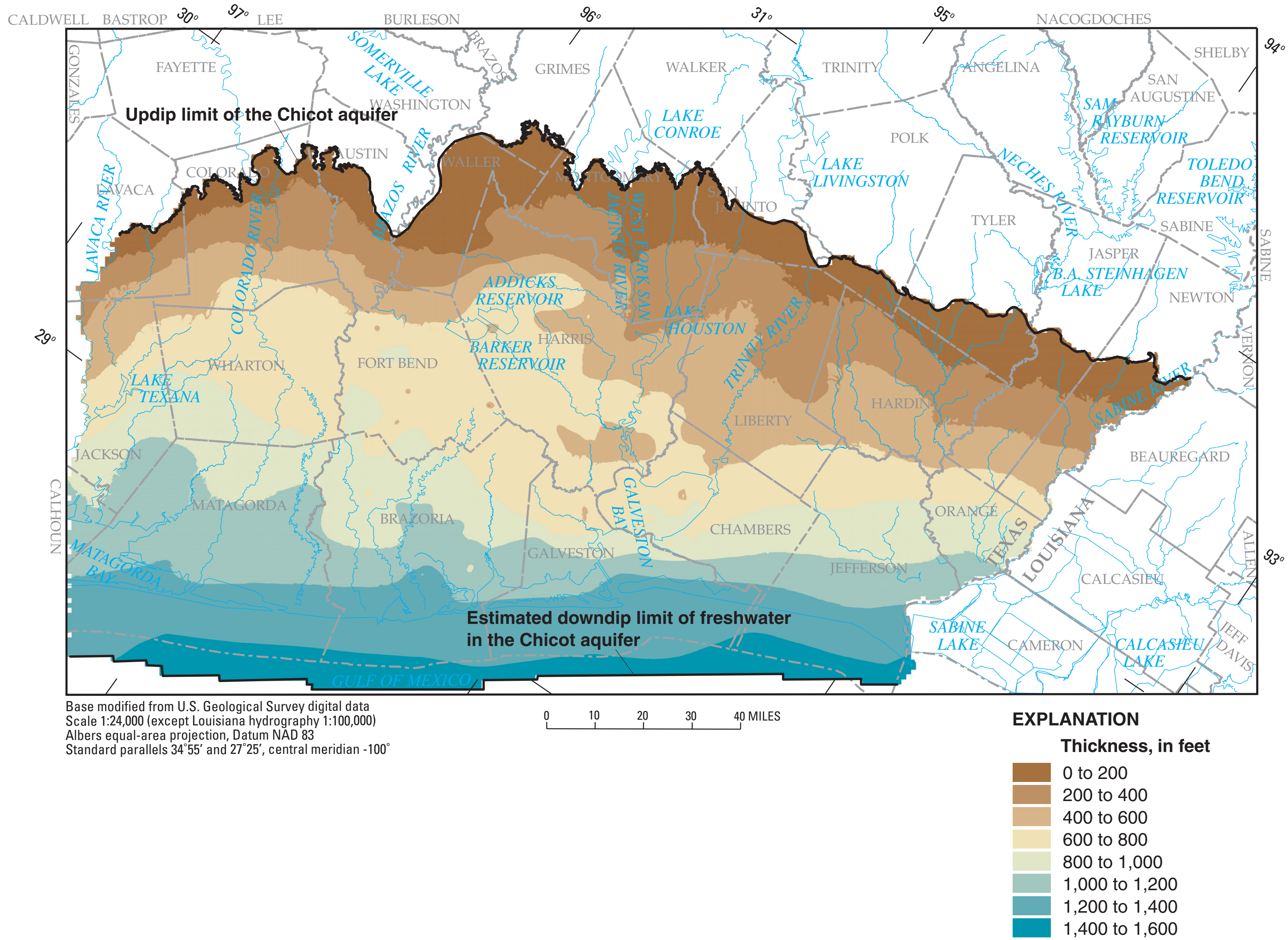
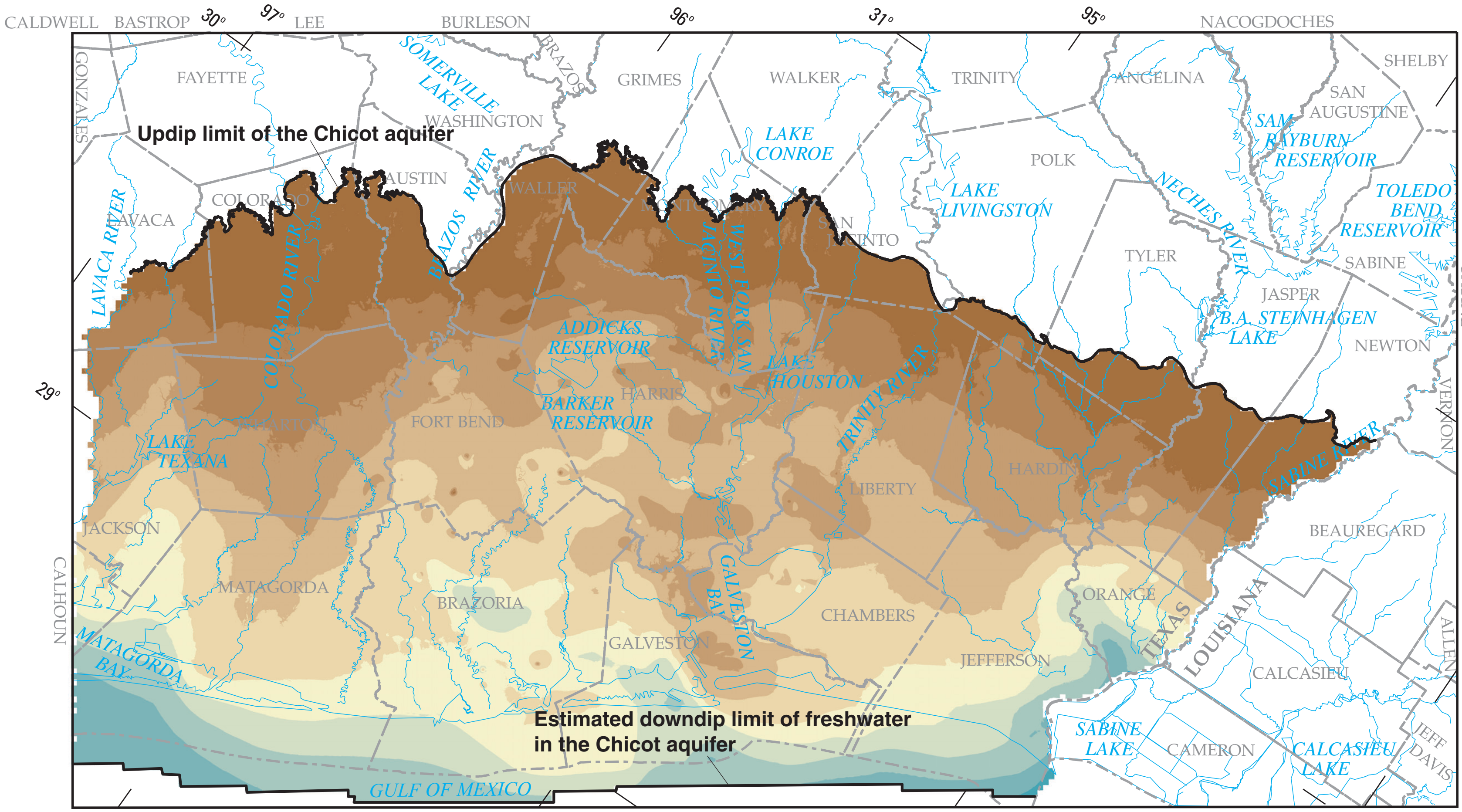


Figure 11. Thickness of the Chicot aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003a).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

EXPLANATION

Thickness, in feet

- 0 to 100
- 100 to 200
- 200 to 300
- 300 to 400
- 400 to 500
- 500 to 600
- 600 to 700
- 700 to 800
- 800 to 900
- 900 to 1,000
- 1,000 to 1,100

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 12. Cumulative clay thickness of the Chicot aquifer in the Ground-Water Availability Model area (modified from Gabrysch, 1982, fig. 37; and Strom and others, 2003a).

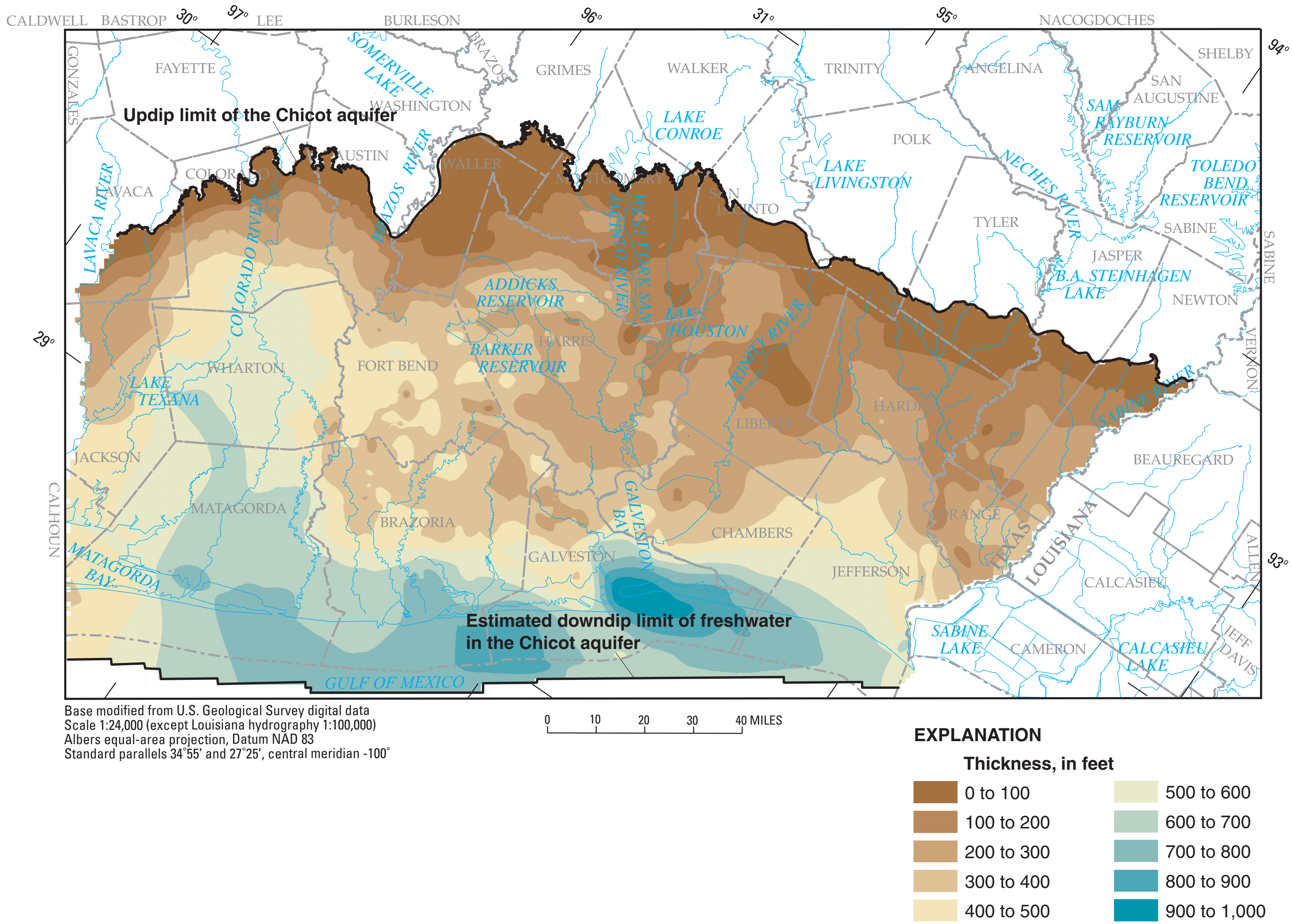


Figure 13. Cumulative sand thickness of the Chicot aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003a).

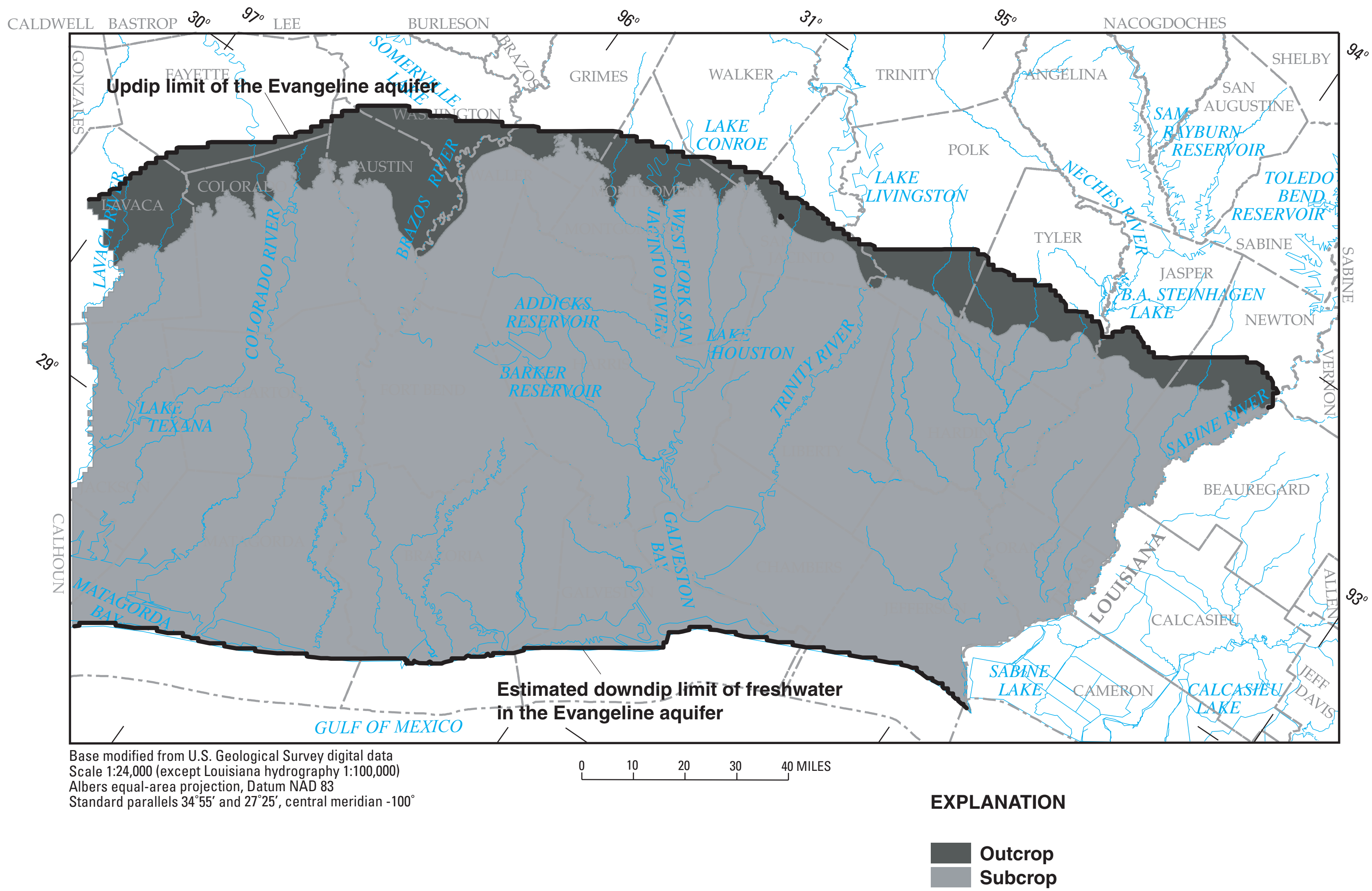
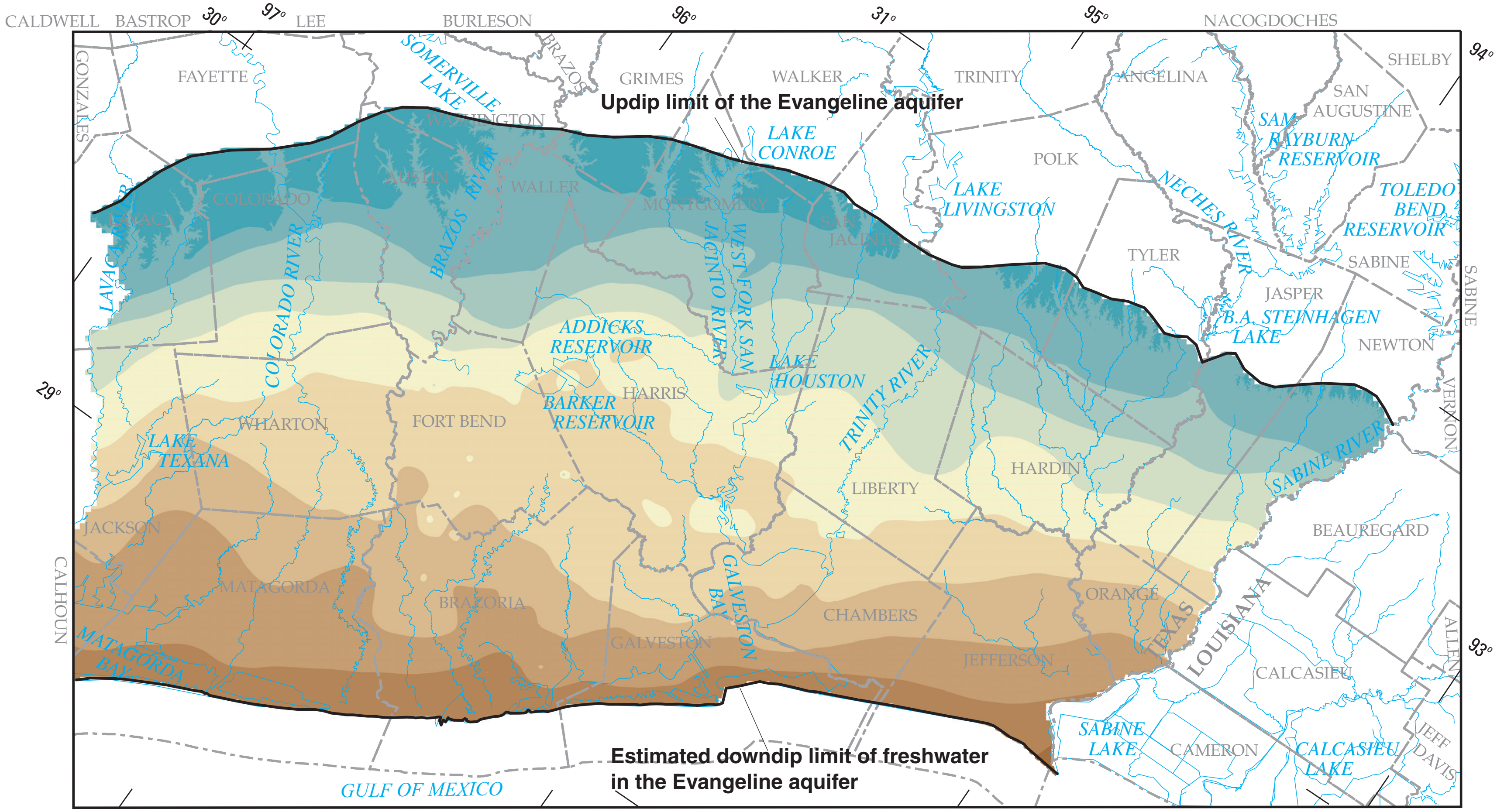


Figure 14. Extent, outcrop area, and subcrop area of the Evangeline aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003b).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

Altitude, in feet above NGVD 29

- -1,550 to -1,350
- -1,350 to -1,150
- -1,150 to -950
- -950 to -750
- -750 to -550
- -550 to -350
- -350 to -150
- -150 to 50
- 50 to 250
- 250 to 450
- 450 to 650

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 15. Altitude of the top of the Evangeline aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003b).

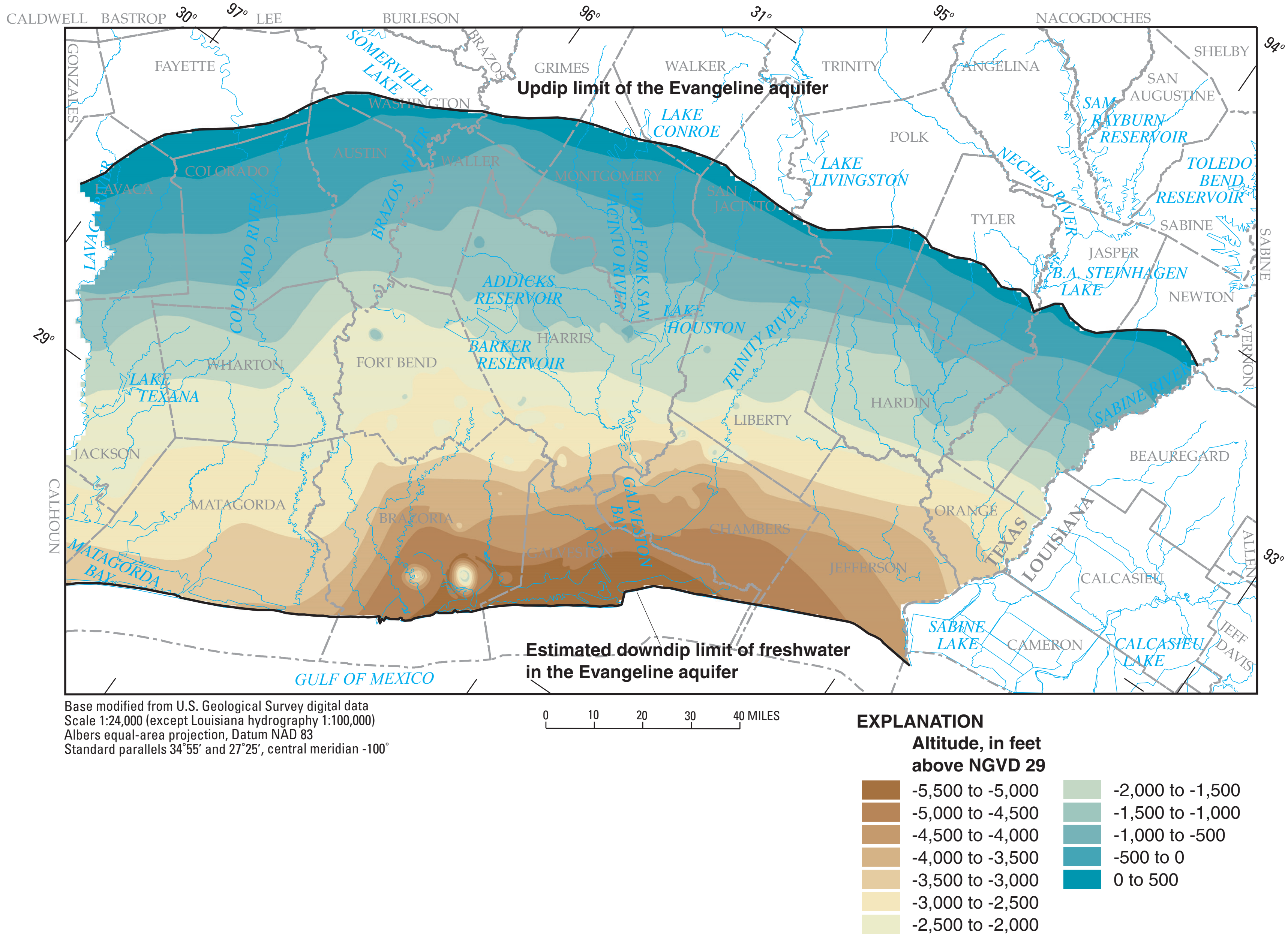


Figure 16. Altitude of the base of the Evangeline aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003b).

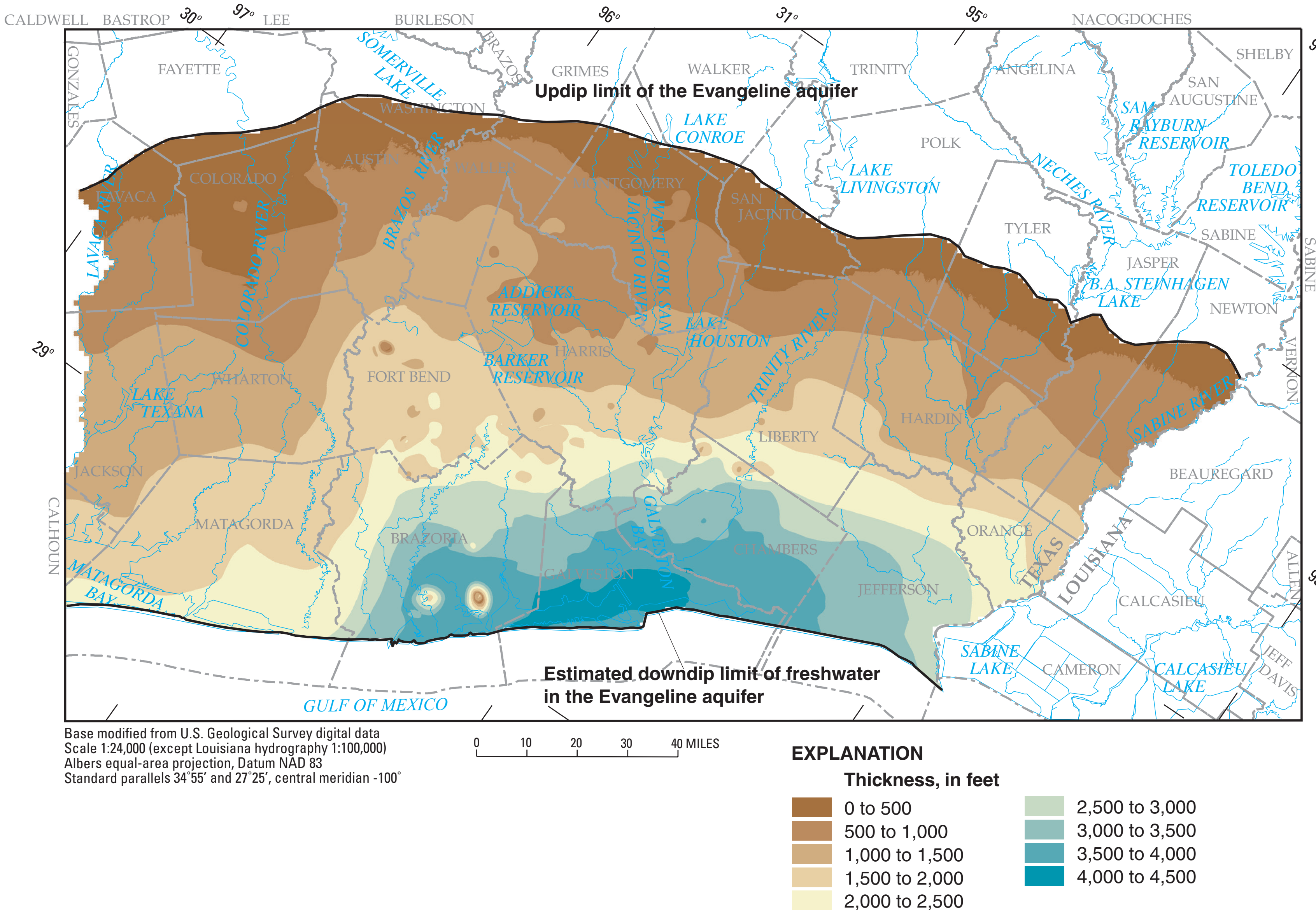
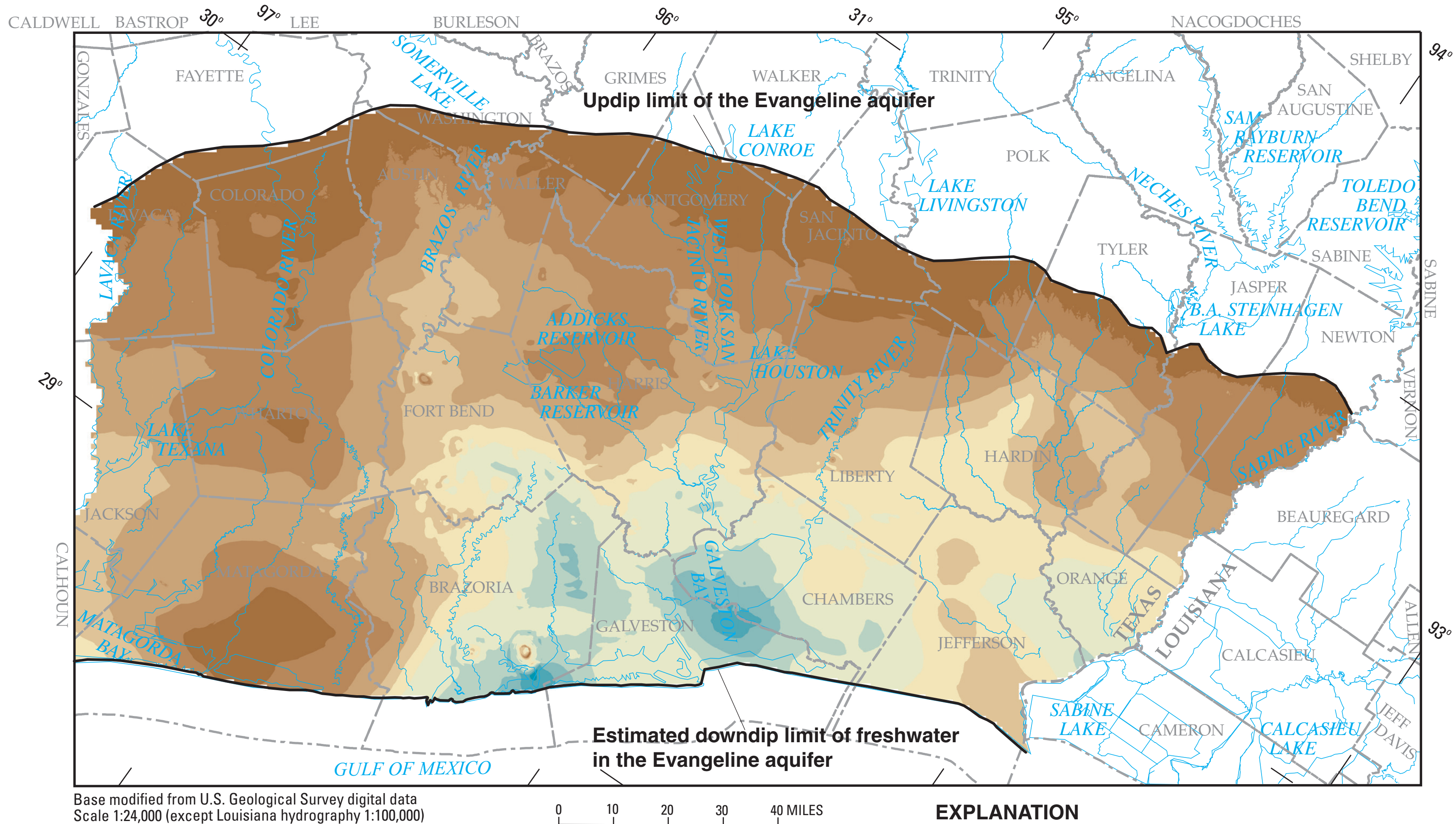


Figure 17. Thickness of the Evangeline aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003b).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

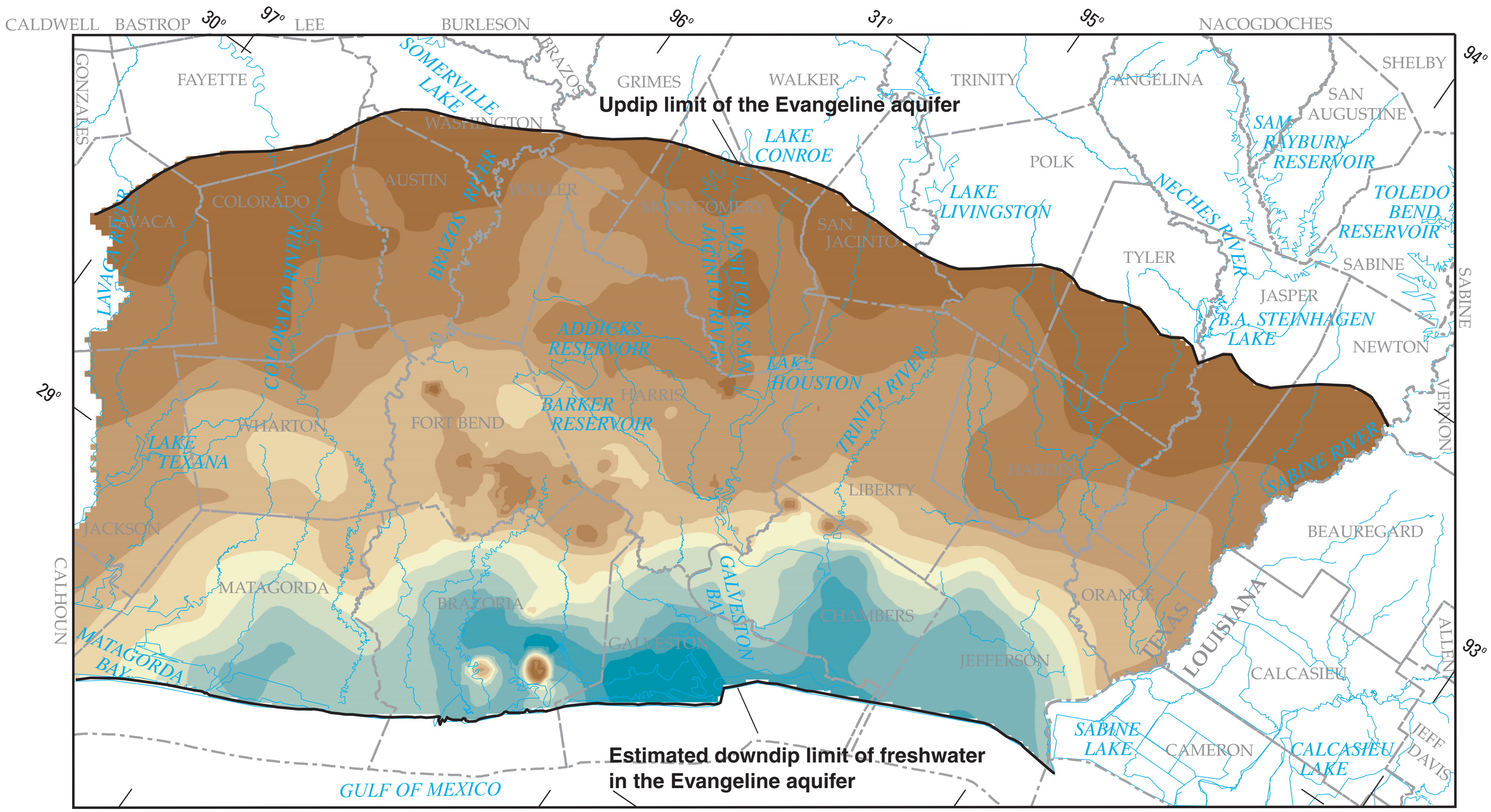
EXPLANATION

Thickness, in feet

- 0 to 250
- 250 to 500
- 500 to 750
- 750 to 1,000
- 1,000 to 1,250
- 1,250 to 1,500
- 1,500 to 1,750
- 1,750 to 2,000
- 2,000 to 2,250
- 2,250 to 2,500

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 18. Cumulative clay thickness of the Evangeline aquifer in the Ground-Water Availability Model area (modified from Gabrysch, 1982, fig. 37; and Strom and others, 2003b).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

EXPLANATION

Thickness, in feet

0 to 250
250 to 500
500 to 750
750 to 1,000
1,000 to 1,250
1,250 to 1,500
1,500 to 1,750
1,750 to 2,000
2,000 to 2,250
2,250 to 2,500
2,500 to 2,750

Figure 19. Cumulative sand thickness of the Evangeline aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003b).

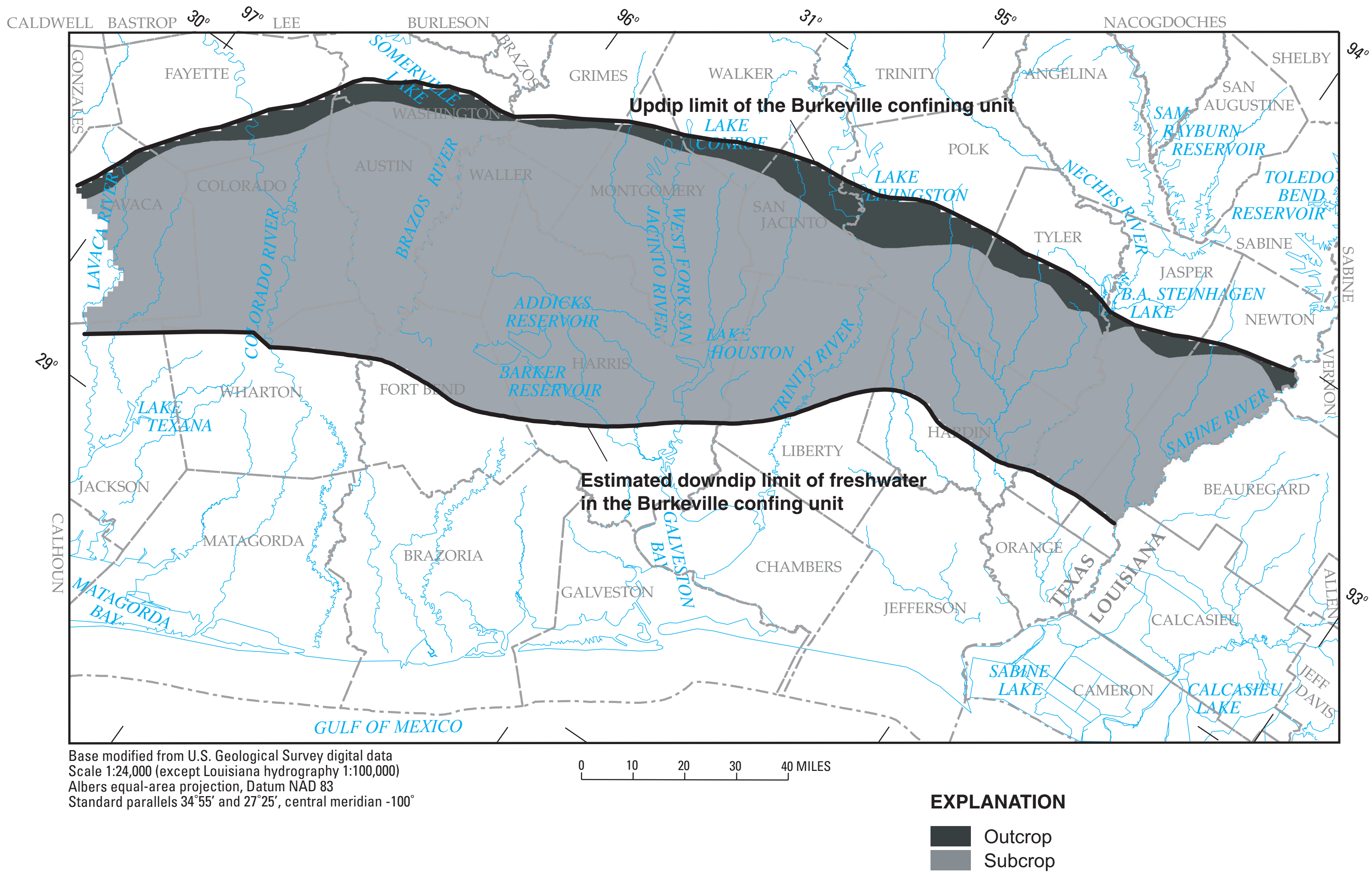


Figure 20. Extent, outcrop area, and subcrop area of the Burkeville confining unit in the Ground-Water Availability Model area (modified from Strom and others, 2003b, c).

confining unit lies stratigraphically below the Evangeline aquifer and above the Jasper aquifer (fig. 5). The Burkeville functions as a confining unit because of its relatively large percentage of silt and clay compared to the percentages of the adjacent aquifers (Baker, 1979). Southeast of the downdip freshwater boundary (fig. 20), this unit is considered (for simulation) a no-flow basal unit that prevents diffuse upward leakage of saline water. In updip areas of the Burkeville confining unit, the sediments are slightly more transmissive and are able to supply small quantities of water for domestic use. In the outcrop area, the altitude of the top of the Burkeville confining unit is equal to the land-surface altitude, and in the subcrop area the top of the Burkeville confining unit is coincident with the base of the Evangeline aquifer. The altitude of the base of the Burkeville confining unit is coincident with the top of the Jasper aquifer and varies locally because of numerous salt domes.

In the GAM area, the updip limit of the Jasper aquifer is an undulating boundary approximately parallel to the coast and extending as far north as Gonzales, Lavaca, Fayette, Washington, Brazos, Grimes, Walker, Trinity, Polk, Tyler, Angelina, Jasper, Newton, and Sabine Counties (fig. 21). Southeast of the downdip boundary of freshwater, this unit is considered (for simulation) a no-flow basal unit that prevents diffuse upward leakage of saline water. The altitude of the top of the Jasper aquifer in the GAM area ranges from less than 2,800 ft below NGVD 29 to about 900 ft above NGVD 29 at its updip limit (fig. 22). The altitude of the base of the freshwater part of the Jasper aquifer (fig. 23) ranges from about 3,800 ft below NGVD 29 near the downdip limit of freshwater to about 500 ft above NGVD 29 in the outcrop area and varies locally because of numerous salt domes. The base of the Jasper aquifer in updip areas transgresses the stratigraphic boundary between the Fleming Formation and the Catahoula Sandstone (figs. 5, 6). The altitudes of the top and base of the Jasper aquifer were created by Strom and others (2003c). The thickness of the Jasper aquifer (fig. 24) also is from Strom and others (2003c). Cumulative clay thickness of the Jasper aquifer (fig. 25) was subtracted from aquifer thickness to construct the cumulative sand thickness (fig. 26).

The Jasper aquifer is underlain by the Catahoula confining unit, which is composed mostly of clay or tuff. The Catahoula prevents any substantial exchange of water between the Jasper aquifer and underlying units (Baker, 1986). Therefore, the Catahoula confining unit is considered the base of the Gulf Coast aquifer system for simulation.

The paleo-depositional environment of the rocks that formed the Gulf Coast aquifer system was a fluvial deltaic or shallow-marine environment that produced interlayered, discontinuous sequences of sand, silt, clay, and gravel. Changes in land-surface altitudes related to naturally occurring land-surface subsidence of the depositional basin and sea-level transgressions and regressions created cyclical sedimentation facies. During periods when the sea level declined, fluvial deltaic processes deposited continental sediments; but as the sea level rose, the deposited continental sediments were reworked and marine sediments were deposited. Because of this complex depositional process, the facies alternate cyclically from the predominantly continental sediments that compose the aquifers to the predominantly marine sediments that compose the confining units and clay layers within aquifers. Therefore, the aquifer system has a high degree of heterogeneity in both lateral and vertical extent (Sellards and others, 1932).

Growth faults are common throughout the unconsolidated sediments of the GAM area, and traces of some of these faults have been mapped and named. On the basis of the study of well logs and seismic-line data, these faults have been delineated to depths of 3,000 to 12,000 ft below land surface (Verbeek and others, 1979). The presence of most of these faults is associated with natural geologic processes. The scale of fault movement is insufficient to completely offset entire hydrogeologic units. However, if an offset results in the juxtaposition of relatively more-permeable sediments against relatively less-permeable sediments, the rate and direction of ground-water flow could be affected. Although growth faults are common in the study area, the frequency with which associated offsets appreciably affect ground-water flow is unknown. Because the distribution and magnitude of such occurrences in the study area are unknown, accounting for them in the GAM was not possible.

Numerous salt domes have been mapped in the GAM area (Beckman and Williamson, 1990) (fig. 27). The salt originated from the Jurassic-age Louann Salt and has risen through the overlying strata (Halbouty, 1967). In some areas, the salt domes have penetrated the aquifers. The upward intrusions of the salt domes decrease the thickness of the adjacent aquifer sediments radially and alter the prevailing hydraulic characteristics and flow paths in the adjacent aquifer sediments. These widely distributed salt domes increase the heterogeneity of the hydraulic characteristics of the aquifers.

Hydraulic Properties

Carr and others (1985) estimated transmissivity and storativity of the Chicot and Evangeline aquifers from simulation for an area essentially the same as that of the GAM. Transmissivity of the Chicot aquifer ranged from about 3,000 to about 50,000 feet squared per day (ft^2/d), and storativity ranged from about 0.0004 to 0.1. Transmissivity of the Evangeline aquifer ranged from about 3,000 to about 15,000 ft^2/d, and storativity ranged from about 5×10^{-4} to 0.1. For both aquifers, the larger storativities are in the updip outcrop areas that are under water-table conditions; the smaller storativities are in downdip areas that are under confined conditions.

Baker (1986) estimated transmissivity of the Jasper aquifer from simulation for an area coincident with most of the Jasper aquifer in the GAM area. The transmissivity of the Jasper aquifer ranged from less than 2,500 to about 35,000 ft^2/d.

Wesselman (1967) estimated transmissivity for all three aquifers and storativity for the Chicot and Evangeline aquifers from aquifer tests in Jasper, Newton, Orange, and Hardin Counties. Transmissivities of the Chicot aquifer ranged from 12,300 to 68,000 ft^2/d, the Evangeline aquifer, 2,130 to 14,800 ft^2/d,

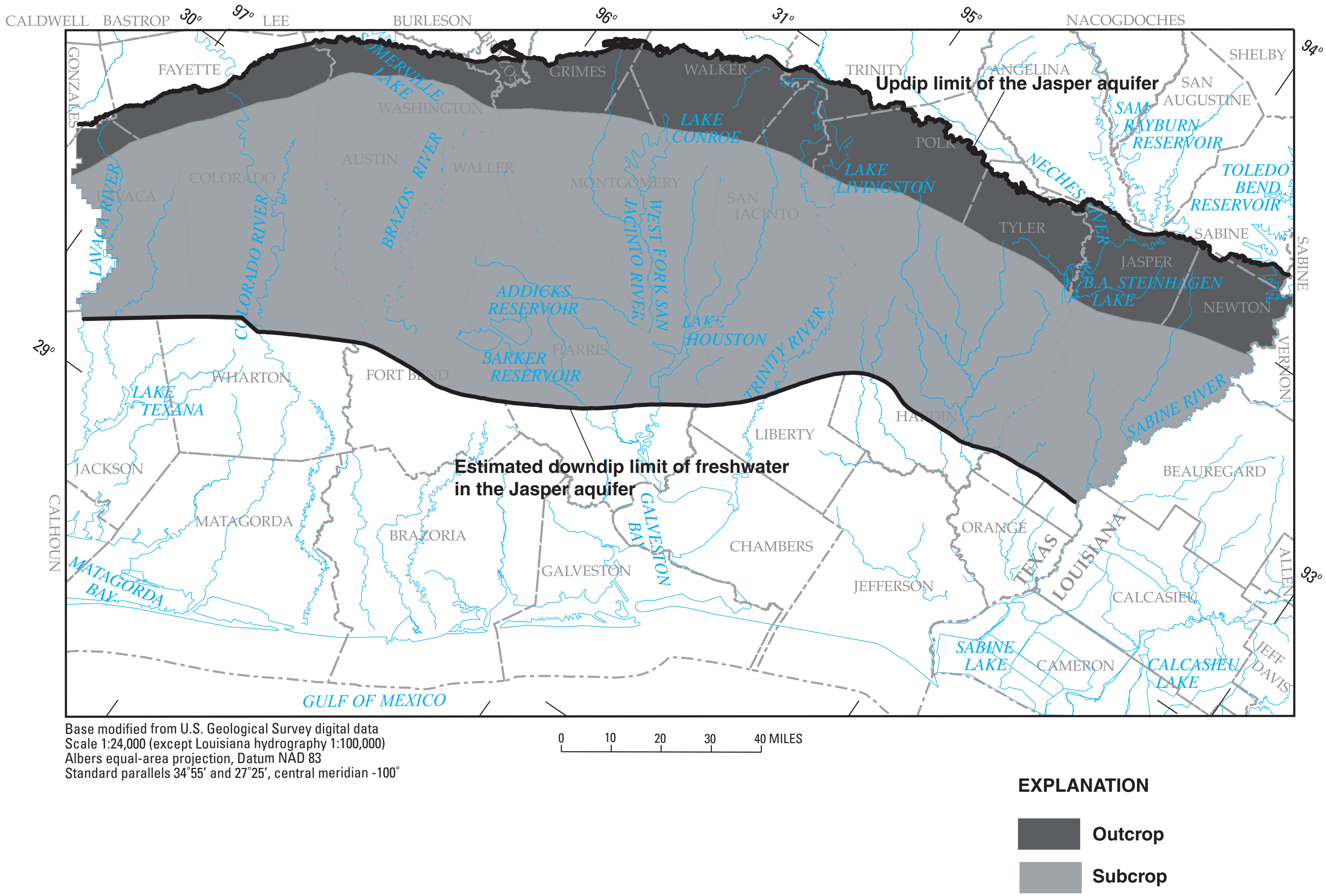


Figure 21. Extent, outcrop area, and subcrop area of the Jasper aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003c).

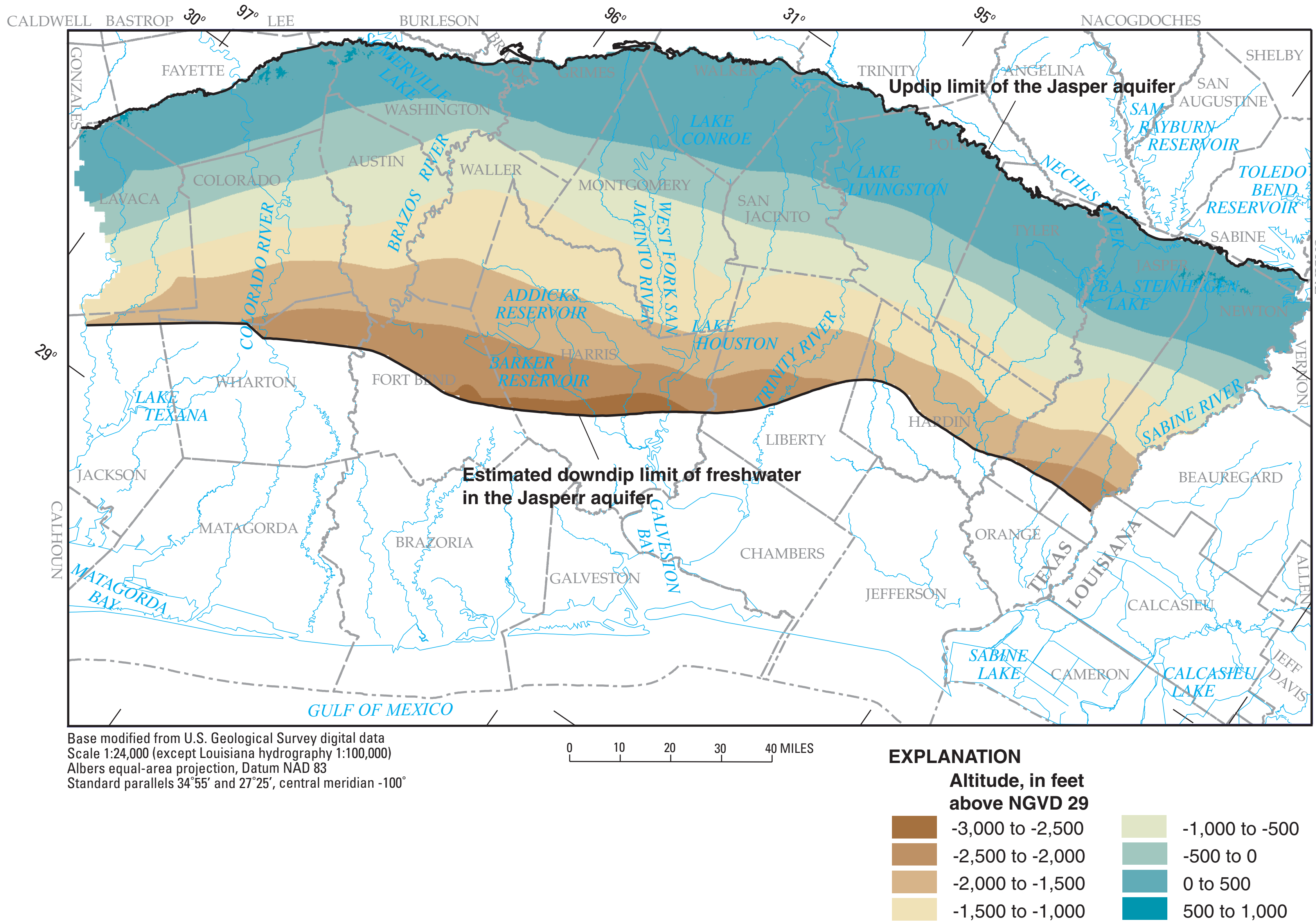


Figure 22. Altitude of the top of the Jasper aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003c).

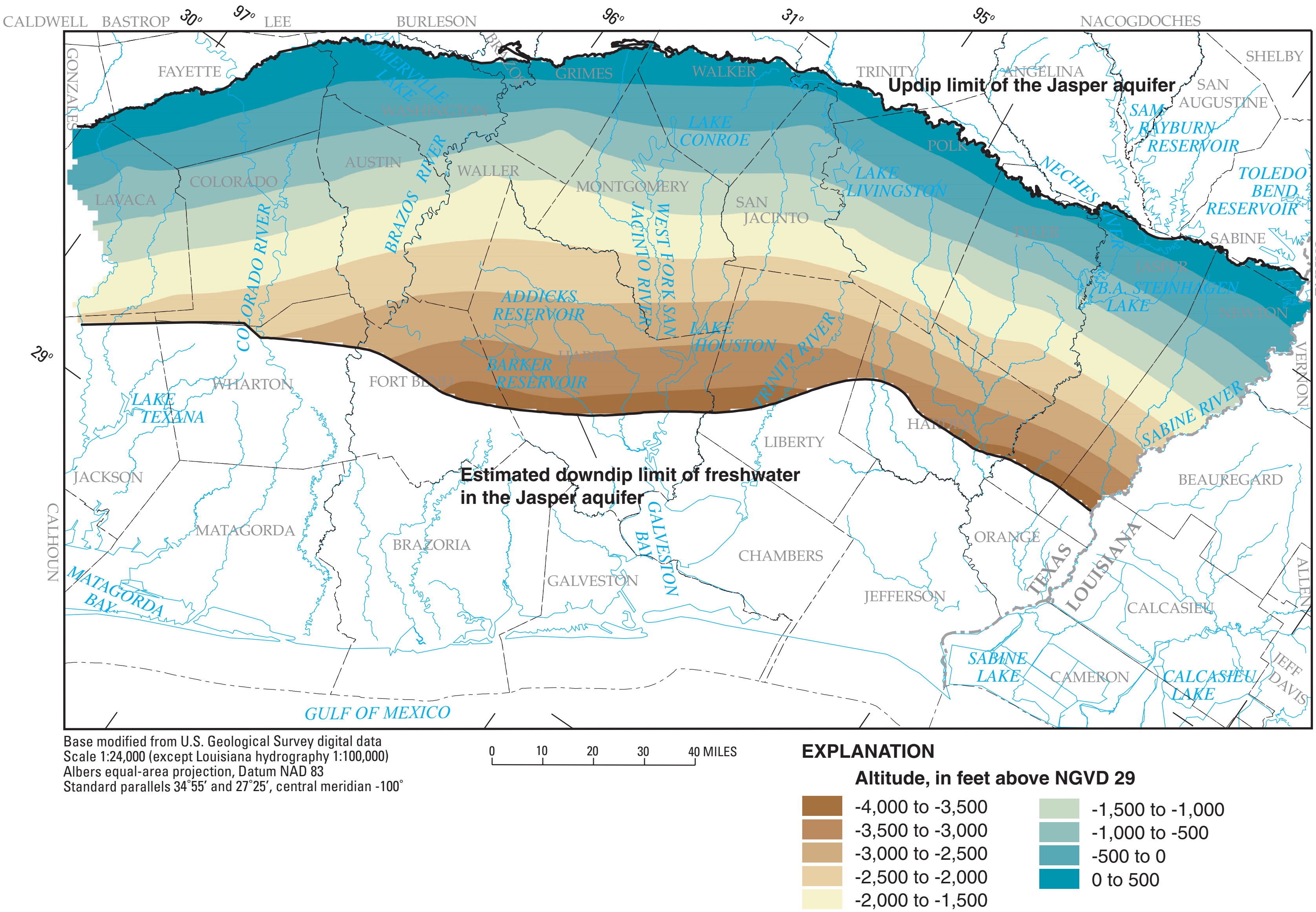


Figure 23. Altitude of the base of the Jasper aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003c).

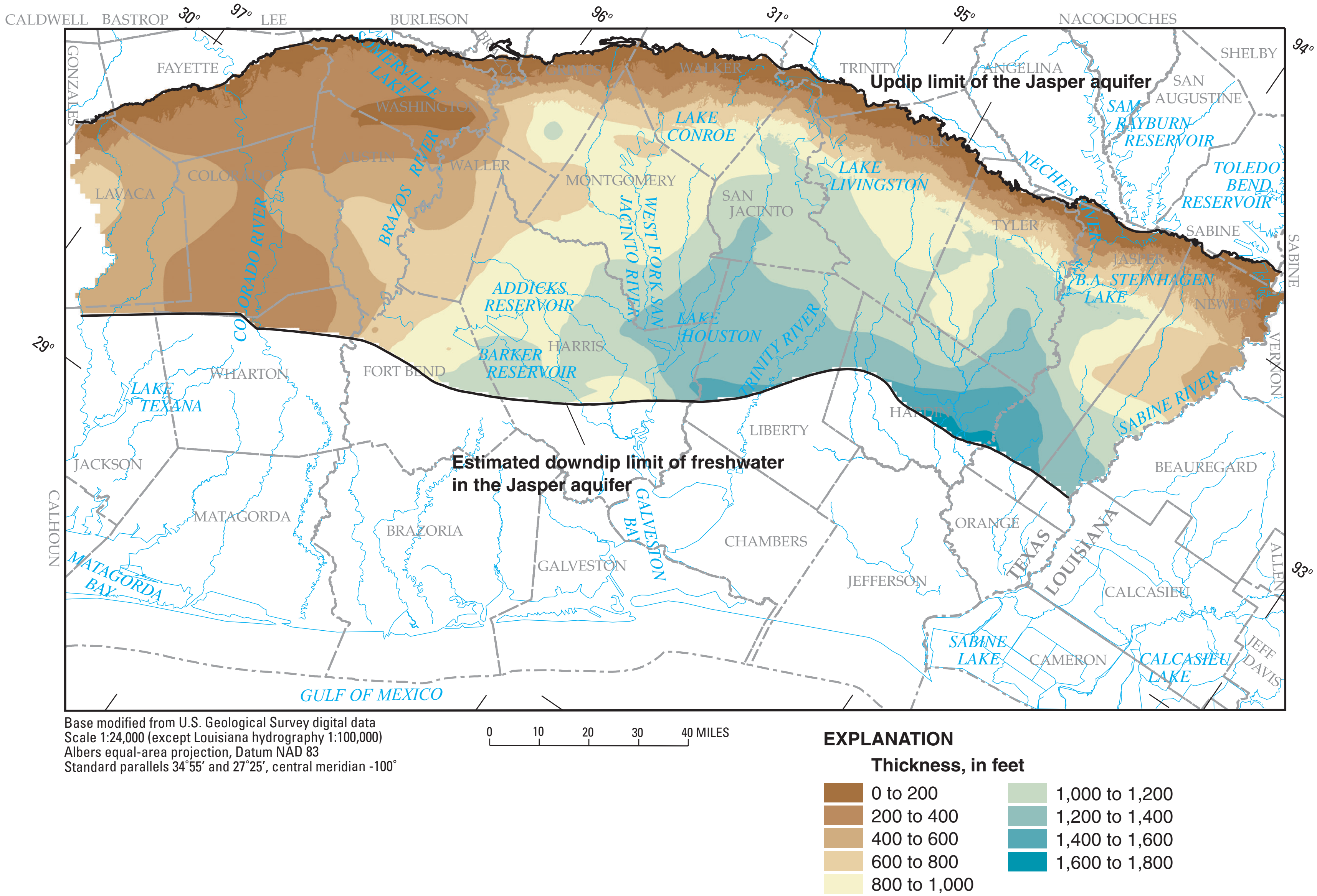
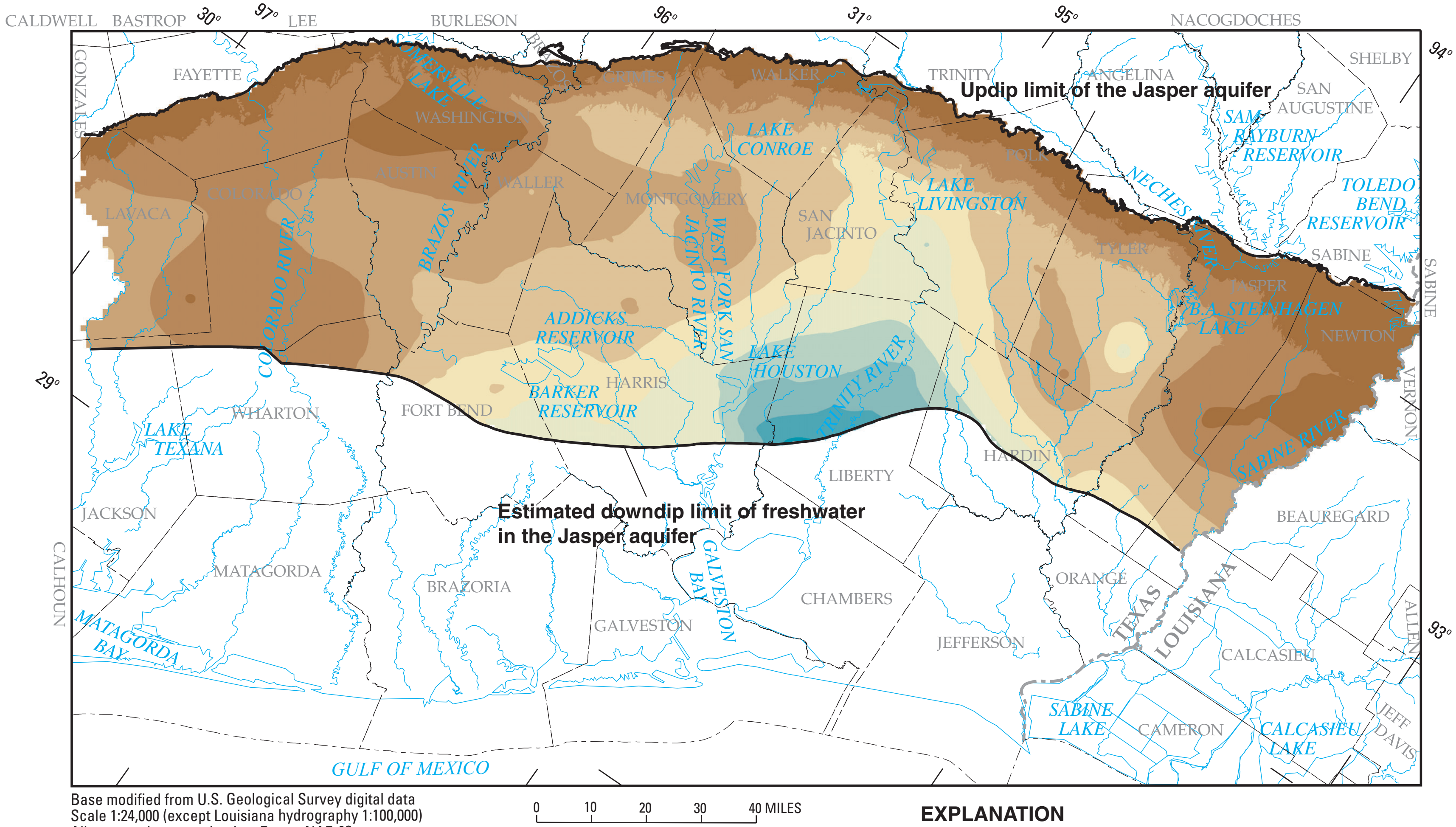


Figure 24. Thickness of the Jasper aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003c).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

Thickness, in feet

0 to 150	750 to 900
150 to 300	900 to 1,050
300 to 450	1,050 to 1,200
450 to 600	1,200 to 1,350
600 to 750	1,350 to 1,500

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 25. Cumulative clay thickness of the Jasper aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003c).

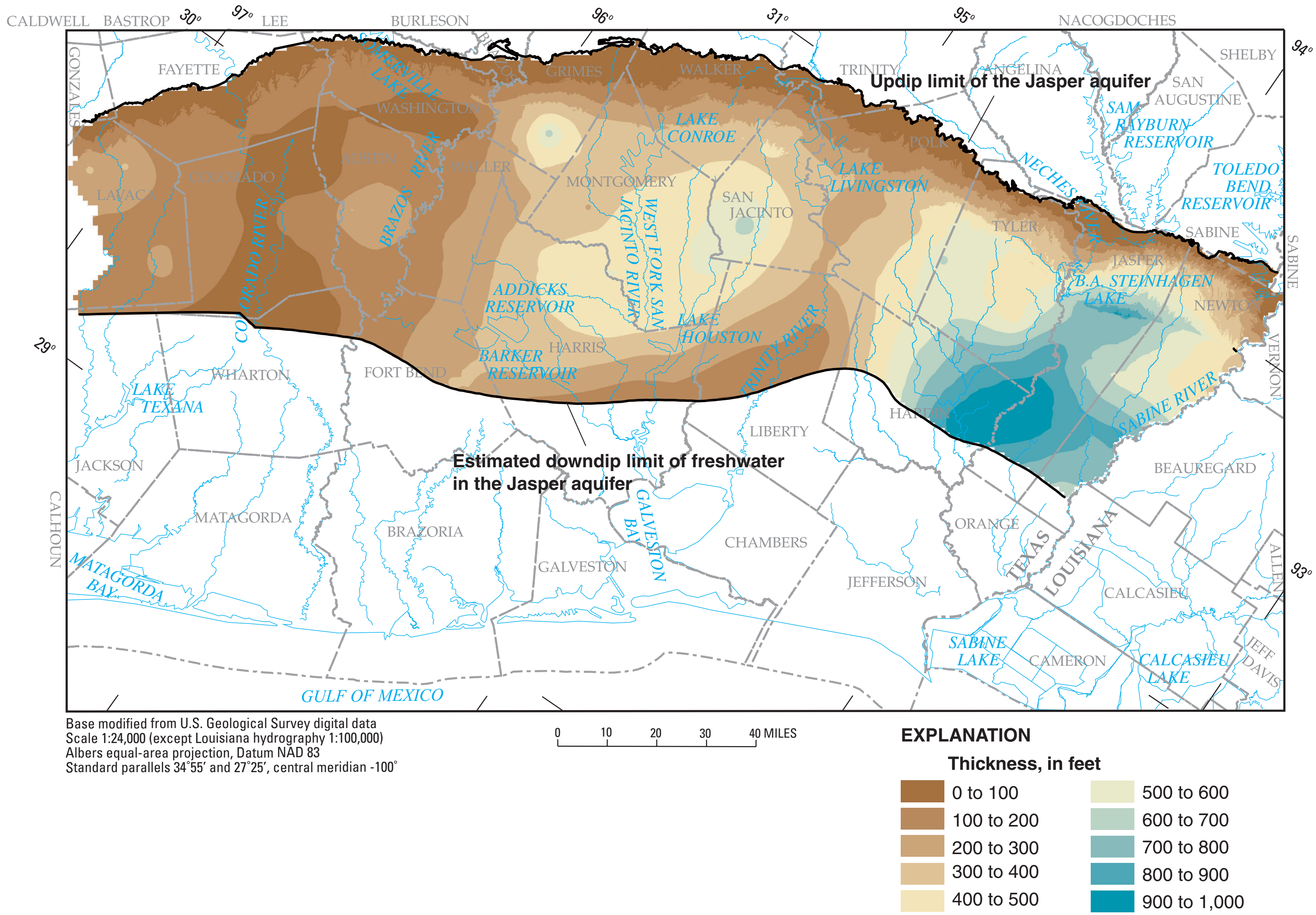
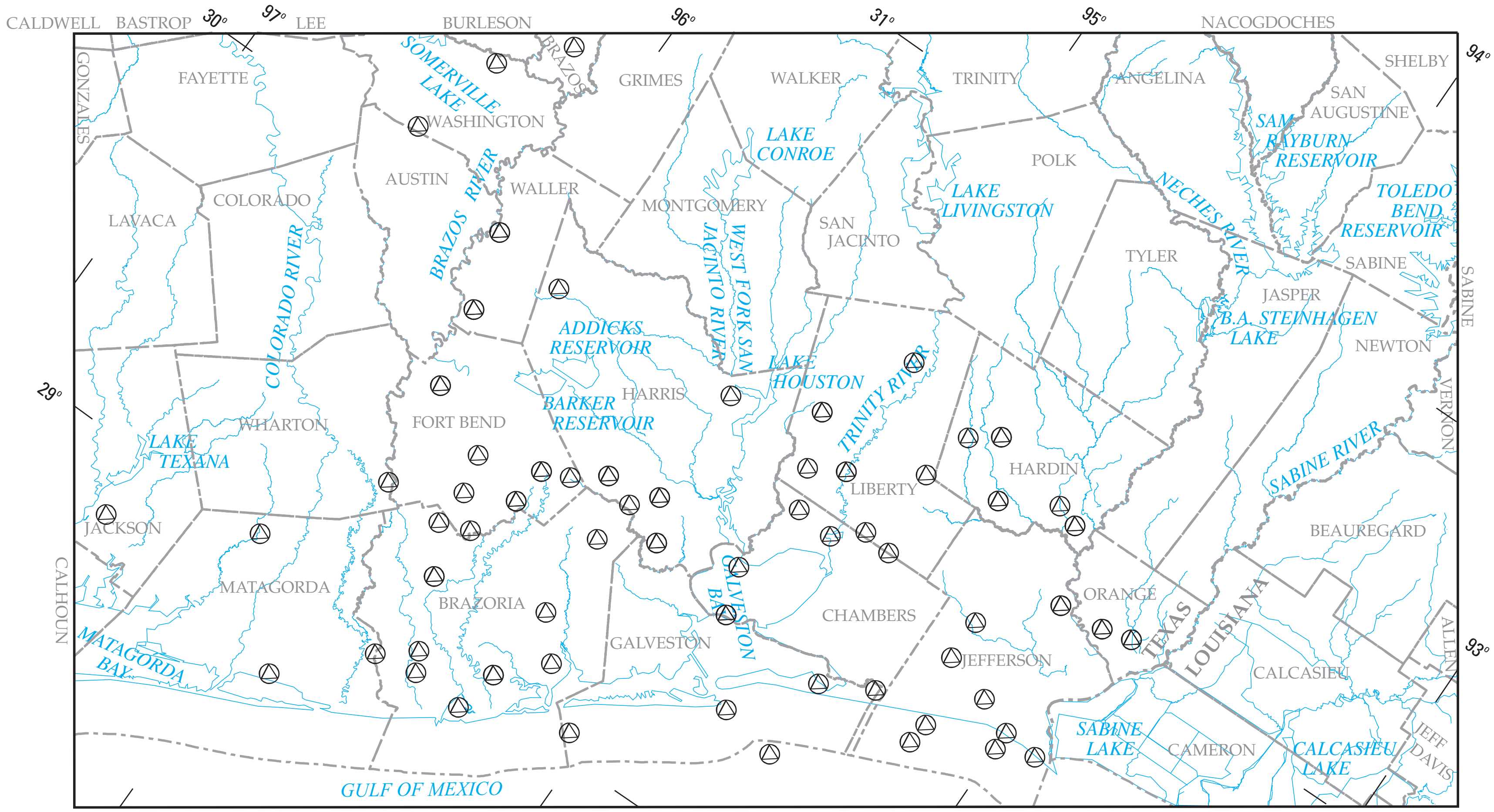


Figure 26. Cumulative sand thickness of the Jasper aquifer in the Ground-Water Availability Model area (modified from Strom and others, 2003c).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

EXPLANATION

△ Salt domes (diapirs)

Figure 27. Salt-dome locations in the Ground-Water Availability Model area (modified from Beckman and Williamson, 1990).

and the Jasper aquifer, 1,070 to 14,000 ft^2/d. Wesselman (1967) also estimated storativities of the Evangeline aquifer ranging from 6.3×10^{-4} to 1.5×10^{-3} and the Jasper aquifer ranging from 3.82×10^{-4} to 1.19×10^{-3}. Strom and others (2003c) reported storativities for the Jasper aquifer as large as 0.2.

Several other previous studies provide estimates of transmissivity for parts of or counties in the GAM area (for example, Jorgensen, 1975); those estimates generally are within the ranges listed above.

Ground-Water-Flow Conditions, Recharge, and Discharge

The uppermost parts of the aquifer system (shallow zones), which include outcrop areas, are under water-table conditions. As depth increases in the aquifer system and as interbedded sand and clay accumulate, water-table conditions evolve into confined conditions. Thus the lowermost parts of the aquifer system (deep zones) are under confined conditions. The middle parts of the aquifer system (intermediate zones) therefore are under semiconfined conditions. Because the transition from water-table to confined conditions with increasing depth is gradual, assigning specific depth horizons to shallow, intermediate, and deep zones is problematic.

As first described by Tóth (1963) and summarized relative to regional aquifer systems by Johnston (1999), natural (predevelopment) ground-water flow can be subdivided into local, regional, and intermediate flow systems. Local flow follows relatively short flow paths in shallow zones and is controlled mainly by topography. Recharge to local flow systems occurs in topographically high areas, and discharge occurs in nearby, topographically low areas. Regional flow follows relatively long flow paths from regional recharge areas through deep zones to distant discharge areas such as the downgradient limits of an aquifer system. Intermediate flow follows flow paths from recharge areas through intermediate zones to downgradient discharge areas. Although implied, to assume an exact, one-to-one correspondence between shallow (water-table) zones and local flow systems, deep (confined) zones and regional flow systems, and intermediate (semiconfined) zones and intermediate flow systems, probably would be an oversimplification.

If this concept of subdividing natural ground-water flow is applied to the Gulf Coast aquifer system, the implications are that an appreciable amount of the precipitation that infiltrates the subsurface (total recharge) in the relatively topographically rugged outcrop areas of the hydrogeologic units joins local flow systems. Thus much of the total recharge enters and exits the shallow subsurface to streams and valleys within relatively small areas. A proportionally smaller amount of the total recharge joins intermediate flow systems, and an even smaller amount of the total recharge joins regional flow systems. Wood (1956, p. 30–33), in an early study of the availability of ground water in the Gulf Coast region of Texas, states that, "Within the rainfall belts of 40–50 inches per year, probably 1 inch or more of the water that enters the outcrop of the aquifers updip from the heavily pumped areas is discharged to the streams in the outcrop area as base flow or rejected recharge."

The natural ground-water-flow system in places (Houston area, for example) has been greatly altered by decades of substantial withdrawals from deep zones. In such places, increased vertical head gradients have induced downward flow from local and intermediate flow systems into the regional flow system, thus capturing some flow that would have discharged naturally.

Few studies that focus specifically on recharge to the system in the GAM area are available. One such study involved use of environmental tritium as a ground-water tracer to estimate the total recharge rate in outcrops of the Chicot and Evangeline aquifers near Houston (Noble and others, 1996). The estimated total recharge rate from that study was 6 inches per year (in/yr). That rate, an estimated average for 1953–90, is considered an upper bound because it is based on the deepest penetration of tritium among 41 sampled wells. A study of potential recharge in the Houston area (R.K. Gabrysch and Fred Liscum, U.S. Geological Survey [retired], written commun., 1995) based on 30-year water-budget computations indicated that about 7 in/yr more precipitation was retained in four stream basins in the outcrops of the Chicot and Evangeline aquifers than in two stream basins atop confined areas of those aquifers, thus implying that the potential recharge rate was as much as 7 in/yr. Both of these studies were in areas likely influenced by withdrawals.

Loskot and others (1982, p. 29) report estimates of "potential recharge," which includes outcrop recharge that discharges to streams, for the Chicot and Evangeline aquifers in Colorado, Lavaca, and Wharton Counties. For the Chicot aquifer, they estimated that about 78,000 acre-feet per year (acre-ft/yr) is available as natural recharge in an outcrop area of about 1,100 mi^2, which is equivalent to about 1.3 in/yr. For the Evangeline aquifer, they estimated that about 38,000 acre-ft/yr is available as natural recharge in an outcrop area of about 600 mi^2, which is equivalent to about 1.2 in/yr.

Tarver (1968, p. 25–26) estimated a recharge rate of about 2 in/yr in outcrops of the Evangeline and Jasper aquifers in Polk County would be necessary to sustain the estimated rate of ground-water flow in the county in those aquifers.

Sandeen (1972, p. 36) provided two estimates of recharge to the Evangeline and Jasper aquifers in Washington County. To sustain the estimated amount of ground-water flow through the county, a recharge rate of about 0.3 in/yr on the aquifer outcrops would be required. Total recharge, which includes rejected recharge (that which discharges to streams), was estimated to be 1.2 in/yr.

Other studies that provide estimates of recharge to the system in the GAM area are based on simulation. Those studies indicate recharge rates over much more of the GAM area than the studies referred to above, but recharge rates derived from ground-water-flow models that require discretization of the simulated area have a disadvantage. The disadvantage, referred to as "the scale problem" by Johnston (1999), is that the fraction of total recharge (and thus total ground-water flow) that a model simulates decreases as the size of model grid blocks increases.

When grid blocks are large (for example, the block size of the Texas Gulf Coast aquifer system model of Ryder and Ardis [2002] is 25 mi^2), local and possibly some intermediate flow that enters and exits the physical system within the area encompassed by a single grid block cannot be simulated except by superimposing sources or sinks; thus only the deeper, more regional flow is simulated. As block size decreases, more of the local and intermediate flow is simulated. The relation between grid-block size and the fraction of total recharge (or flow) that a model simulates is unknown. The relation probably is most strongly influenced by topography and drainage density, which typically vary spatially across regional areas, thus further complicating the problem.

The fact that total recharge rates are underestimated in regional simulations lends perspective when considering estimates of recharge from simulation. Ryder and Ardis (2002, pl. 3) show areal distributions of simulated recharge to the entire Gulf Coast aquifer system in Texas for predevelopment and for 1982. Predevelopment recharge to the system in the GAM area ranges from 0 to 1 in/yr over most of the recharge area, 1 to 2 in/yr in a series of very small (relative to the total recharge area), isolated areas, and 2 to 3 in/yr in one area of a few square miles on the Austin-Colorado County line. For 1982, simulated rates of 5 to 6 in/yr are shown in topographically flat areas of Wharton and Jackson Counties (owing to withdrawals for irrigation [mostly rice]) and rates of 4 to 5 in/yr in areas of Harris, Fort Bend, and Waller Counties (owing to withdrawals for irrigation and public supply). The areas of 1 to 2 in/yr have enlarged and coalesced around the areas of 4 to 6 in/yr, but the most prevalent range for 1982 (range accounting for the largest area) remained 0 to 1 in/yr.

Ryder and Ardis (2002, fig. 9) show simulated recharge to the entire Gulf Coast aquifer system in Texas for predevelopment of 85 million cubic feet per day (ft^3/d), which is equivalent to 0.12 in/yr over the entire 114,000 mi^2 modeled area. Ryder and Ardis (2002, fig. 12) show simulated net recharge to the entire Gulf Coast aquifer system in Texas for 1982 of 179 million ft^3/d, which is equivalent to 0.25 in/yr over the entire 114,000 mi^2 modeled area.

Dutton and Richter (1990, figs. 42, 54) show simulated recharge for the Chicot and Evangeline aquifers in Wharton, Matagorda, and parts of adjacent counties, an area of irrigation withdrawals, for selected conditions including predevelopment and 1985. Recharge rates of as much as 0.4 in/yr were obtained in a small upland area east and west of the Colorado River for both predevelopment and 1985 conditions. The order-of-magnitude differences in recharge rates in the area under developed conditions between Ryder and Ardis (2002) and Dutton and Richter (1990) appear to be the result of a difference in simulation approach: Ryder and Ardis (2002) assumed a stable (unchanging) shallow water table because of sufficient recharge by precipitation and irrigation return flow and thus applied confined storativities; therefore the source of most of the water to sustain pumpage was induced recharge. Dutton and Richter (1990) assumed unconfined to semiconfined storativities; therefore the source of most of the water to sustain pumpage was aquifer storage rather than induced recharge.

Baker (1986, fig. 15) shows the areal distribution of simulated recharge to the Jasper aquifer in its outcrop in nearly all of the GAM area for predevelopment conditions, which were similar to developed conditions because development of the Jasper was "relatively limited" (Baker, 1986, p. 24). Recharge rates range from about 0.25 in/yr over much of the outcrop to as much as 1.5 in/yr in a small area in Newton County. Baker (1986, p. 39) implicitly acknowledged the scale problem, stating that "* * * a large part of the precipitation that reaches the zone of saturation in the outcrop moves to streams where it is discharged as seepage and spring flow * * * [and] only a small quantity * * * becomes * * * recharge * * * that moves into the downdip part of the aquifer."

Natural discharge occurs by seepage to streams, evapotranspiration, and diffuse upward leakage in topographically low and downdip (coastal) areas. Simulation can provide estimates of discharge rates distributed over regional areas. However, because simulated recharge is less than total recharge, simulated discharge will be less than total discharge. Ryder and Ardis (2002, pl. 3) show areal distributions of simulated discharge from the Gulf Coast aquifer system throughout the GAM area for predevelopment and for 1982. Predevelopment discharge rates range from 0 to 1 in/yr over all of the discharge area except for two small areas along streams where the indicated range is 1 to 2 in/yr. For 1982, simulated rates range from 0 to 1 in/yr throughout the discharge area. The most noticeable change between simulated predevelopment and 1982 discharge is the decrease in size of the discharge area compared to the recharge area, evidence that ground-water development reduces natural discharge, in addition to inducing recharge.

Dutton and Richter (1990) show simulated discharge for the Chicot and Evangeline aquifers in Wharton, Matagorda, and parts of adjacent counties for predevelopment and for 1985. For both conditions, discharge rates are less than 0.1 in/yr. As with Ryder and Ardis (2002), the size of the discharge area decreased substantially between predevelopment and 1985.

Ground-Water Development

Rates of recharge to and discharge from the Chicot, Evangeline, and Jasper aquifers are affected by ground-water withdrawals from the aquifers. "Predevelopment" relative to the GAM refers to aquifer conditions before 1891 or before the aquifers were measurably stressed by ground-water withdrawals; "postdevelopment" refers to aquifer conditions after the stress of withdrawals became measurable.

One of three principal areas of concentrated ground-water withdrawals from the aquifer system in the GAM area is Harris and Galveston Counties (the Houston area). Much of the early ground-water-use information for the area as summarized here is from Lang and Winslow (1950) and Wood and Gabrysch (1965).

Houston was founded in 1836 and initially used surface water to meet water-supply demands. In 1886, the first well was drilled to a depth of 140 ft and was reported as free flowing at more than 1,000 gallons per minute (gal/min). By 1905, as population and water demand increased, 65 wells ranging from 115 to 1,130 ft deep in the Chicot or Evangeline aquifers, or both, were in production. In 1906, the City of Houston had the capacity to supply as much as 19 million gallons per day (Mgal/d) of water, of which only 11 Mgal/d was actually used. By 1935, ground-water withdrawals averaged 24.5 Mgal/d, and by 1941, had increased to 27.2 Mgal/d. From 1941 to 1950, ground-water use more than doubled.

In 1954, water released from the newly constructed Lake Houston began to be used to augment ground-water supplies. The additional surface-water supply resulted in reduced ground-water withdrawals from 1954 to 1960. From the early 1960s to the mid-1970s, ground-water withdrawals increased at rates comparable to pre-1954 rates.

In 1975, because of increasing ground-water withdrawals and subsequent land-surface subsidence in Harris and Galveston Counties, the HGCSD was created to control land-surface subsidence by regulating ground-water withdrawals. In late 1976, ground-water withdrawals began to decrease in eastern Harris County because part of the demand began to be supplied by water from Lake Livingston. The policies of the newly created HGCSD resulted in decreased ground-water withdrawals in the Baytown and southeastern Harris County areas. The ground-water withdrawal rate exceeded 450 Mgal/d in 1976 and decreased to about 390 Mgal/d in the early 1980s; but then the trend reversed, and by 1990 withdrawals had increased to 493 Mgal/d. A downward trend began again in the 1990s, and withdrawals were about 463 Mgal/d by 1996.

The second principal area of withdrawals is the coastal irrigation area centered in Wharton and Jackson Counties but also extending into adjacent counties. Most of the irrigation withdrawals are from the Chicot aquifer for rice. Loskot and others (1982) recount the history of withdrawals in the area to the mid-1970s. As in the Houston area before appreciable development, wells flowed, but by the mid-1940s, most wells had ceased flowing. Withdrawals increased sharply in the early-to-mid-1950s with the introduction of the two-crop rice season, coupled with a period of below-normal precipitation. By the late 1960s, irrigation withdrawals in Wharton County (97 percent from the Chicot aquifer), which historically account for about 70 to 80 percent of the irrigation total for the area, had reached 172 Mgal/d. The irrigation withdrawal rate in Wharton County was about 155 Mgal/d in 1985, about 121 Mgal/d in 1990, about 129 Mgal/d in 1995 (U.S. Geological Survey, 2004), and about 183 Mgal/d in 2000 (D.L. Barbie, U.S. Geological Survey, written commun., 2004).

The third principal area of withdrawals is the Evadale-Beaumont area. Industrial withdrawals associated with wood-pulp processing at Evadale in southwestern Jasper County began in 1955. The initial withdrawal rate, about 18 Mgal/d from the Evangeline aquifer, increased to more than 45 Mgal/d by early 1965 (Wesselman, 1967, p. 46). Public-supply withdrawals from the Beaumont well field in southeastern Hardin County began about 1958. By 1965, Beaumont was withdrawing 6 Mgal/d (Ryder and Ardis, 2002, p. E33). The combined Evadale-Beaumont withdrawals from the Chicot and Evangeline aquifers for 1977 were about 24 Mgal/d; by 2000, the rate had increased to about 44 Mgal/d (Texas Water Development Board, written commun., 2003).

Potentiometric Surfaces and Land-Surface Subsidence

In the updip outcrop area of the Chicot aquifer and the outcrop areas of the Evangeline and Jasper aquifers and Burkeville confining unit (figs. 8, 14, 21, 20), water-table conditions generally exist. The water table is assumed to be a subdued replica of the topography (Williams and Williamson, 1989). In outcrops of the Chicot and Evangeline aquifers in parts of Harris and Montgomery Counties, seismic refraction has indicated that the water table ranges from about 10 to 30 ft below land surface (Noble and others, 1996). Hydrographs indicate that the water table in the Chicot and Evangeline aquifers, where not influenced by pumping wells, has remained fairly stable (fig. 28).

The USGS annually has measured water levels and constructed potentiometric surfaces of the Chicot and Evangeline aquifers in the greater Houston area since 1977 and of the Jasper aquifer since 2000. For example, the potentiometric-surface map of the Chicot aquifer, January–February 2000 (Coplin and Santos, 2000), shows a range in water-level altitudes from 150 ft above NGVD 29 in northwestern Harris County and southeastern Waller County to 200 ft below NGVD 29 in north-central Harris County (fig. 29). The potentiometric-surface map of the Evangeline aquifer, January–February 2000, shows a range in water-level altitudes from 200 ft above NGVD 29 in southwestern Montgomery County and southeastern Waller County to 400 ft below NGVD 29 in west-central Harris County (fig. 30). The potentiometric-surface map of the Jasper aquifer, spring 2000 (Coplin, 2001), shows a range in water-level altitudes from 250 ft above NGVD 29 in Grimes and northern Montgomery Counties to 50 ft below NGVD 29 in south-central Montgomery County (fig. 31). The small areal extent of the Jasper potentiometric surface reflects a scarcity of water-level data compared to the amount of data available for the Chicot and Evangeline aquifers.

In the GAM area away from the Houston area, no periodic potentiometric surfaces are constructed, as measured synoptic water-level data are few.

Before appreciable ground-water withdrawals from the system in the GAM area, the potentiometric surfaces in the confined parts of the aquifers were higher than land surface in places. Ground-water development has caused substantial (as much as 350 ft) declines of the potentiometric surfaces of the aquifers (and subsequent land-surface subsidence) primarily in the Houston area, as will be shown in the “Model Calibration” section.

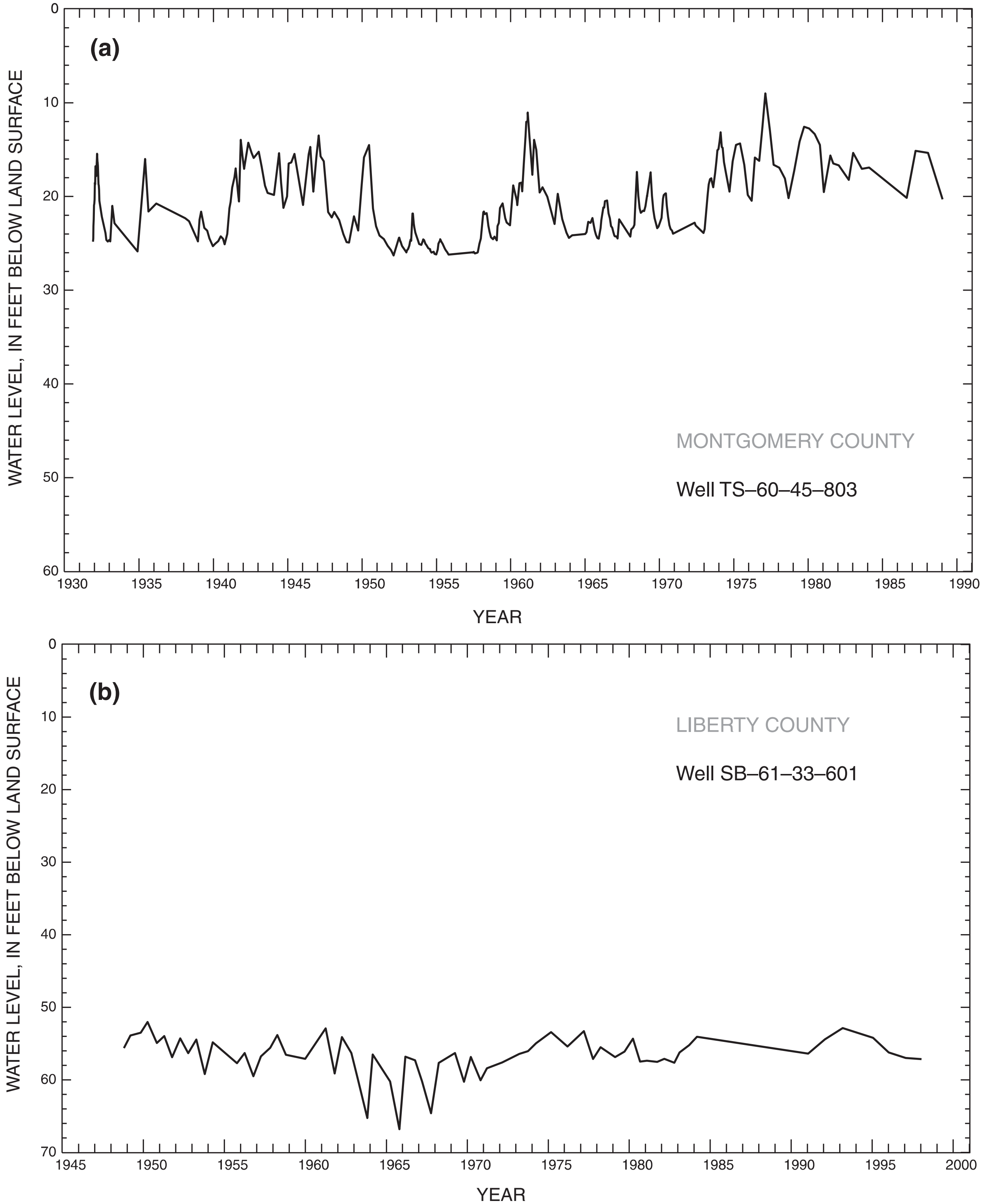
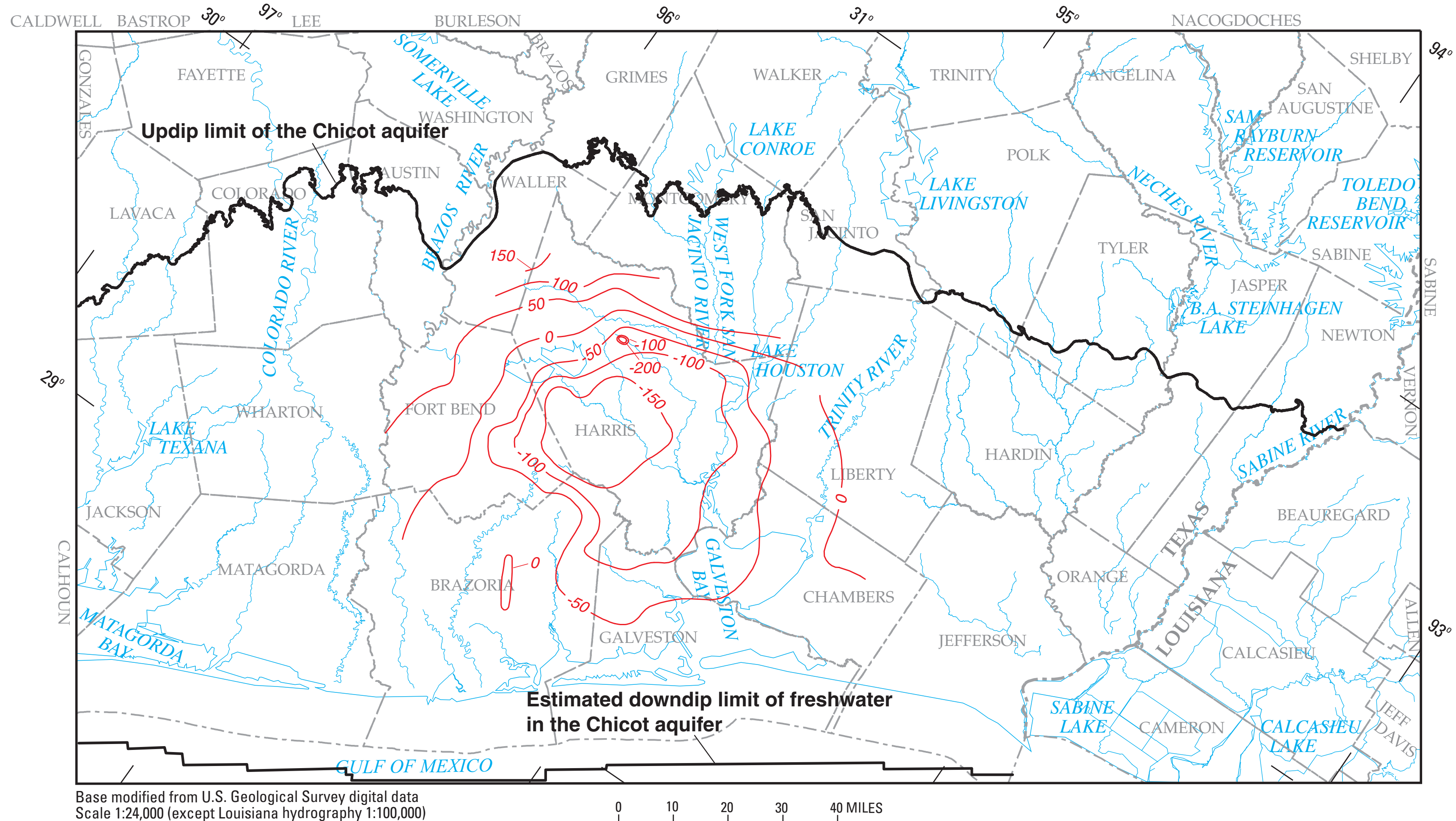


Figure 28. Hydrographs of wells in the Ground-Water Availability Model area screened in the outcrops of (a) the Chicot aquifer in Montgomery County and (b) the Evangeline aquifer in Liberty County.

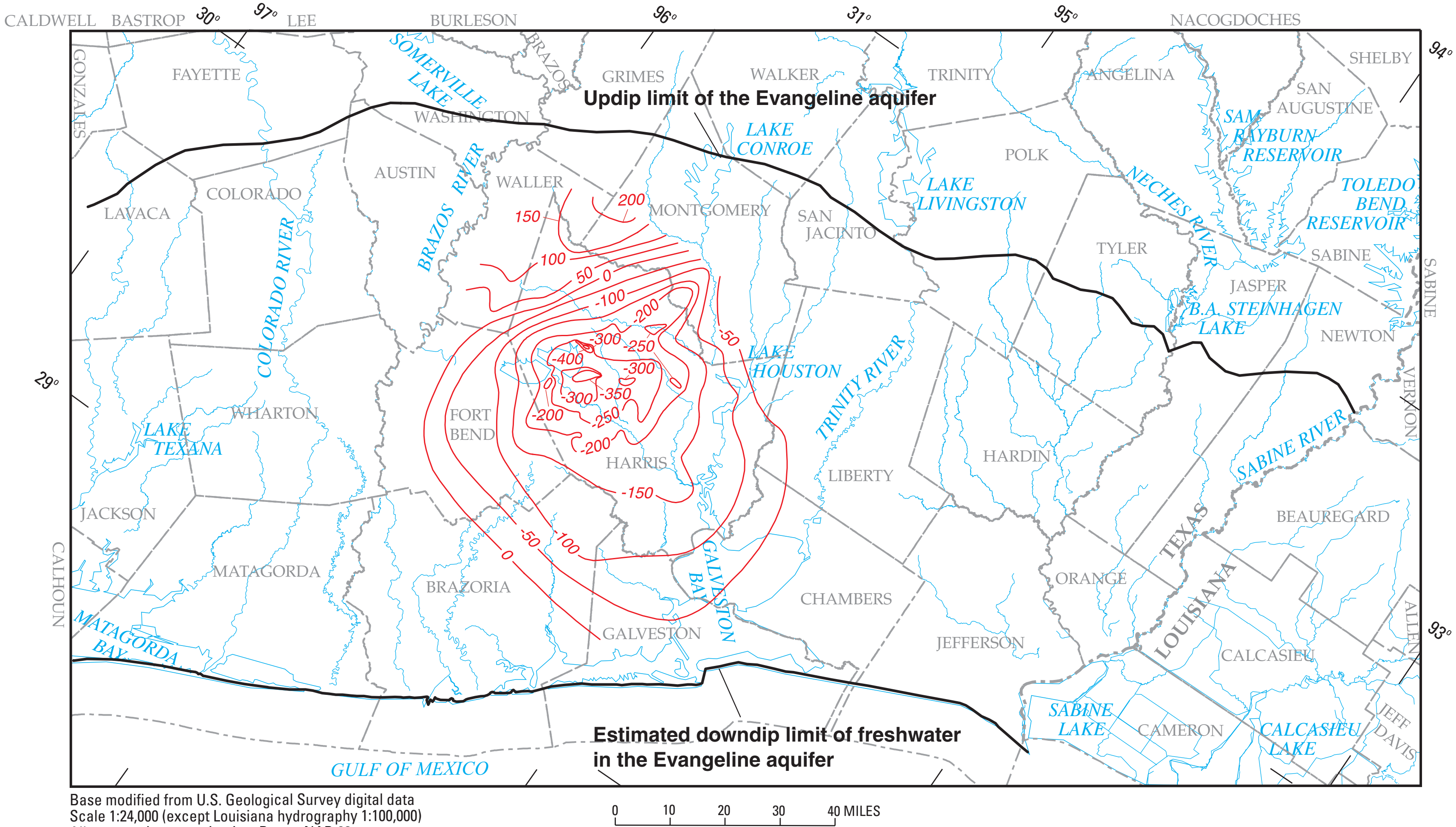


Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

—— -50 —— **Potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Intervals 50 and 100 feet. Datum is NGVD 29

Figure 29. Measured 2000 potentiometric surface of the Chicot aquifer in the Ground-Water Availability Model area (modified from Coplin and Santos, 2000, fig. 1).

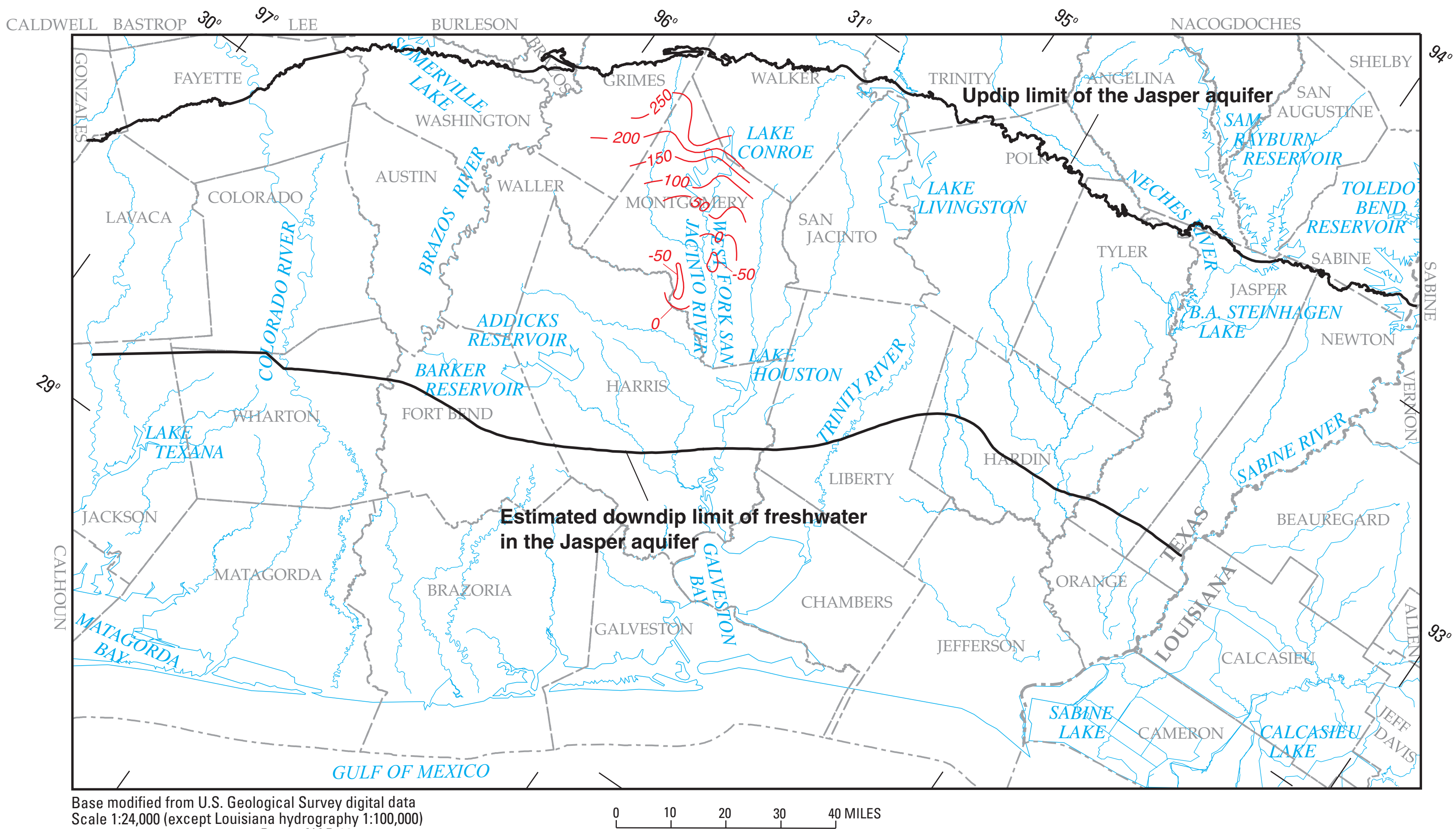


Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

-50 **Potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Intervals 50 and 100 feet. Datum is NGVD 29

Figure 30. Measured 2000 potentiometric surface of the Evangeline aquifer in the Ground-Water Availability Model area (modified from Coplin and Santos, 2000, fig. 4).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

—— -50 —— **Potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Intervals 50 and 100 feet. Datum is NGVD 29

Figure 31. Measured 2000 potentiometric surface of the Jasper aquifer in the Ground-Water Availability Model area (modified from Coplin, 2001).

Although appreciable amounts of water have been withdrawn from the Chicot aquifer in the coastal irrigation area for decades, relatively little long-term drawdown (tens of feet) has occurred there. Rice-irrigation return flow (by one estimate [Tuck, 1974, *in* Loskot and others, 1982, p. 33] as much as 30 percent) and withdrawals from relatively shallow zones under water-table conditions that are readily recharged probably have helped to lessen long-term water-level declines in the area.

If recent synoptic water-level measurements were available in the Evadale-Beaumont area, cones of depression caused by withdrawals undoubtedly would appear. Ryder and Ardis (2002, p. 33) estimate 150 to 200 ft of drawdown in the Evangeline aquifer centered at Evadale in 1982.

Potentiometric-surface declines in unconsolidated confined aquifers cause a decrease in hydraulic pressure that creates a load on the skeletal matrix of the aquifer. Because the sand layers are more transmissive than the clay layers, the depressuring of the sand layers is relatively rapid, causing only slight skeletal matrix consolidation of the sand layers. However, the depressurizing and subsequent dewatering of the clay layers require more time compared to that of the sand layers and are dependent on the thickness and hydraulic characteristics of the clay layers as well as the vertical stress of the sediment overburden. The delayed drainage of the clay layers continues to occur until the excess (transient) pore pressure in the clay layers equals the pore pressure of the adjacent sand layers. Until pressure equilibrium is attained, dewatering of the clay layers continues to apply a load to the skeletal matrix of the clay layers. This loading process is similar to what occurs in the sand layers; but additionally, the orientation of the individual clay grains changes, becoming perpendicular to the applied vertical load. Therefore, the dewatering caused by the depressurization of the clay layers combined with clay-grain realignment reduces the porosity and ground-water-storage capacity of the clay layers, which in turn allows them to compact.

Because of the weight of the overburden and the inelastic compaction characteristics of the clay layers, about 90 percent of the compaction is permanent (Gabrysch and Bonnet, 1975). Thus, when potentiometric surfaces rise and repressure compacted clay layers there is little, if any, rebound of the land surface (Gabrysch and Bonnet, 1975). Although the compaction of one clay layer generally will not cause a noticeable decrease in the land-surface altitude, if numerous stacked clay-layer sequences (which are characteristic of the Gulf Coast aquifer system) depressure and compact, then appreciable decreases in land-surface altitude can and do occur (Gabrysch and Bonnet, 1975). More than 10 ft of land-surface subsidence has been documented in the Baytown and Houston Ship Channel area in southwestern Harris County (Harris-Galveston Coastal Subsidence District, 1998), as will be shown in the "Model Calibration" section. Subsidence of smaller but still destructive magnitudes has occurred in places throughout most of Harris County and to a lesser extent in parts of Galveston and Fort Bend Counties.

A substantial amount of the total water withdrawn is derived from the dewatering of the numerous clay layers of the aquifer. As early as 1959, Winslow and Wood (1959, p. 1,034) computed about one-fifth of the water withdrawn from wells in the Katy-Houston-Pasadena-Baytown area during 1954–59 was derived from compaction of clays. Wood and Gabrysch (1965, p. 16) considered water derived from compaction in construction of the first analog model of the ground-water-flow system but estimated only 1 percent of the water withdrawn by wells was derived from compaction of clays. Later, Jorgensen (1975, p. 49) showed water derived from compaction ranged from 17 to 22 percent of the water withdrawn from wells for different periods. Most recently, model simulations indicated that as much as 19 and 10 percent of the total water budget of the Chicot and Evangeline aquifers, respectively, is derived from the dewatering of the clay layers of the aquifers (Kasmarek and Strom, 2002).

Simulation of Ground-Water Flow and Land-Surface Subsidence in the Northern Part of the Gulf Coast Aquifer System

Model Description

A numerical model was developed to simulate ground-water flow and land-surface subsidence in the northern Gulf Coast aquifer system from predevelopment through 2000. The finite-difference computer code MODFLOW96 (Harbaugh and McDonald, 1996) was used in this application. The Interbed-Storage Package designed for the MODFLOW model (Leake and Prudic, 1991) was used to simulate clay compaction and storage, and thus land-surface subsidence, in the Chicot and Evangeline aquifers. The Chicot, Evangeline, and Jasper aquifers and the Burkeville confining unit were simulated as four separate layers and discretized into two-dimensional finite-difference grids. Using GIS applications, model input data were georeferenced and assigned to model grid blocks.

Mathematical Representation

The MODFLOW model uses finite-difference methods to solve the partial differential equation for three-dimensional movement of ground water of constant density through heterogeneous, anisotropic porous materials. The equation can be written as

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right) - W = S_s\frac{\partial h}{\partial t}, \qquad (1)$$

where

K_{xx}, K_{yy}, and K_{zz} = hydraulic conductivity along the x, y, and z coordinate axes, which are assumed parallel to the major axes of hydraulic conductivity [Lt^{-1}],

S_s = specific storage [Lt^{-1}],

W = source or sink term [Lt^{-1}],
h = hydraulic head [Lt^{-1}], and
t = time

(Harbaugh and McDonald, 1996). This equation, with specification of appropriate boundary and initial conditions, constitutes a mathematical representation of the ground-water flow system. In this application, the aquifer system was assumed to be horizontally isotropic; thus there was no preferred direction of hydraulic conductivity in the horizontal.

Grid Design

The finite-difference grid for the numerical model (fig. 32) covers 33,565 mi^2 in southeastern Texas and southwestern Louisiana. The model grid was rotated 37.6 degrees clockwise so that the orientation of the model was parallel to the Texas Gulf Coast. This rotation allowed the grid axes to more closely coincide with natural ground-water divides, model boundaries, and predevelopment and postdevelopment flow paths. The four layers of the model together contain 134,260 grid blocks. Each layer consists of 137 rows and 245 columns. Layer 1 represents the Chicot aquifer, layer 2 the Evangeline aquifer, layer 3 the Burkeville confining unit, and layer 4 the Jasper aquifer. The grid blocks are uniformly spaced, and the area of each block is 1 mi^2.

Boundaries

Model boundaries control where and how much water enters and leaves the simulated aquifer system. The selection of model boundaries for the aquifers in this model was based on a conceptual interpretation of the flow system developed using information reported by Meyer and Carr (1979), Carr and others (1985), Williamson and others (1990), and data supplied by the TWDB.

Lateral and Base of System

The northwestern boundaries of the three aquifers and the Burkeville confining unit are the northwestern extent of the updip outcrop sediments for each unit (figs. 8, 14, 20, 21). Northwest of these boundaries, the model grid blocks were assigned zero transmissivity to simulate no-flow boundaries.

The downdip limit of freshwater (defined for this study as the location where the dissolved solids concentration is as much as 10,000 milligrams per liter [mg/L]) was chosen as the southeastern boundary of flow in each hydrogeologic unit. Southeast of these limits, the model grid blocks were assigned zero transmissivity to simulate no-flow boundaries. The location of the 10,000-mg/L line in each hydrogeologic unit was estimated from geophysical log data and (for the Evangeline aquifer) from the coastward extent of freshwater withdrawals.

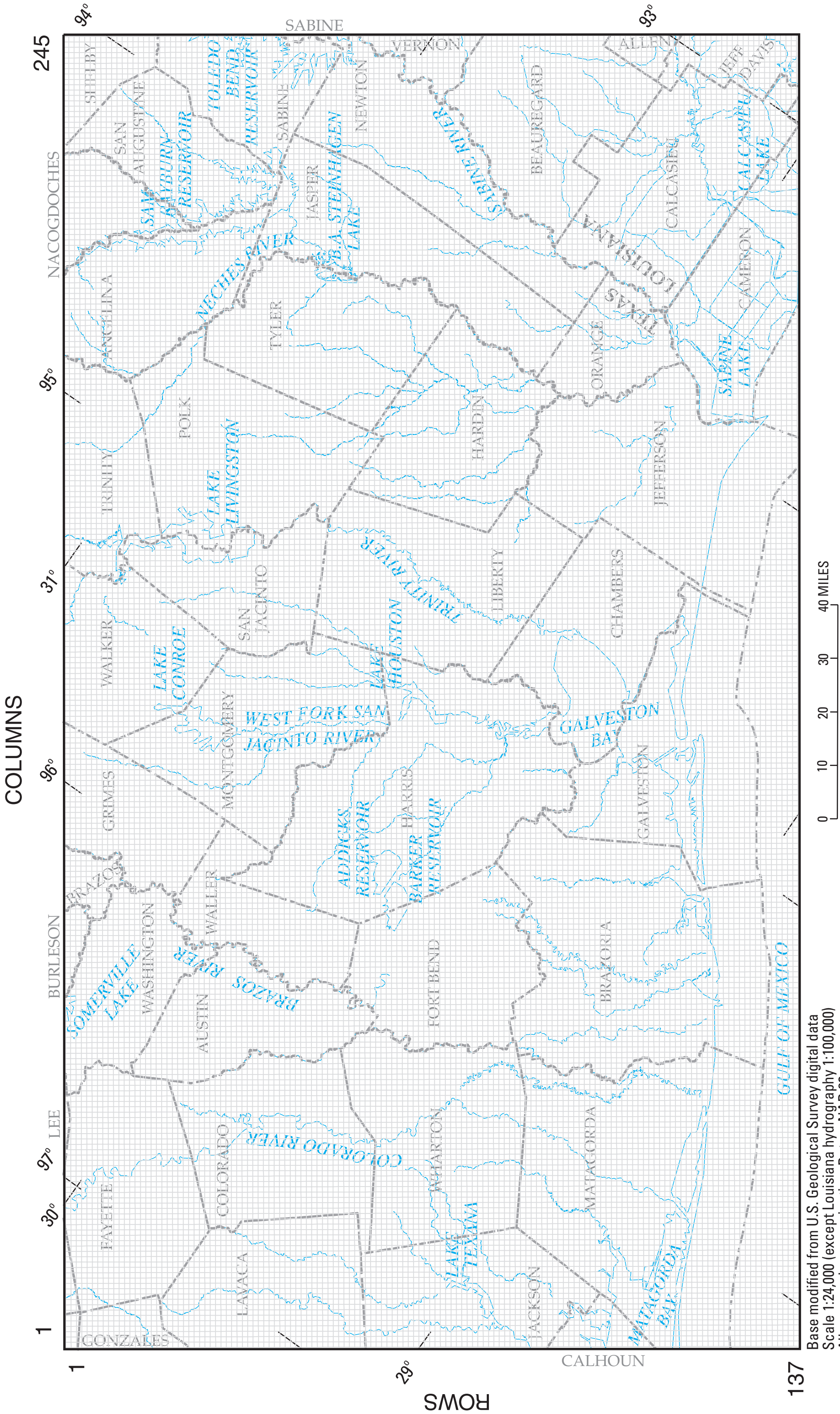
A no-flow boundary at a specified location reflects an assumption of a stable downdip freshwater/saline-water interface. Along the coast in most of the GAM area, this assumption probably is valid. Little or no human-induced stresses on the aquifer system in most of the coastal region likely have allowed long-term equilibrium to be established between the freshwater and the slightly more-dense saline water that lies laterally adjacent to and beneath the freshwater. However, in the Houston-Galveston area, reduced freshwater heads caused by withdrawals have induced saline-water encroachment in the Chicot and Evangeline aquifers in places, as noted by several previous investigators and summarized by Ryder and Ardis (2002). Such encroachment was not simulated in the GAM for two reasons: The first is that the MODFLOW model does not have the ability to simulate variable-density flow, and the second is that data are lacking to indicate whether the documented encroachment in the Houston-Galveston area represents regional-scale changes in the locations of the interfaces of the aquifers during the decades of ground-water development. Although simulating downdip freshwater/saline-water interfaces in coastal areas as fixed boundaries in regional-scale finite-difference models is common practice (for example, Bush and Johnston [1988], Mallory [1993], Arthur [1994], Barker and Pernik [1994], Strom and Mallory [1995], and Strom [1998]), the inability to simulate movement of a freshwater/saline-water interface is an acknowledged weakness of the GAM.

The southwestern-northeastern lateral boundaries for the Chicot, Evangeline, and Jasper aquifers and the Burkeville confining unit were selected to coincide with ground-water-flow divides associated with major streams. The southwestern lateral boundary was located generally along the Lavaca River, and the northeastern lateral boundary was located along the Sabine River (fig. 4). The assumption is that little lateral flow occurs across these boundaries, and thus they can reasonably be simulated as no-flow boundaries.

The Jasper aquifer is underlain by the Catahoula confining unit. The assumption is that the unit sufficiently restricts the exchange of water between the Jasper aquifer and deeper units so that the Catahoula confining unit can reasonably be simulated as a no-flow base-of-system boundary.

Recharge and Discharge

The MODFLOW general-head boundary package was used to simulate recharge and discharge in the outcrops of the Chicot, Evangeline, and Jasper aquifers and the Burkeville confining unit. This package allows the water table of an aquifer system to function as a head-dependent flux (flow per unit area) boundary (Franke and others, 1987). That is, a condition in which the rate of flow between the water table and the adjacent deeper zone of the system is controlled by the difference between the water table (constant head) and the head in the adjacent deeper zone (which changes with time), and by the vertical hydraulic conductance between the water table and the immediately adjacent deeper zone. In interstream outcrop areas, the head differences likely are downward (recharge area), and in stream and downdip areas, the head differences likely are upward (discharge areas).



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

Figure 32. Finite-difference grid used for the Ground-Water Availability Model.

Simulating the water table as a constant-head source (or sink) of water to the system requires an assumption that no long-term trends in the water table are indicated, as in the example hydrographs in figure 28. This assumption is believed reasonable over most of the GAM area, although the assumption might not be valid in some areas of intense withdrawals.

Water-table-altitude data for the shallow zones of the hydrogeologic units from the model of Kasmarek and Strom (2002) were used for GAM grid blocks in areas where the two models are coincident. These water-table-altitude data were created using the method described by Williams and Williamson (1989) in which multiple linear regressions of depth-to-water data and topographic data were used to derive relations between depth to water and topography. For the GAM outside the area of coincidence with the model of Kasmarek and Strom (2002), water-table altitudes were estimated using GIS methods of Strom and others (2003a, b, c). These methods involved constructing an initial water table on the basis of topography (60-meter digital elevation model data in this application) and subtracting a "trend surface" (a dataset of measured depths to water supplemented by interpolated depths to water) to obtain water-table altitudes.

Flow between streams and the aquifer system (essentially discharge from aquifers to incised streams in outcrops) was not explicitly simulated by imposing sinks along streams (MODFLOW river package) in the model. The rationale for this decision is that the general-head boundary package, assuming the model is adequately calibrated, would account for stream discharge to the level of accuracy that such discharge is known. Few measured data are available on streamflow gains/loses for the major streams that flow across the outcrops of the Gulf Coast aquifers. A recent compilation of the results of historical streamflow gain-loss studies in Texas (Slade and others, 2002) lists total streamflow gains in GAM aquifer outcrops of about 10, 4, and 37 cubic feet per second (ft^3/s) for the West Fork San Jacinto, East Fork San Jacinto, and Trinity Rivers on the basis of data collected before 1970. Because aquifer discharge to streams is not well known, such data are not particularly helpful for comparison with simulated data for purposes of calibration; there was little incentive to add more complexity to an already complex model by explicitly computing flow between streams and aquifers. Ensuring that simulated discharge to streams is physically reasonable could be done by assessing the amount of overall discharge in stream areas, which is available using the general-head boundary package.

Initial Conditions

Initial conditions for head and hydraulic properties were prepared for the model area for input to MODFLOW. The initial values for hydraulic properties were then varied within reasonable ranges, as described in the "Model Calibration" section, to construct the calibrated model.

Heads

Distributions of head in each hydrogeologic unit for an initial predevelopment steady-state simulation were estimated on the basis of land-surface altitudes. Simulated predevelopment steady-state heads were used as starting heads for transient simulations from predevelopment to 1977 and to 2000.

Hydraulic Properties Associated With Ground-Water Flow

Initial transmissivity distributions for the aquifers were constructed with data from Wesselman (1967), Carr and others (1985), Baker (1986), and Kasmarek and Strom, (2002) using GIS applications. The initial transmissivity of the Burkeville confining unit was computed by multiplying values of hydraulic conductivity representative of a mid-range between silty sand and marine clay (average 0.01 foot per day [ft/d]) (Freeze and Cherry, 1979, table 2.2) by the areally distributed thickness of the confining unit.

For outcrop areas, the initial vertical hydraulic conductance between the water table and the immediately adjacent deeper zone was computed by dividing a constant vertical hydraulic conductivity by the cumulative clay thickness from land surface to the centerline of the outcropping hydrogeologic unit and multiplying by grid-block area. The vertical hydraulic conductivity for the computation of water table-Chicot and water table-Evangeline conductances was 0.001 ft/d; for the water table-Burkeville conductance, 5×10^{-5} ft/d; and for the water table-Jasper conductance, 5×10^{4} ft/d. These hydraulic conductivities were selected on the basis of published ranges of hydraulic conductivity (Freeze and Cherry, 1979, table 2.2) for the types of sediments that compose the hydrogeologic units. The initial water table-Chicot vertical hydraulic conductances ranged from negligible (at updip featheredge of outcrop) to 51,000 ft^2/d; initial water table-Evangeline, 46 to 139,000 ft^2/d; initial water table-Burkeville, 22 to 1,060 ft^2/d; and initial water table-Jasper, 38 to 13,900 ft^2/d.

For subcrop areas, vertical hydraulic conductance is computed internally by MODFLOW by multiplying a leakance by the grid-block area. Initial leakances in the GAM area coincident with the area of the model of Kasmarek and Strom (2002) were the calibrated leakances of that model. For the area of the system outside the Kasmarek and Strom (2002) model area, initial leakances were computed by dividing a constant vertical hydraulic conductivity by the cumulative clay thickness from centerline to centerline of adjacent hydrogeologic units. The vertical hydraulic conductivity for the computation of the Chicot-Evangeline leakance was 0.001 ft/d; for the Evangeline-Burkeville leakance, 5×10^{-5} ft/d; and for the Burkeville-Jasper leakance, 5×10^{4} ft/d. The initial Chicot-Evangeline leakances ranged from 1.2×10^{-7} to 5.0×10^{-3} foot per day per foot (d^{-1}); initial Evangeline-Burkeville leakances, 7.2×10^{-8} to 7.4×10^{-6} d^{-1}; and initial Burkeville-Jasper leakances, 6.2×10^{-7} to 9.0×10^{-6} d^{-1}.

Initial storativities of the sands in the Chicot and Evangeline aquifers are from Kasmarek and Strom (2002) in the areas where the model of that report is coincident with the GAM and from Carr and others (1985) in other areas of the GAM. Initial storativities of the sands for the Chicot and Evangeline aquifers ranged from 4×10^{-4} to 0.1 and from 5×10^{-4} to 0.1, respectively. The ranges of storativities reflect subsurface conditions from confined to semiconfined to water table.

The storativity of the sands for the Burkeville confining unit was derived by multiplying the sand thickness of that unit times 1×10^{-6}, a value that Lohman (1972) states is typical for specific storage of confined aquifers. Storativities thus derived range from 1.0×10^{-5} to 0.05 and again reflect confined through water-table conditions. Initial storativities of the Burkeville confining unit were not varied during model calibration.

The storativity of the sands for the Jasper aquifer are from Strom and others (2003c) augmented with available Jasper aquifer-test data (Wesselman, 1967). Confined through water-table storativities range from 2.0×10^{-5} to 0.2 and were not varied during calibration.

Land-Surface Subsidence and Storage in Clays

Simulation of land-surface subsidence (actually, compaction of clays) and release of water from storage in the clays of the Chicot and Evangeline aquifers was accomplished using the Interbed-Storage Package designed for use with MODFLOW developed by Leake and Prudic (1991). Compaction of clays in the Jasper aquifer and the Burkeville confining unit were not simulated because the sediments of those units are geologically older, more deeply buried, and therefore more consolidated relative to the sediments of the Chicot and Evangeline aquifers. Additionally, substantial potentiometric-surface declines such as have occurred in the Chicot and Evangeline aquifers in the greater Houston area have not occurred in the Jasper aquifer, and probably not in the Burkeville confining unit.

As explained in Leake and Prudic (1991), effective stress is defined as the difference between geostatic pressure (overburden load) and fluid pressure (head). Head decreases in a confined aquifer do not change geostatic pressure if, as assumed in this application, water-table heads remain constant. With constant geostatic pressure, effective stress thus will increase by the same amount that heads decrease. Previous studies (Riley, 1969; Helm, 1975) indicate that compaction (or expansion) of interbedded clays is proportional, or nearly so, to change in effective stress. So, for sediments in confined aquifers with constant geostatic pressure, compaction also is proportional, or nearly so, to change in head. The relation is

$$\Delta b = -\Delta h S_s b_o , \qquad (2)$$

where

Δb = amount of compaction or expansion [L],

Δh = change in head [L],

S_s = skeletal component of elastic or inelastic specific storage [L^{-1}], and

b_o = thickness of the interbed [L]

(modified from Leake and Prudic, 1991).

For changes in hydraulic head that are less than a given preconsolidation head, an elastic response is computed. For changes in hydraulic head that are greater than a given preconsolidation head, an inelastic response is computed, and the resultant head becomes the new preconsolidation head. An initial preconsolidation head of 70 ft was used in the model, which means that if 70 ft of head decline occurs in a grid block, the model converts from an elastic to an inelastic storativity value. A preconsolidation head of 70 ft was used by Meyer and Carr (1979), Carr and others (1985), and Kasmarek and Strom (2002).

For the Chicot and Evangeline aquifers in the GAM area coincident with the model of Kasmarek and Strom (2002), the initial values of elastic and inelastic clay storativity (elastic and inelastic skeletal specific storage multiplied by cumulative clay thickness) are the calibrated values from Kasmarek and Strom (2002). For the rest of the area of the system, initial inelastic clay storativities were computed by multiplying areally distributed values of clay thickness from Gabrysch (1982) and Strom and others (2003a, b) (figs. 12, 18) by values of inelastic clay specific storage from Meyer and Carr (1979, p. 13) (8.7×10^{-5} ft^{-1} for Chicot aquifer and 1.5×10^{-5} ft^{-1} for Evangeline aquifer). Elastic clay storativity typically is about two orders of magnitude less than inelastic clay storativity (S.A. Leake, U.S. Geological Survey, oral commun., 1999). Initial elastic clay storativities for the Chicot and Evangeline aquifers thus were computed by multiplying inelastic storativities by 0.01.

Withdrawals

Simulations were made under transient conditions from 1891 through 2000 for 68 withdrawal (stress) periods of variable, but mostly annual, length (fig. 33, table 1). Monthly stress periods were applied for 3 years: 1980, 1982, and 1988. Substantially lower-than-average precipitation was recorded in the GAM area for those years. Monthly rather than annual stress periods would allow the model to represent ground-water withdrawals on a monthly or seasonal basis should the model be used to simulate hypothetical drought scenarios in the future. Total ground-water withdrawals increased from an estimated 41 Mgal/d in 1891 to about 1,130 Mgal/d in 1976, peaked at about 1,135 Mgal/d in 1980, and varied during the next 20 years but generally trended downward to about 850 Mgal/d in 2000.

Historical ground-water withdrawal (pumpage) data used in the model were compiled from numerous sources. Withdrawals were separated into seven categories on the basis of water use: municipal, manufacturing, mining, power generation, livestock, irrigation, and county-other. The sources and methods used to distribute withdrawals to the appropriate model layers and grid blocks are summarized in the appendix.

Model Calibration

The GAM was calibrated by trial-and-error adjustment of selected model input data (the aquifer properties that control

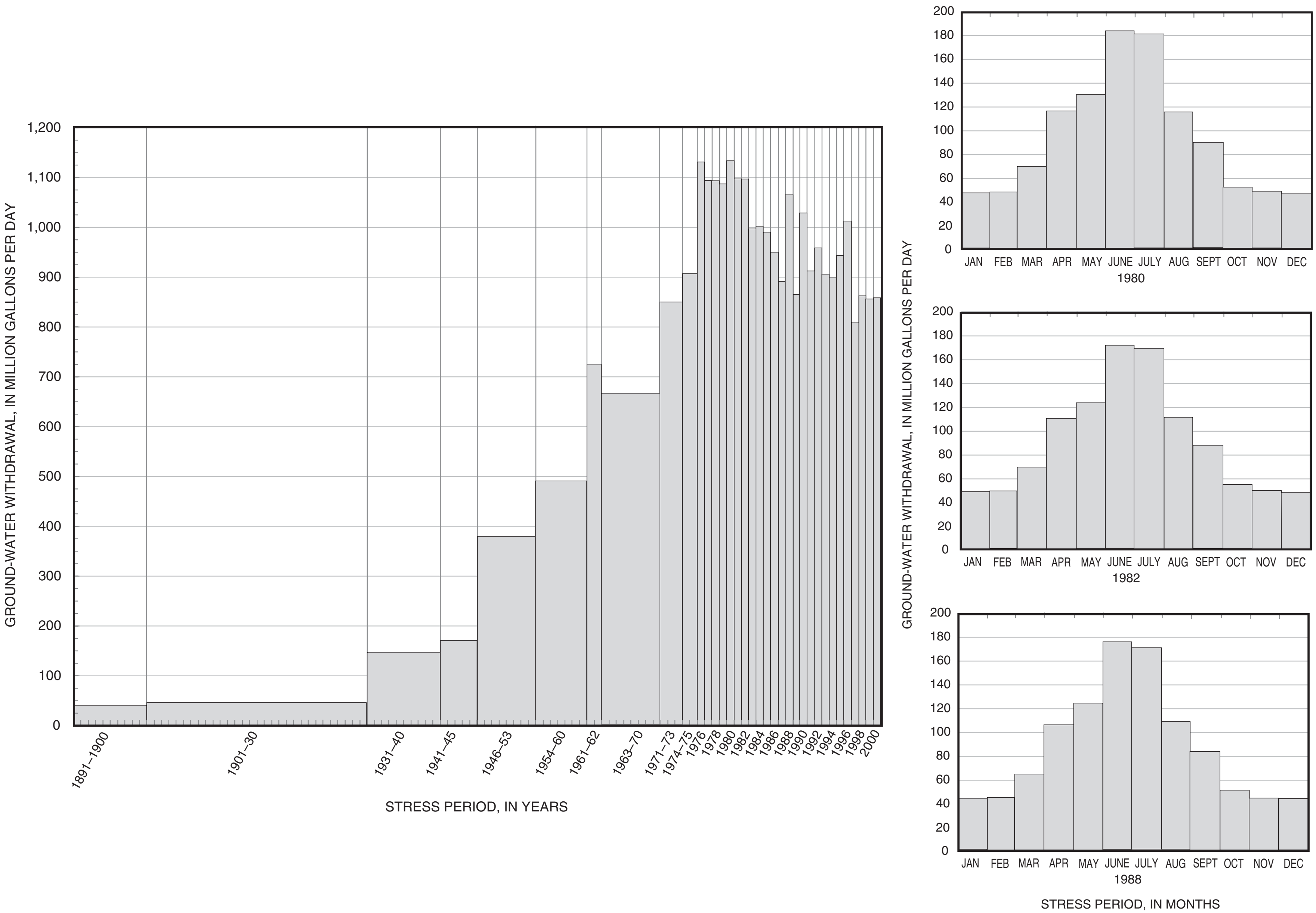


Figure 33. Total ground-water withdrawals used in transient Ground-Water Availability Model simulations by stress periods, 1891–2000.

Table 1. Ground-water withdrawal (stress) periods used in the Ground-Water Availability Model.

Stress period	Length of time (years)	Time interval	Stress period	Length of time (years)	Time interval	Stress period	Length of time (years)	Time interval
1	10	1891–1900	24	0.084	Oct. 1980	47	0.084	Mar. 1988
2	30	1901–30	25	.082	Nov. 1980	48	.082	Apr. 1988
3	10	1931–40	26	.084	Dec. 1980	49	.084	May 1988
4	5	1941–45	27	1	1981	50	.082	June 1988
5	8	1946–53	28	.084	Jan. 1982	51	.084	July 1988
6	7	1954–60	29	.076	Feb. 1982	52	.084	Aug. 1988
7	2	1961–62	30	.084	Mar. 1982	53	.082	Sept. 1988
8	8	1963–70	31	.082	Apr. 1982	54	.084	Oct. 1988
9	3	1971–73	32	.084	May 1982	55	.082	Nov. 1988
10	2	1974–75	33	.082	June 1982	56	.084	Dec. 1988
11	1	1976	34	.084	July 1982	57	1	1989
12	1	1977	35	.084	Aug. 1982	58	1	1990
13	1	1978	36	.082	Sept. 1982	59	1	1991
14	1	1979	37	.084	Oct. 1982	60	1	1992
15	.084	Jan. 1980	38	.082	Nov. 1982	61	1	1993
16	.076	Feb. 1980	39	.084	Dec. 1982	62	1	1994
17	.084	Mar. 1980	40	1	1983	63	1	1995
18	.082	Apr. 1980	41	1	1984	64	1	1996
19	.084	May 1980	42	1	1985	65	1	1997
20	.082	June 1980	43	1	1986	66	1	1998
21	.084	July 1980	44	1	1987	67	1	1999
22	.084	Aug. 1980	45	.084	Jan. 1988	68	1	2000
23	.082	Sept. 1980	46	.076	Feb. 1988			

water flow, recharge, discharge, and storage) in a series of transient simulations until the model output (potentiometric surfaces, land-surface subsidence, selected water-budget components) reasonably reproduced field measured (or estimated) aquifer responses. The calibration objective was to minimize the differences between simulated and measured aquifer responses.

Before calibration began, an initial predevelopment (no withdrawals) steady-state simulation was run to obtain starting heads for the hydrogeologic units for transient calibration simulations. Periodically during calibration, predevelopment steady-state simulations were run with the most current input data to obtain starting heads for successive transient calibration simulations. The input data that were adjusted from initial values on the basis of model output from successive transient simulations were transmissivity of the aquifers (actually, hydraulic conductivity, which is multiplied by aquifer thickness), storativity of sands, vertical hydraulic conductance between the water table and deeper zones of each hydrogeologic unit in outcrop areas, leakance between hydrogeologic units in subcrop areas, and inelastic clay storativity (actually, inelastic clay specific storage, which is multiplied by aquifer thickness) in the Chicot and Evangeline aquifers. Water-table heads, transmissivity and storativity of the Burkeville confining unit, storativity of the Jasper aquifer, and temporal and spatial distributions of withdrawals were not adjusted. Elastic specific storage of clays in the Chicot and Evangeline aquifers were computed by multiplying inelastic storativities by 0.01.

Model calibration comprised four elements: The first was qualitative comparison of simulated and measured potentiometric surfaces (or comparison of simulated potentiometric surfaces to measured point heads in areas of sparse measured heads) in the aquifers for 1977 and 2000; and quantitative comparison of simulated and measured potentiometric surfaces by computation and areal distribution of the root-mean-square (RMS) error (square root of the sum of the squares of the differences between simulated and measured heads divided by the total number of head measurements [240 for the three aquifers for 1977, 422 for the three aquifers for 2000]). In addition, graphical relations between simulated and measured heads for each aquifer for 1977 and 2000 were developed. The 1977 potentiometric surfaces were chosen for "history matching" because that was the first year of comprehensive, synoptic head measurements in the Chicot and Evangeline aquifers, at least in the Houston area (Gabrysch, 1979), following record-high withdrawals during the 1970s (fig. 33). The 2000 potentiometric surfaces were chosen because they represented the most recent year for which comprehensive, synoptic head measurements in the Chicot, Evangeline, and Jasper aquifers, again in the Houston area (Coplin and Santos, 2000; Coplin, 2001), were available, and withdrawals were substantially less in 2000 than in 1977. Also, cumulative long-term land-surface subsidence in the Houston area simulated as of 2000 could reasonably be compared to observed subsidence in 1995, the year of the most recent map of subsidence in the area. Final predevelopment potentiometric surfaces of the aquifers were simulated using calibrated distributions of input data for comparison with conceptualized configurations of the actual predevelopment surfaces.

The second calibration element was comparison of simulated and measured hydrographs from wells in the aquifers primarily in the Houston area, the coastal irrigation area, and selected counties away from those areas of withdrawal. Hydrographs for comparison were selected on the basis of adequate record length (at least 10 years) and period of record (primarily 1977–2000).

The third calibration element was comparison of simulated water-budget components—primarily recharge and discharge—to estimates of physically reasonable ranges of actual water-budget components. Comparisons of simulated distributions of recharge and discharge in the outcrops of aquifers to estimates of physically reasonable distributions on the basis of knowledge of the hydrology of the aquifer system also were used.

The fourth calibration element was comparison of simulated land-surface subsidence from predevelopment to 2000 to measured land-surface subsidence from 1906 through 1995 (Harris-Galveston Coastal Subsidence District, 1998) in the Chicot and Evangeline aquifers in Harris, Galveston, and Fort Bend Counties (the counties in which historical subsidence has been monitored). The amount of subsidence that occurred between 1995 and 2000, as indicated by measured compaction at 11 borehole extensometer sites in the Houston area (Coplin and others, 2001, fig. 8), was about 0.1 ft or less at 10 of the 11 sites and about 0.6 ft at one site (in west-central Harris County); thus the comparison of simulated to measured subsidence was judged a "like comparison," despite the 5-year time difference between the two datasets.

Model Results

Simulated Hydraulic Properties Associated With Ground-Water Flow and Subsidence

The calibrated areal distributions of simulated transmissivity in the hydrogeologic units are shown in figures 34–37. For the Chicot aquifer, transmissivities range from negligible to about 77,000 ft^2/d (fig. 34). For the Evangeline aquifer, transmissivities range from negligible to about 43,000 ft^2/d (fig. 35). Transmissivities near the maximums for both aquifers occur in only a few grid blocks. Transmissivities of the Burkeville confining unit (unadjusted from initial values during calibration) are very small (maximum about 8 ft^2/d) and shown here only for completeness (fig. 36). For the Jasper aquifer, transmissivities range from negligible to about 14,500 ft^2/d (fig. 37).

Transmissivities of the aquifers are of the same orders of magnitude as those reported for the respective aquifers in previous studies (Wesselman, 1972; Jorgensen, 1975; Carr and others, 1985; Baker, 1986; Kasmarek and Strom, 2002; Ryder and Ardis, 2002). However, the distributions of transmissivity show that, generally, the largest values are coincident with areas of large withdrawals. The coincidence of large transmissivity and large withdrawals probably is an artifice of the model (see "Model Limitations/Input Data" section). Numerous trial-and-error adjustments of the input properties could not simultaneously maintain adequate history matching of heads and subsidence (in the Houston area) and eliminate large transmissivities coincident with large withdrawals. The implication for the simulated system, assuming withdrawals are estimated accurately, is that larger-than-actual amounts of water moving to centers of withdrawal from adjacent areas are compensating for smaller-than-actual amounts of water from storage or induced recharge, or both.

Storativities of the Chicot and Evangeline aquifers (1×10^{-4} to 0.2 and 4×10^{-5} to 0.2, respectively, figs. 38, 39) reflect aquifer conditions from confined to semiconfined to water table. Chicot aquifer storativities (fig. 38) generally are largest in the updip, outcrop areas where water-table conditions prevail. Also notable is the area of water-table storativities in the coastal irrigation area centered in Wharton County. A relatively large fraction of the withdrawals there is supplied by storage in the shallow zones of the Chicot aquifer. Storativities of the Burkeville confining unit and the Jasper aquifer (1×10^{-5} to 5×10^{-2} and 2×10^{-5} to 0.2, respectively) also are generally largest in the updip, outcrop areas where water-table conditions prevail (figs. 40, 41). As with the distributions of transmissivity, the distributions of Chicot and Evangeline aquifer storativity (aside from water-table storativities in updip, outcrop areas) are somewhat artifices of the model (See "Model Limitations/Input Data" section).

The calibrated distribution of vertical hydraulic conductance between the water table and deeper zones of all hydrogeologic units in outcrop areas is shown in figure 42. Hydraulic conductances range from negligible to nearly 70,000 ft^2/d, the maximum only in a few small areas. Largest values generally are near the updip limits of the Chicot and Evangeline aquifers.

The calibrated distributions of leakance between hydrogeologic units in subcrop areas are shown in figures 43–45. The largest leakances generally occur near the updip limits of the overlying units for the Chicot-Evangeline distribution (fig. 43) and the Burkeville-Jasper distribution (fig. 45), which is consistent with the fact that the units contain relatively more sand than clay in updip areas. An exception is the Evangeline-Burkeville distribution (fig. 44), which could reflect a relatively less sandy composition of the Burkeville beneath the Evangeline outcrop. Another exception is the relatively large Chicot-Evangeline aquifer leakance coincident with areas of large withdrawals and consequent large cones of depression. As with coincident large transmissivity and withdrawals/cones of depression, the similar configuration in Chicot-Evangeline aquifer leakance could be an artifice of the model.

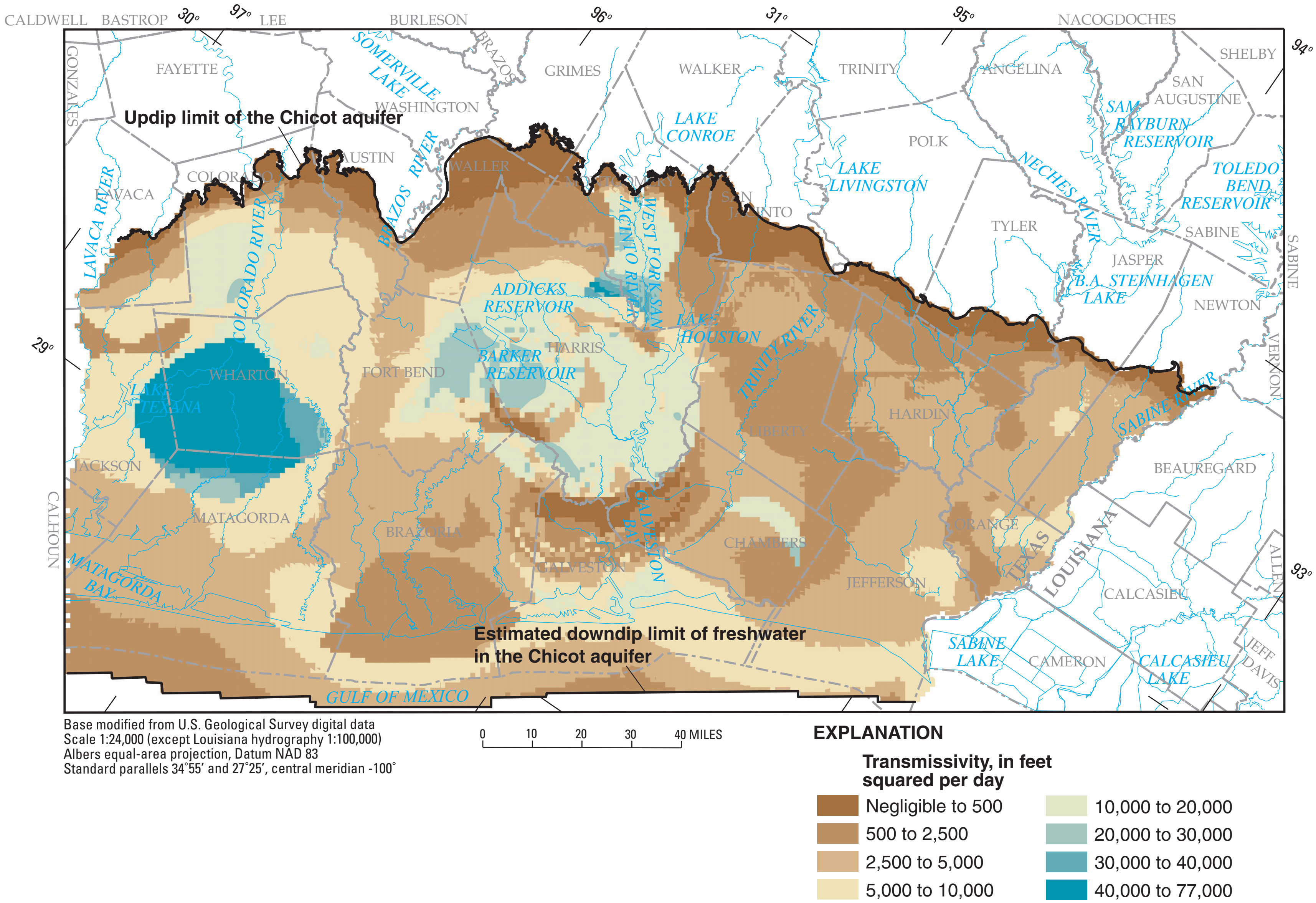


Figure 34. Simulated transmissivity of the Chicot aquifer in the Ground-Water Availability Model area.

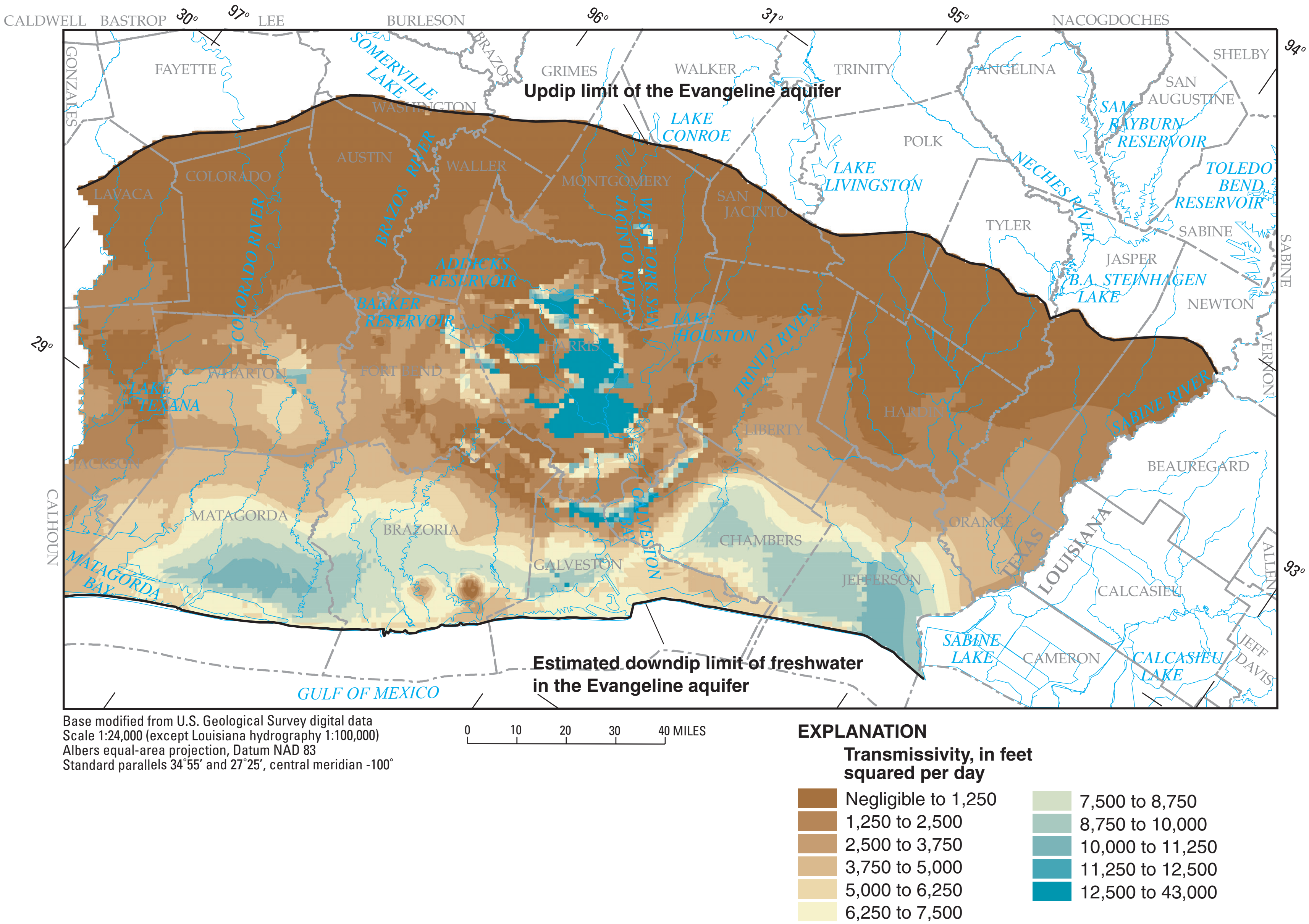


Figure 35. Simulated transmissivity of the Evangeline aquifer in the Ground-Water Availability Model area.

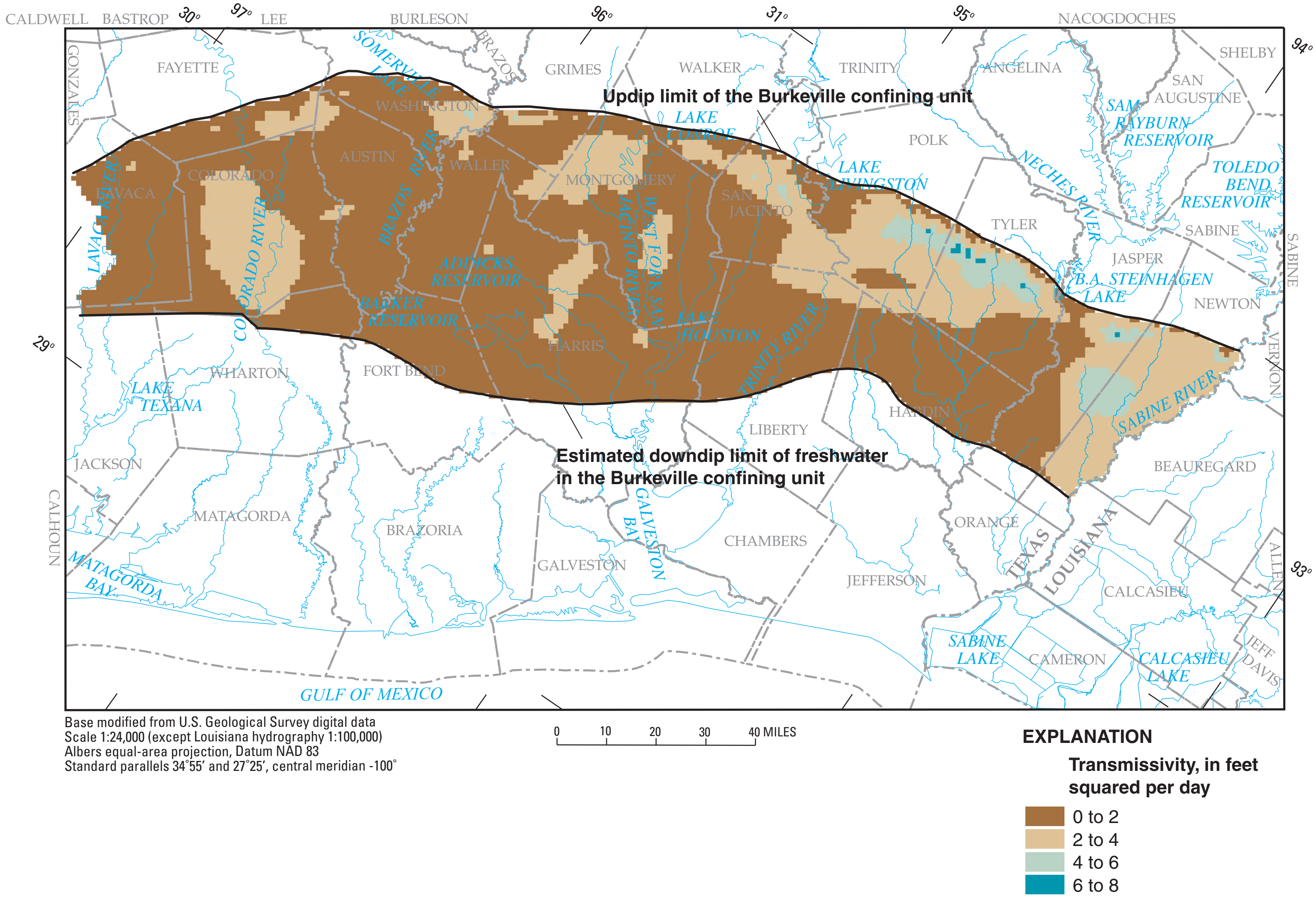
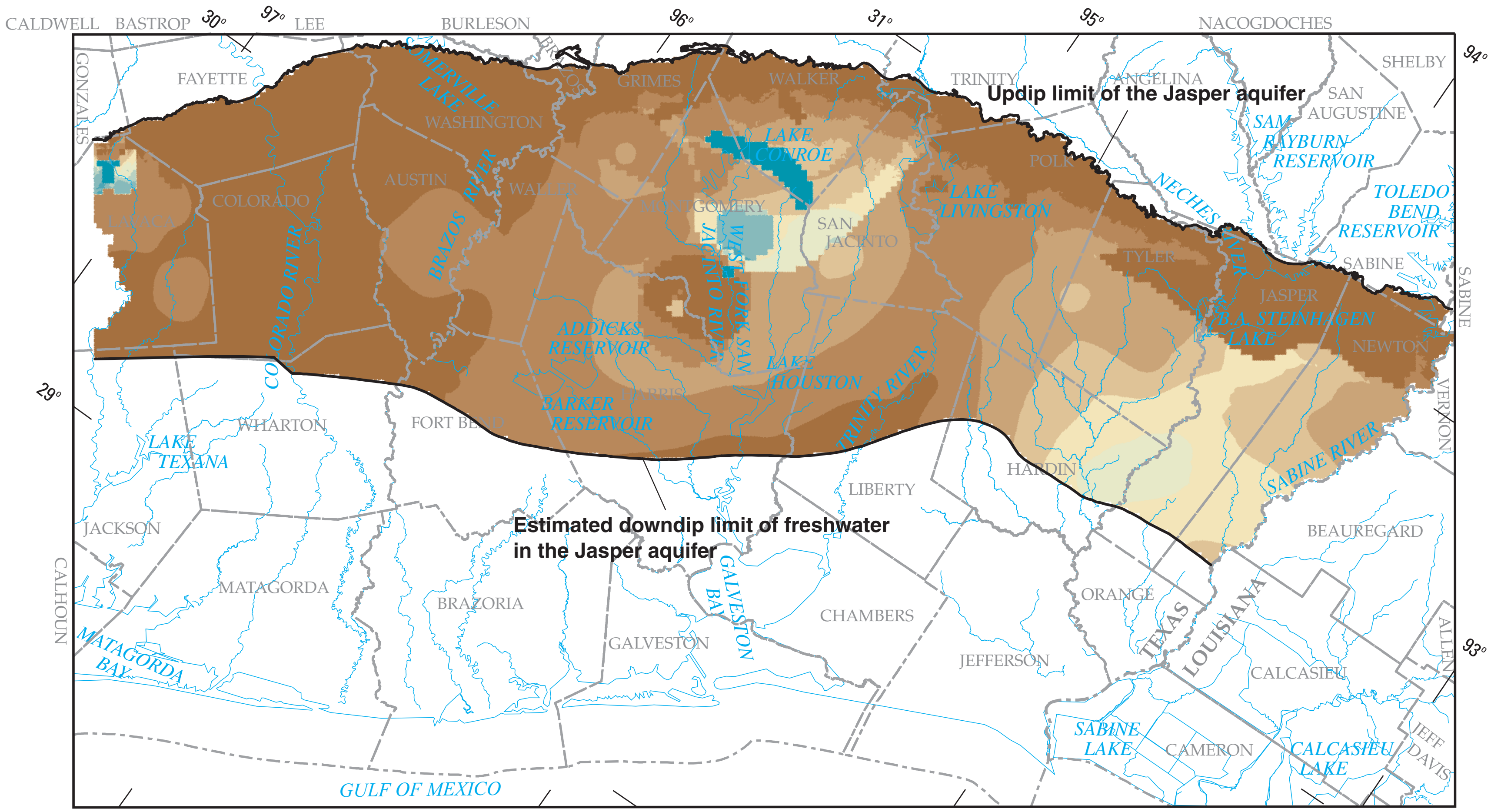


Figure 36. Simulated transmissivity of the Burkeville confining unit in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

EXPLANATION

Transmissivity, in feet squared per day

- Negligible to 250
- 250 to 500
- 500 to 750
- 750 to 1,000
- 1,000 to 1,250
- 1,250 to 1,500
- 1,500 to 1,750
- 1,750 to 2,500
- 2,500 to 5,000
- 5,000 to 14,500

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 37. Simulated transmissivity of the Jasper aquifer in the Ground-Water Availability Model area.

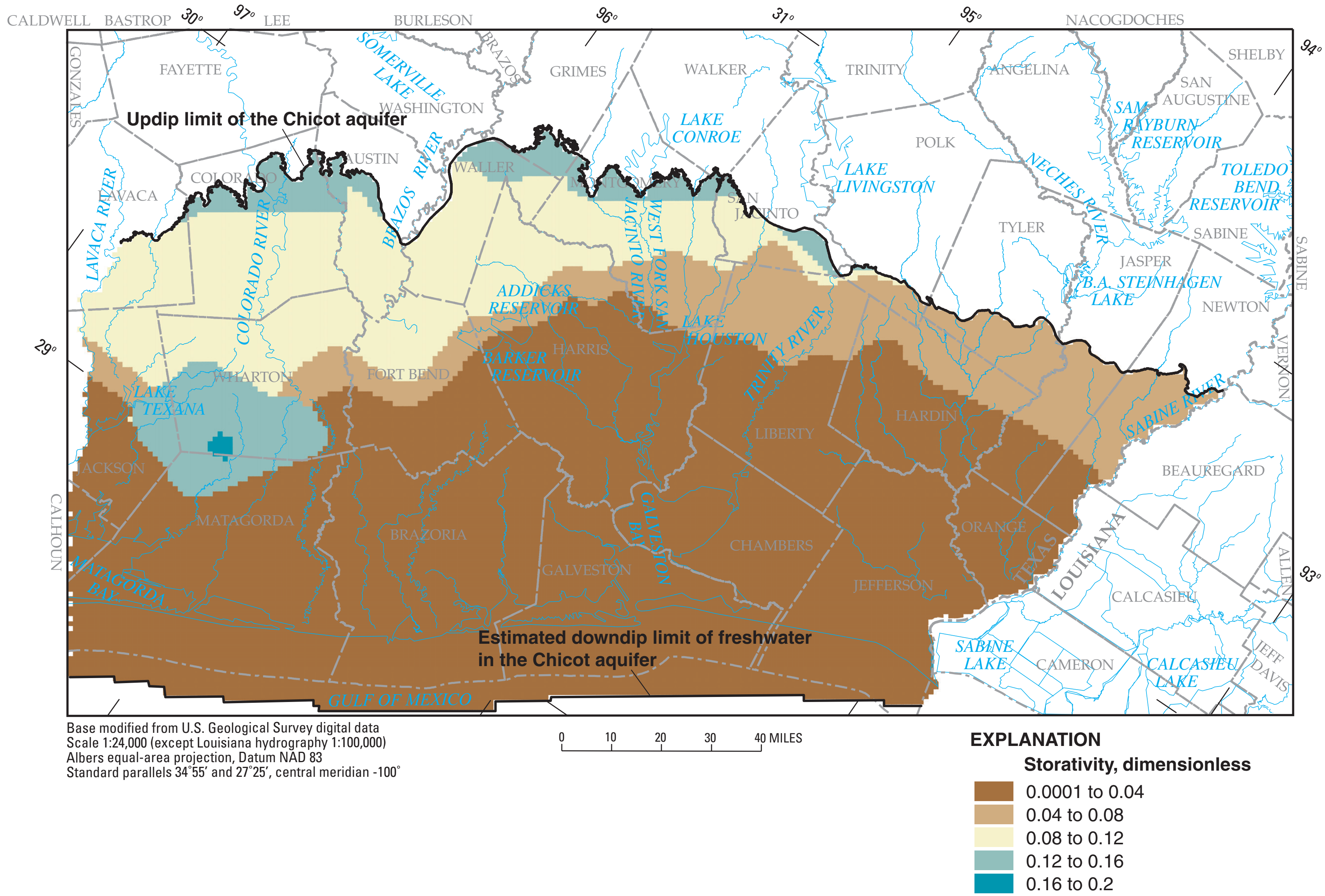
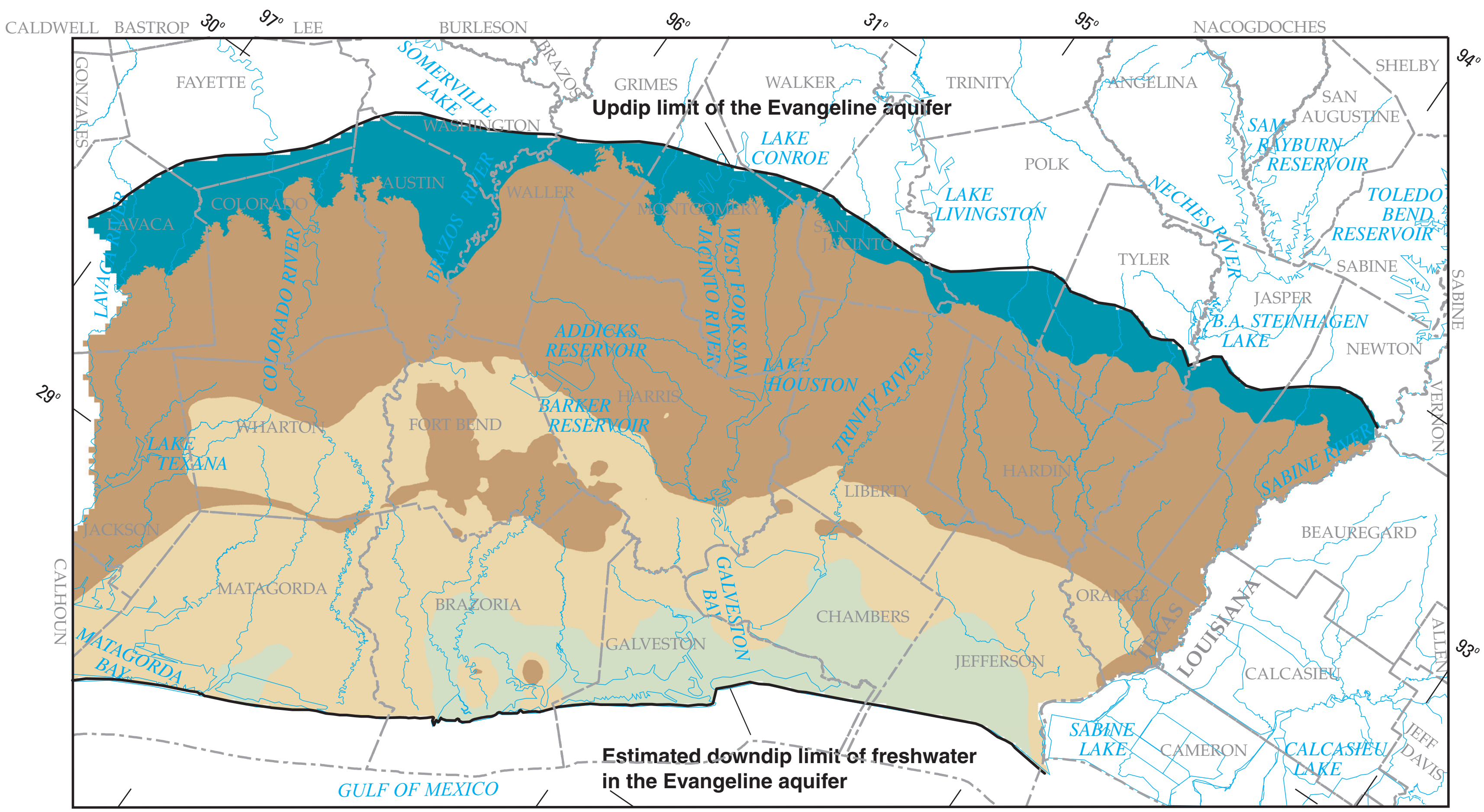


Figure 38. Simulated storativity of the Chicot aquifer in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

Storativity, dimensionless

- 4×10^{-5}
- 4×10^{-5} to 8×10^{-4}
- 8×10^{-4} to 2×10^{-3}
- 2×10^{-3} to 3×10^{-3}
- 3×10^{-3} to 0.2
- 0.2

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 39. Simulated storativity of the Evangeline aquifer in the Ground-Water Availability Model area.

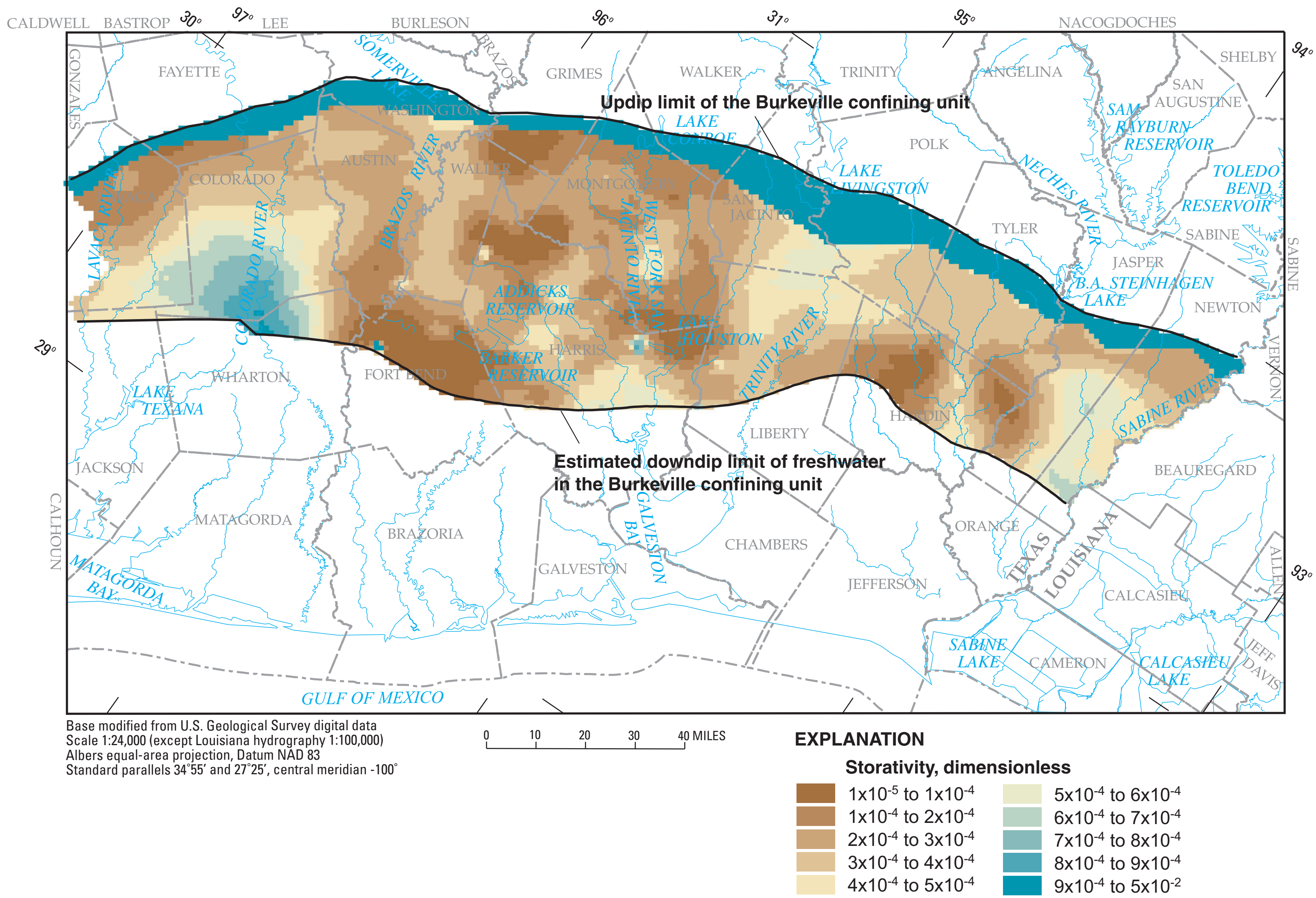


Figure 40. Simulated storativity of the Burkeville confining unit in the Ground-Water Availability Model area.

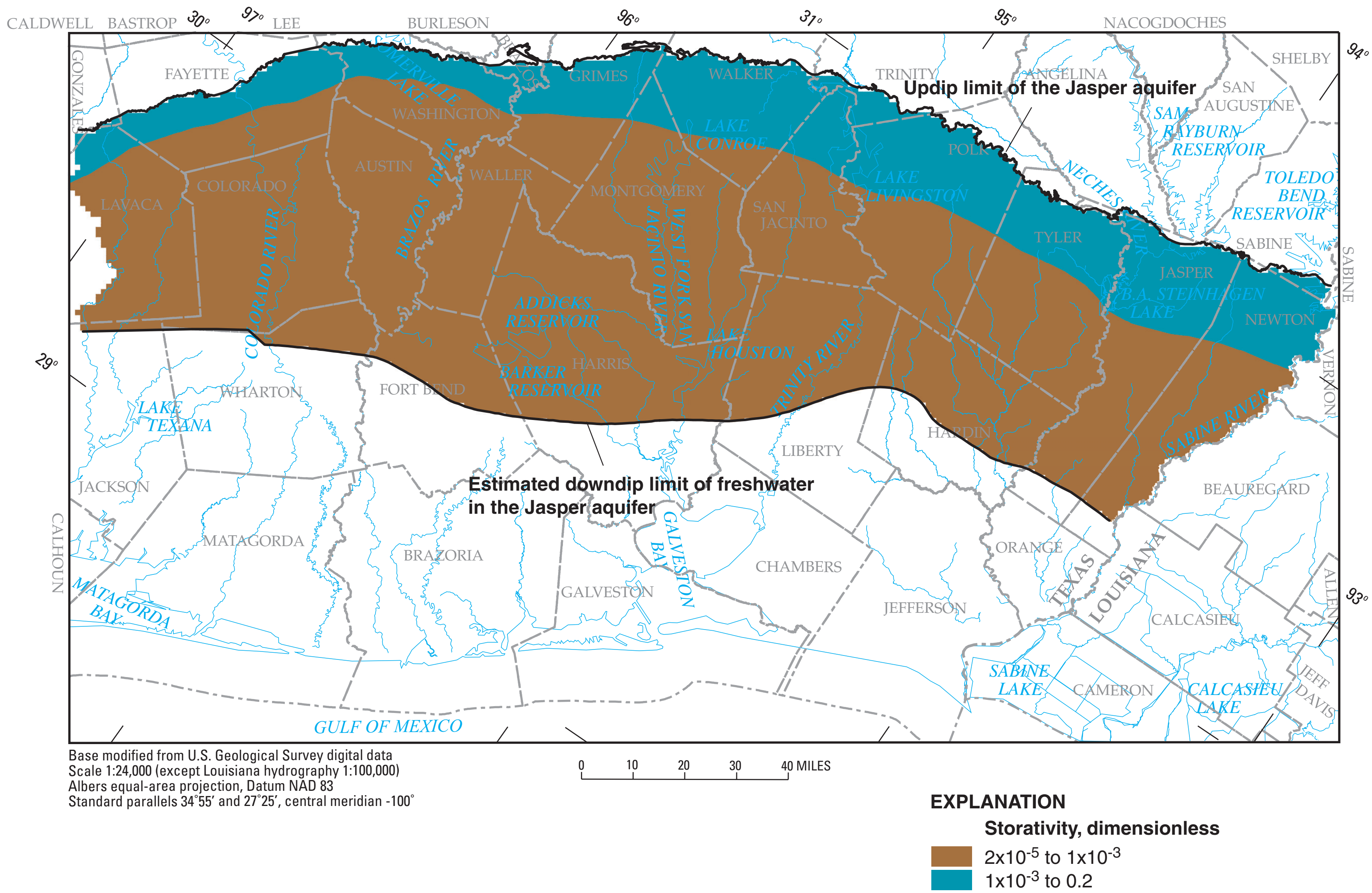


Figure 41. Simulated storativity of the Jasper aquifer in the Ground-Water Availability Model area (modified from Wesselman, 1967; and Strom and others, 2003c).

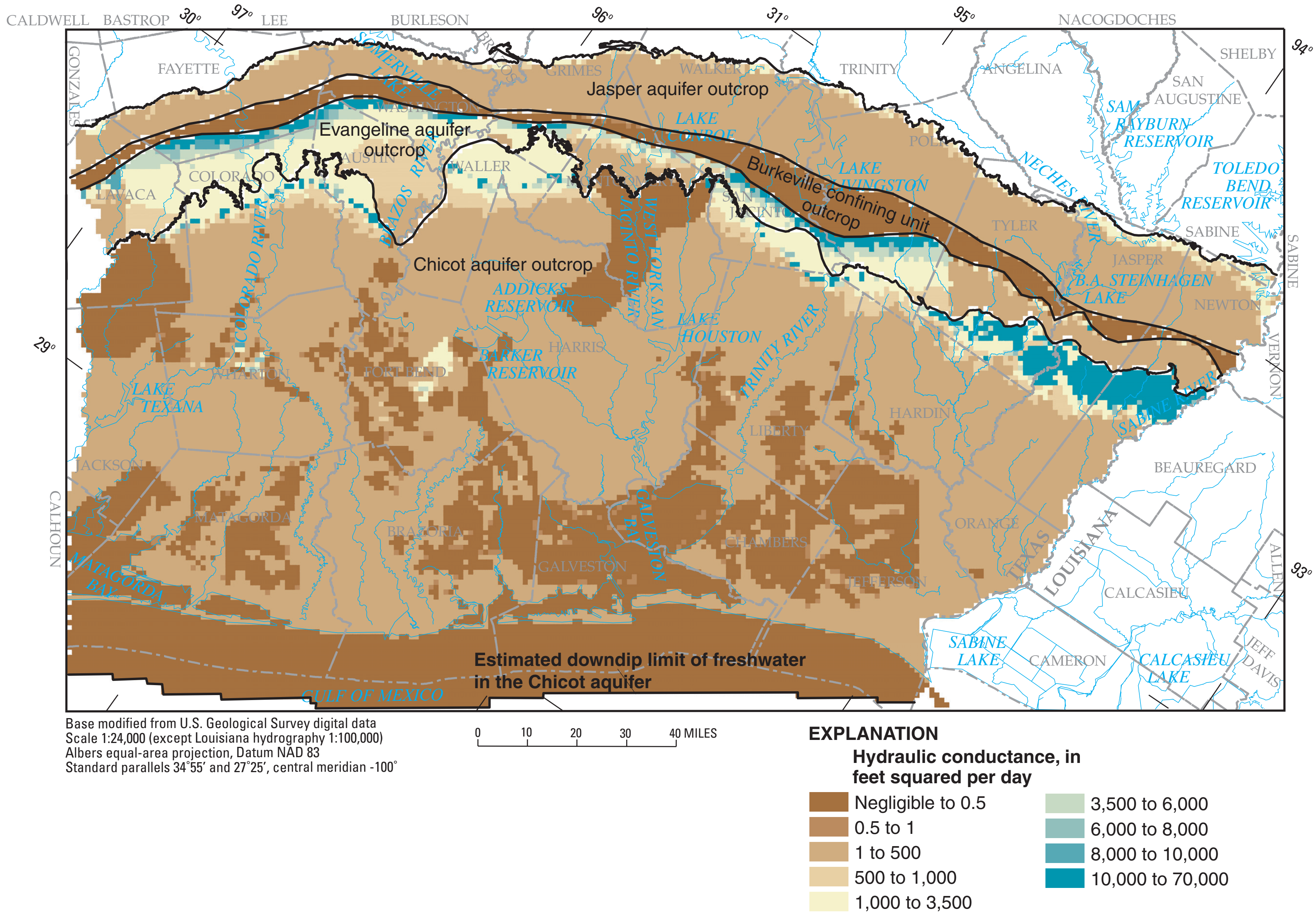


Figure 42. Simulated vertical hydraulic conductance between water table and deeper zones of the hydrogeologic units in the Ground-Water Availability Model area.

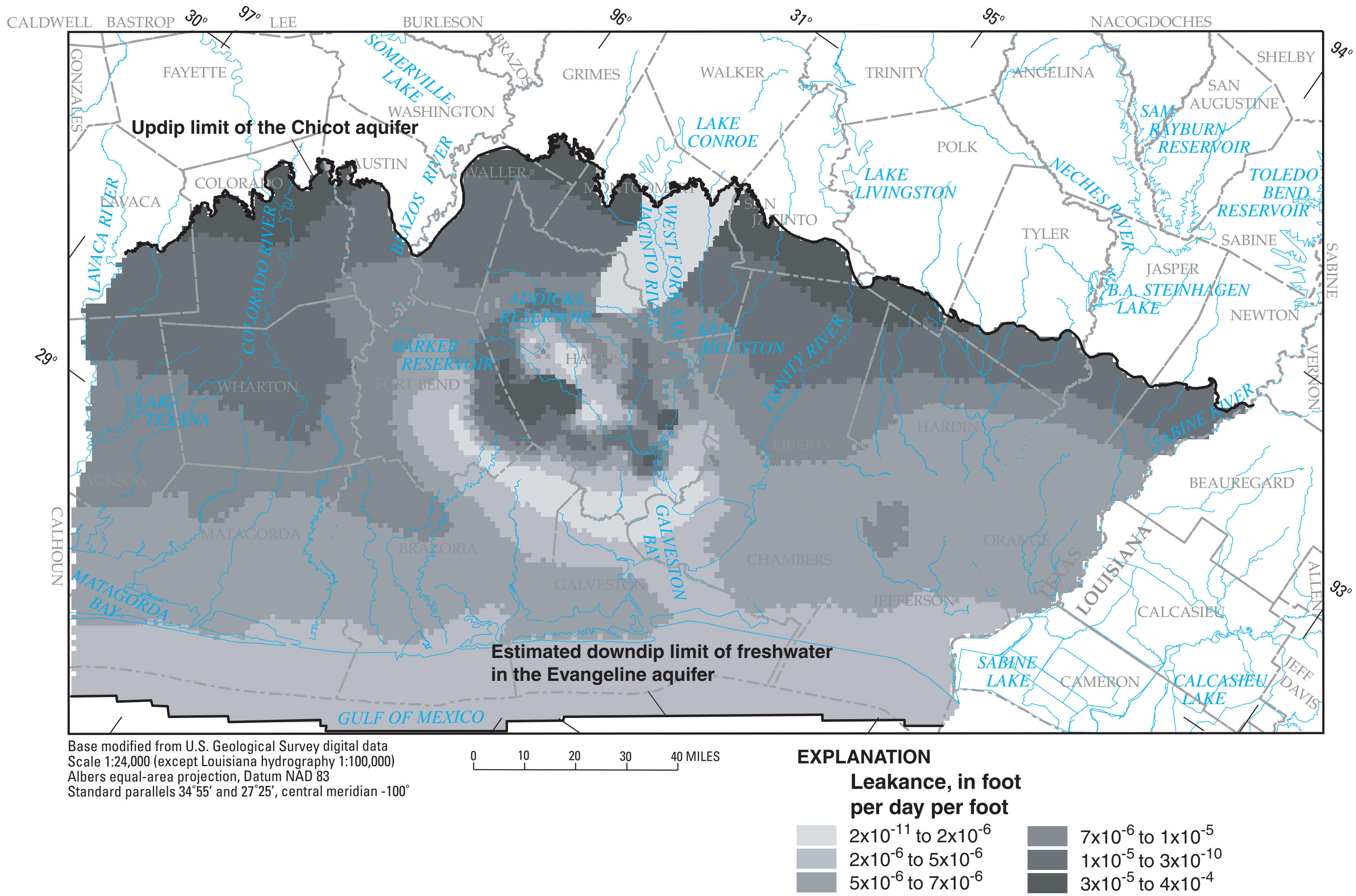
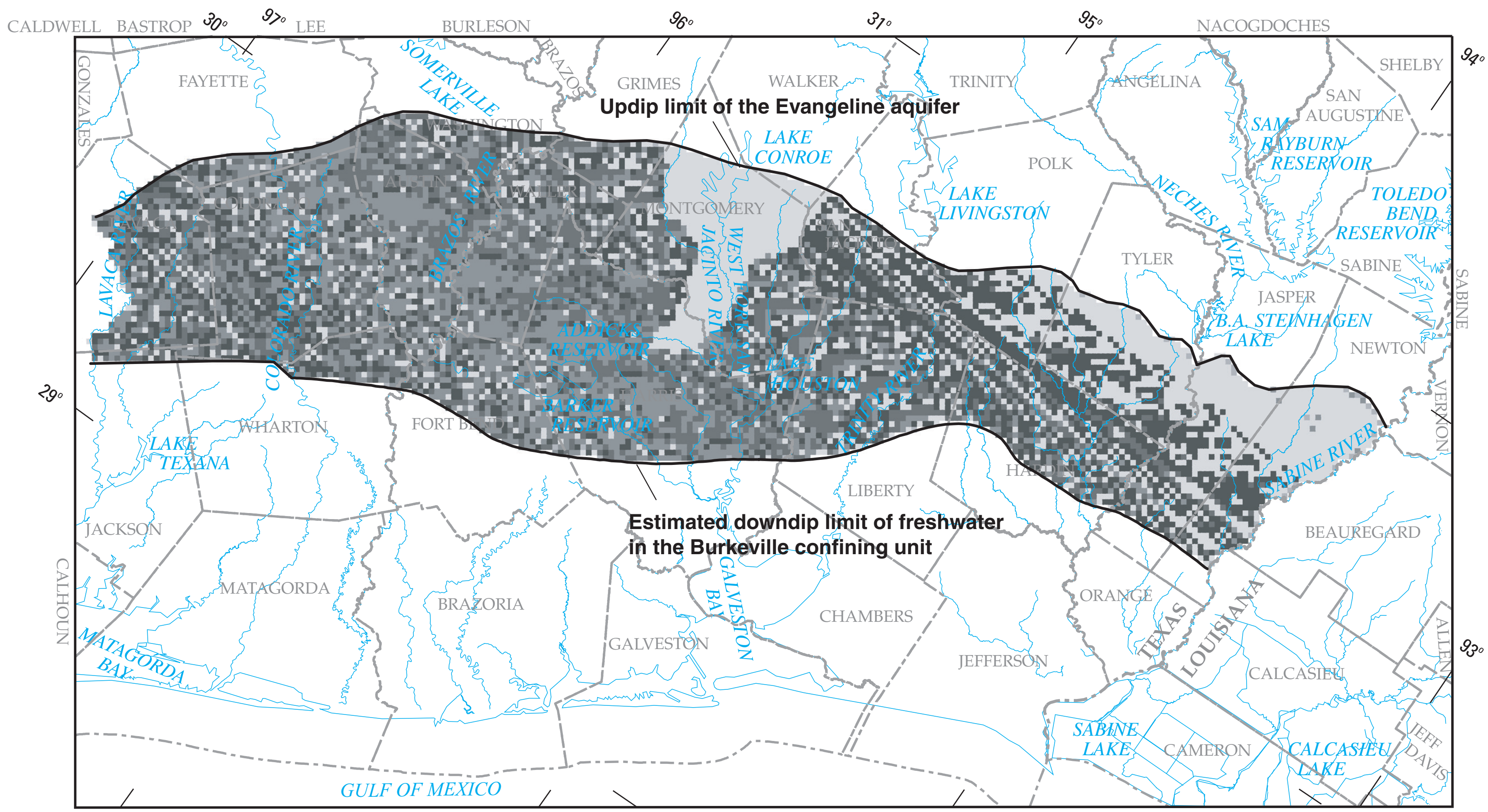


Figure 43. Simulated leakance between the Chicot and Evangeline aquifers in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

EXPLANATION

Leakance, in foot per day per foot

- 5×10^{-11} to 3.7×10^{-6}
- 3.7×10^{-6} to 3.8×10^{-6}
- 3.8×10^{-6} to 7×10^{-6}
- 7×10^{-6} to 1.5×10^{-5}
- 1.5×10^{-5} to 5×10^{-3}

Figure 44. Simulated leakance between the Evangeline aquifer and the Burkeville confining unit in the Ground-Water Availability Model area.

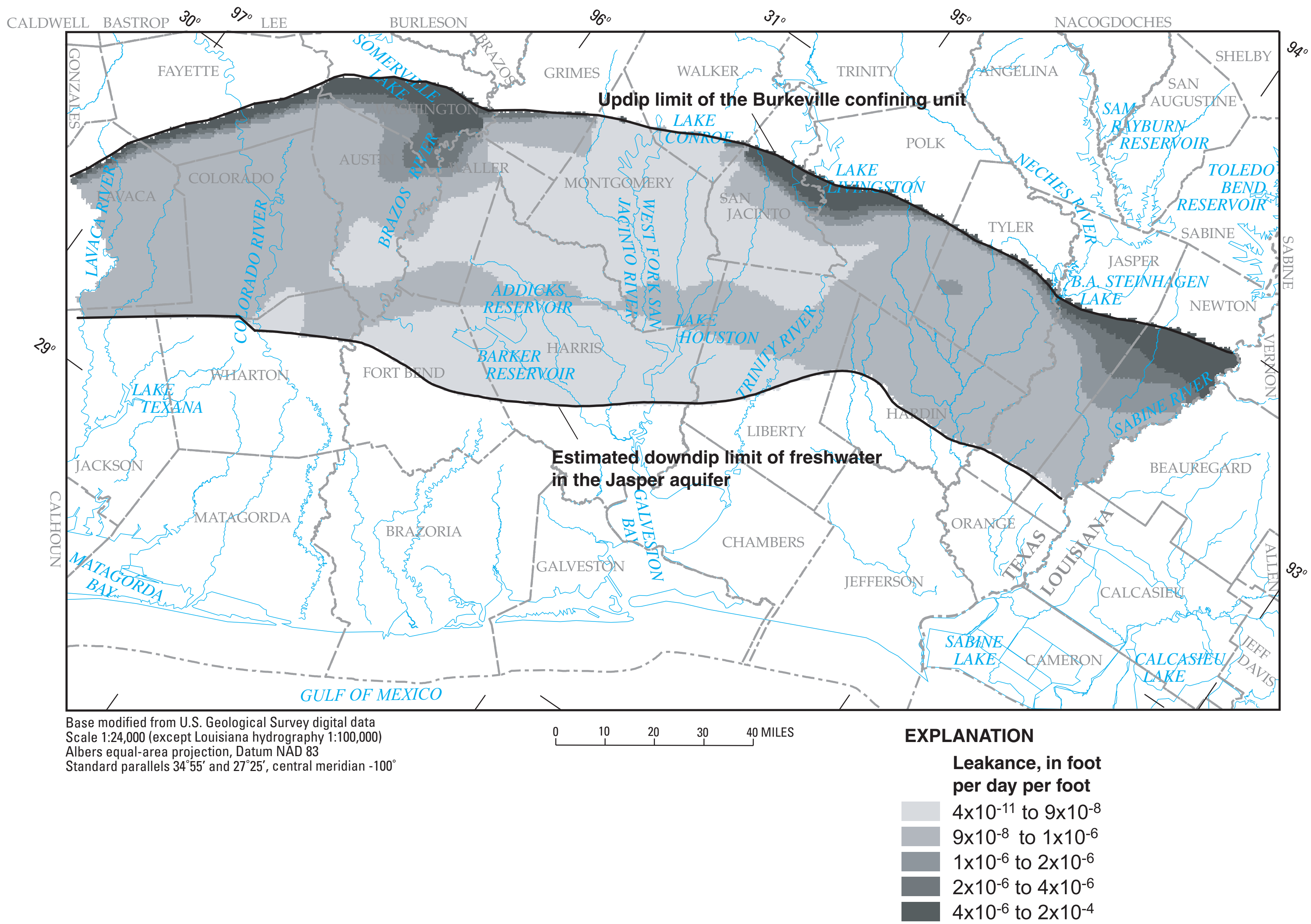


Figure 45. Simulated leakance between the Burkeville confining unit and the Jasper aquifer in the Ground-Water Availability Model area.

The calibrated distributions of inelastic clay storativity for the Chicot and Evangeline aquifers are shown in figures 46 and 47. Because the history matching of simulated and measured subsidence was limited to the Harris-Galveston-Fort Bend County area, that area is the only part of the GAM area for which inelastic clay storativity can be considered calibrated. Inelastic clay storativities range from negligible to about 0.1 for both aquifers.

Simulated and Measured Potentiometric Surfaces, 1977, 2000; and Simulated Predevelopment Surface

The simulated potentiometric surfaces of the Chicot, Evangeline, and Jasper aquifers for 1977 show general agreement with the measured potentiometric surfaces (or with measured point head data in areas where data are sparse) (figs. 48–50). The simulated and measured cones of depression centered in Harris County in the Chicot and Evangeline aquifers caused by major withdrawals are essentially coincident, although the maximum depths (relative to NGVD 29) of the simulated cones are less than those of the measured cones (simulated Chicot, -150 ft, measured Chicot, -250 ft; simulated Evangeline, -250 ft, measured Evangeline, -350 ft). The only area away from the area of major withdrawals in which simulated cones of depression appear in the Chicot and Evangeline aquifers is the Evadale-Beaumont area on the Hardin-Jasper County line. Although no measured heads were available for comparison, the location of the cones is consistent with the center of withdrawals in the area. The simulated 1977 potentiometric surface for the Jasper aquifer shows a configuration generally similar to those of the other two aquifers but without cones of depression associated with withdrawals in the Houston area.

The RMS errors for the aquifer potentiometric surfaces, which reflect the average difference between 1977 simulated and measured heads, were about 34 ft for the Chicot aquifer, about 43 ft for the Evangeline aquifer, and about 47 ft for the Jasper aquifer (table 2). The RMS errors are about 7, 8, and 17 percent, respectively, of the total range in measured heads for the respective aquifers.

Table 2. Number of water-level (head) measurements and root-mean-square errors of simulated water levels in the Chicot, Evangeline, and Jasper aquifers, 1977 and 2000.

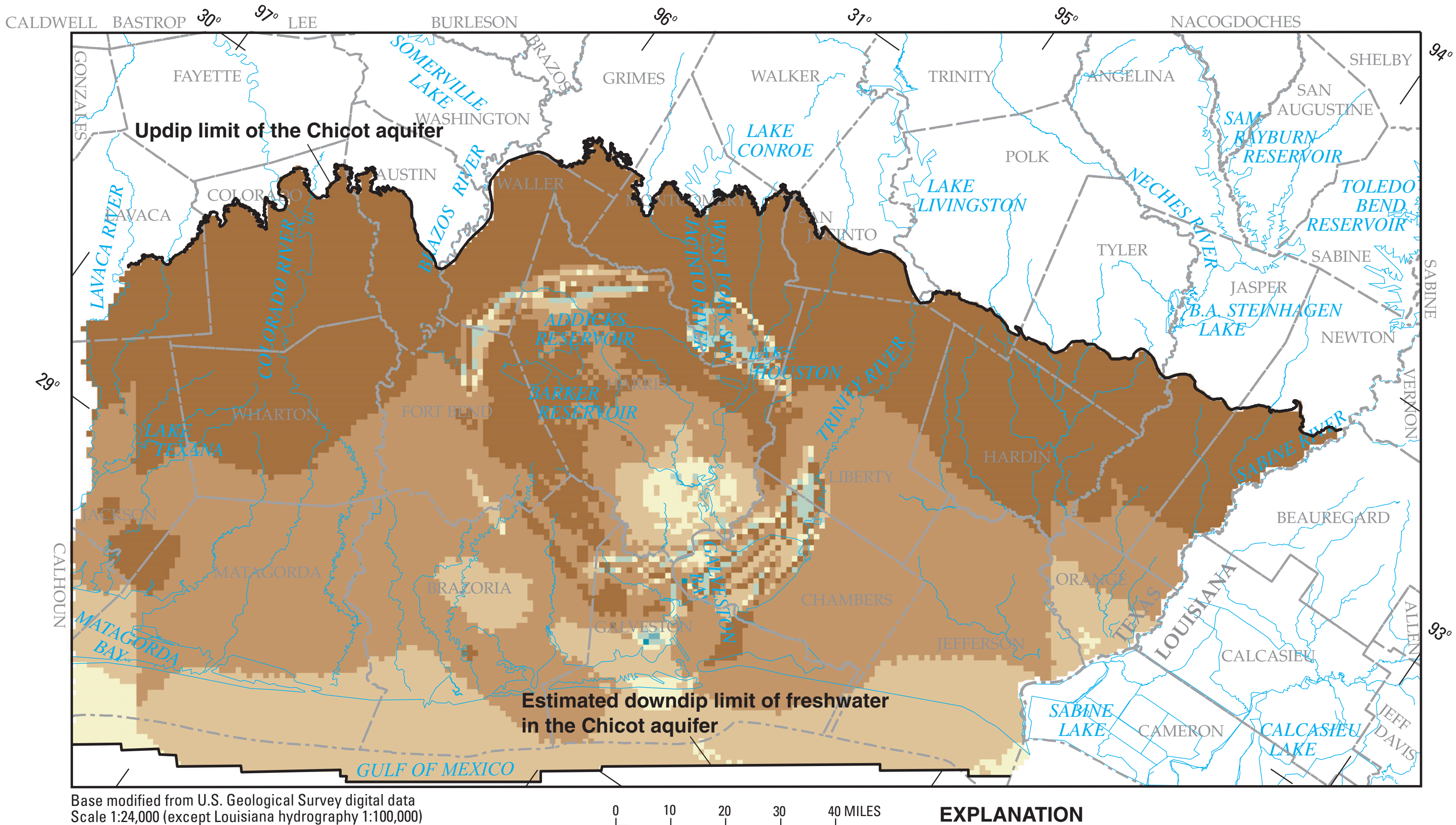
	Aquifer	Number of water-level measurements	Root-mean-square error of simulated water levels (feet)
1977			
	Chicot	204	34.0
	Evangeline	169	43.1
	Jasper	33	47.4
2000			
	Chicot	200	30.7
	Evangeline	153	40.1
	Jasper	69	33.8

Graphical comparison of simulated and measured 1977 heads for the Chicot and Evangeline aquifers (fig. 51) generally shows little bias toward simulated heads greater than or less than measured heads throughout the middle and upper ranges of head values for those aquifers; but toward the lower end of the ranges of head values for those aquifers, simulated heads tend to be somewhat greater than measured heads. For the Jasper aquifer (fig. 51) more simulated heads are greater than measured heads throughout the entire range of head values.

The maps showing distributions of 1977 head residuals (difference between measured and simulated heads, computed as measured minus simulated) (figs. 52–54) show where in the GAM area simulated heads tend to be greater than or less than measured heads and by how much. For the Chicot aquifer as previously described, simulated heads are greater than measured heads by the largest amount in the area of major withdrawals centered in Harris County. Simulated heads generally are larger than measured heads in the southwestern part of the GAM area, particularly in parts of another area of major withdrawals, the coastal irrigation area centered in Wharton County. The same pattern generally characterizes residuals in the Evangeline aquifer, although simulated Evangeline aquifer heads tend to be greater than measured heads in more of the northeastern GAM area than is the case for simulated Chicot aquifer heads. Simulated Jasper aquifer heads tend to be greater than measured heads over most of the GAM area, which is consistent with the graph of figure 51.

The simulated potentiometric surfaces of the Chicot and Evangeline aquifers for 2000 also show general agreement with the measured potentiometric surfaces (or with measured point head data in areas where data are sparse) (figs. 55, 56); and for 2000, sufficient measured head data for a Jasper aquifer potentiometric surface were available, at least in the Montgomery County area. Jasper aquifer simulated and measured surfaces also are reasonably close (fig. 57). The simulated and measured 2000 Chicot and Evangeline potentiometric surfaces, compared with those for 1977, show substantial shifts to the northwest in the major cones of depression, which reflect shifts northwestward of the centers of withdrawals during 1977–2000. The measured 2000 Chicot aquifer potentiometric surface also shows about 100 ft of recovery in the major cone of depression, which is consistent with the overall reduction in withdrawals from the system during 1977–2000. However, the measured 2000 cone of depression in the Evangeline aquifer actually deepened by about 50 ft.

Unlike the simulated major cones of depression in the 1977 Chicot and Evangeline aquifer potentiometric surfaces, those simulated cones of depression for 2000 are not less than the maximum depths of the measured cones. After a series of calibration simulations for both years with various adjustments to input data, and discussion with HGCSD (Tom Michel, Harris-Galveston Coastal Subsidence District, oral commun., 2004), the authors believe the simulated and measured cones of depression for 2000 match more closely than for 1977 because



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

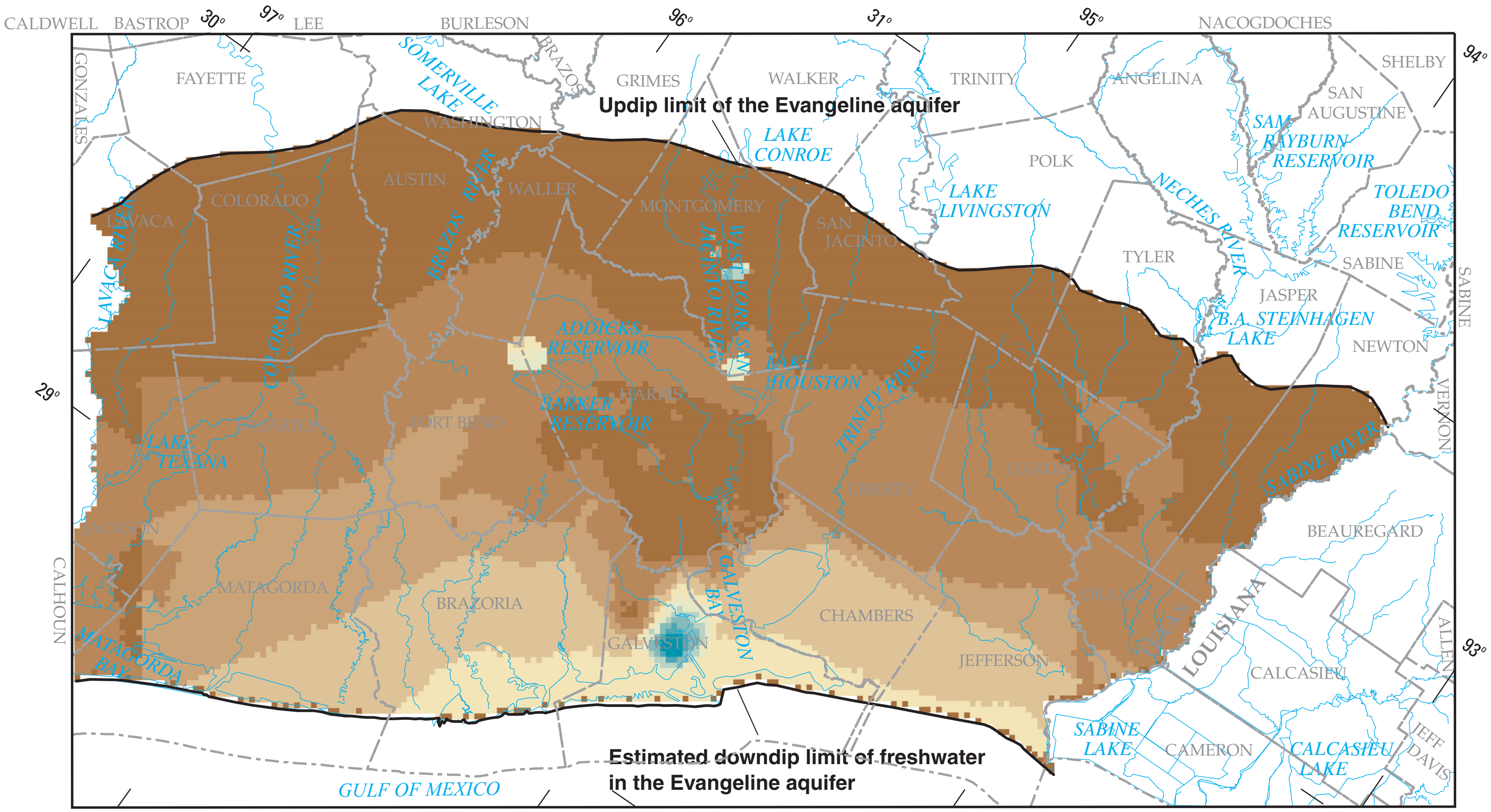
EXPLANATION

Inelastic storativity, dimensionless

- Negligible to 0.02
- 0.02 to 0.04
- 0.04 to 0.06
- 0.06 to 0.08
- 0.08 to 0.1
- 0.1 to 0.12
- 0.12 to 0.14

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 46. Simulated inelastic clay storativity of the Chicot aquifer in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

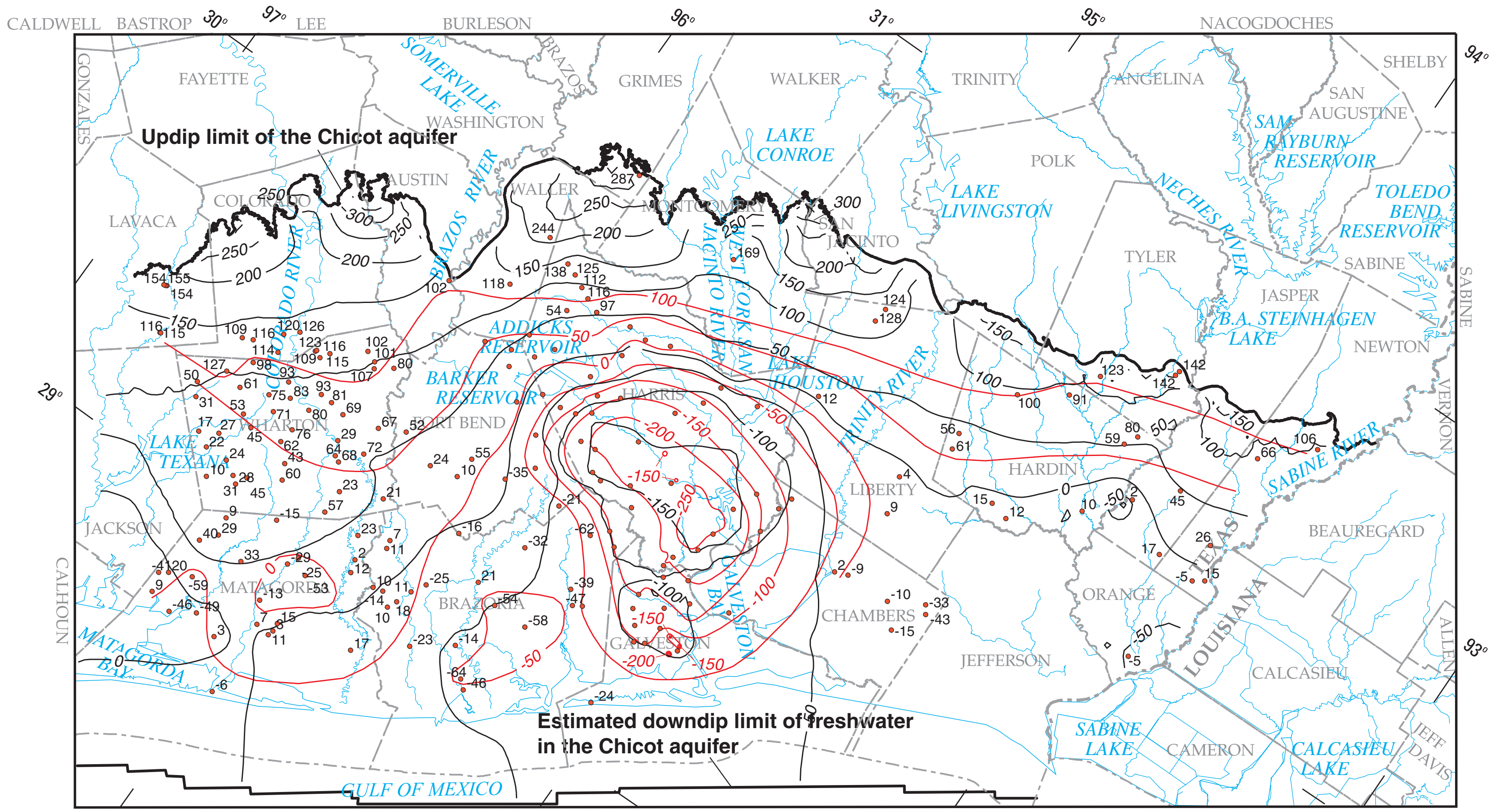
EXPLANATION

Inelastic storativity, dimensionless

- Negligible to 0.01
- 0.01 to 0.02
- 0.02 to 0.03
- 0.03 to 0.04
- 0.04 to 0.05
- 0.05 to 0.06
- 0.06 to 0.07
- 0.07 to 0.08
- 0.08 to 0.09
- 0.09 to 0.1

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 47. Simulated inelastic clay storativity of the Evangeline aquifer in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

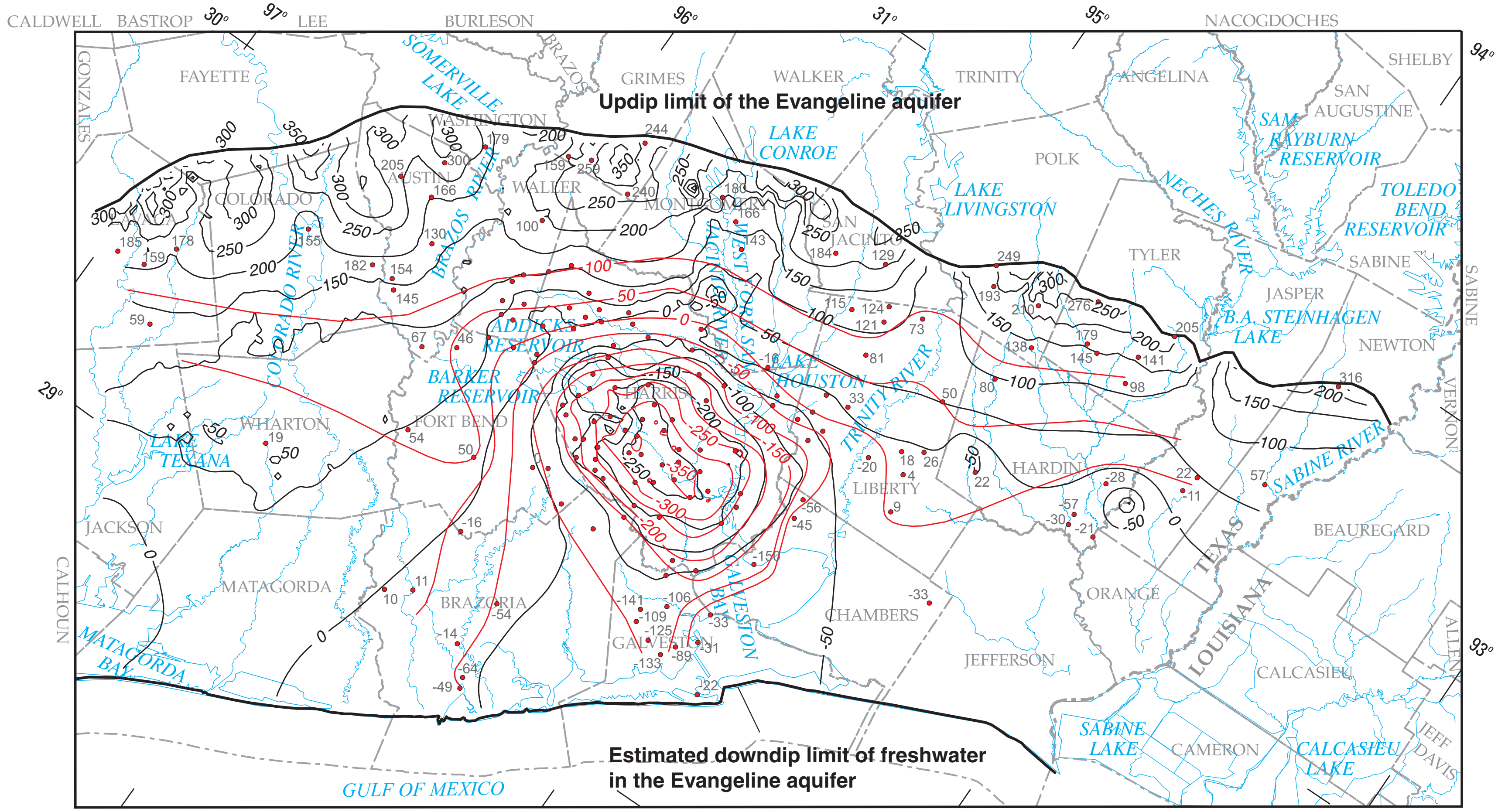
EXPLANATION

-150 **Simulated potentiometric contour—**Shows altitude at which water would have stood in tightly cased well. Interval 50 feet. Datum is NGVD 29

-150 **Measured potentiometric contour—**Shows altitude at which water would have stood in tightly cased well. Interval 50 feet. Datum is NGVD 29

-29 **Data point—**Well in which water-level measurement was made. Number is water-level altitude (shown in areas not having published contours)

Figure 48. Simulated and measured 1977 potentiometric surfaces of the Chicot aquifer and 1977 water-level measurements from wells screened in the Chicot aquifer (modified from Gabrysch, 1979) in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

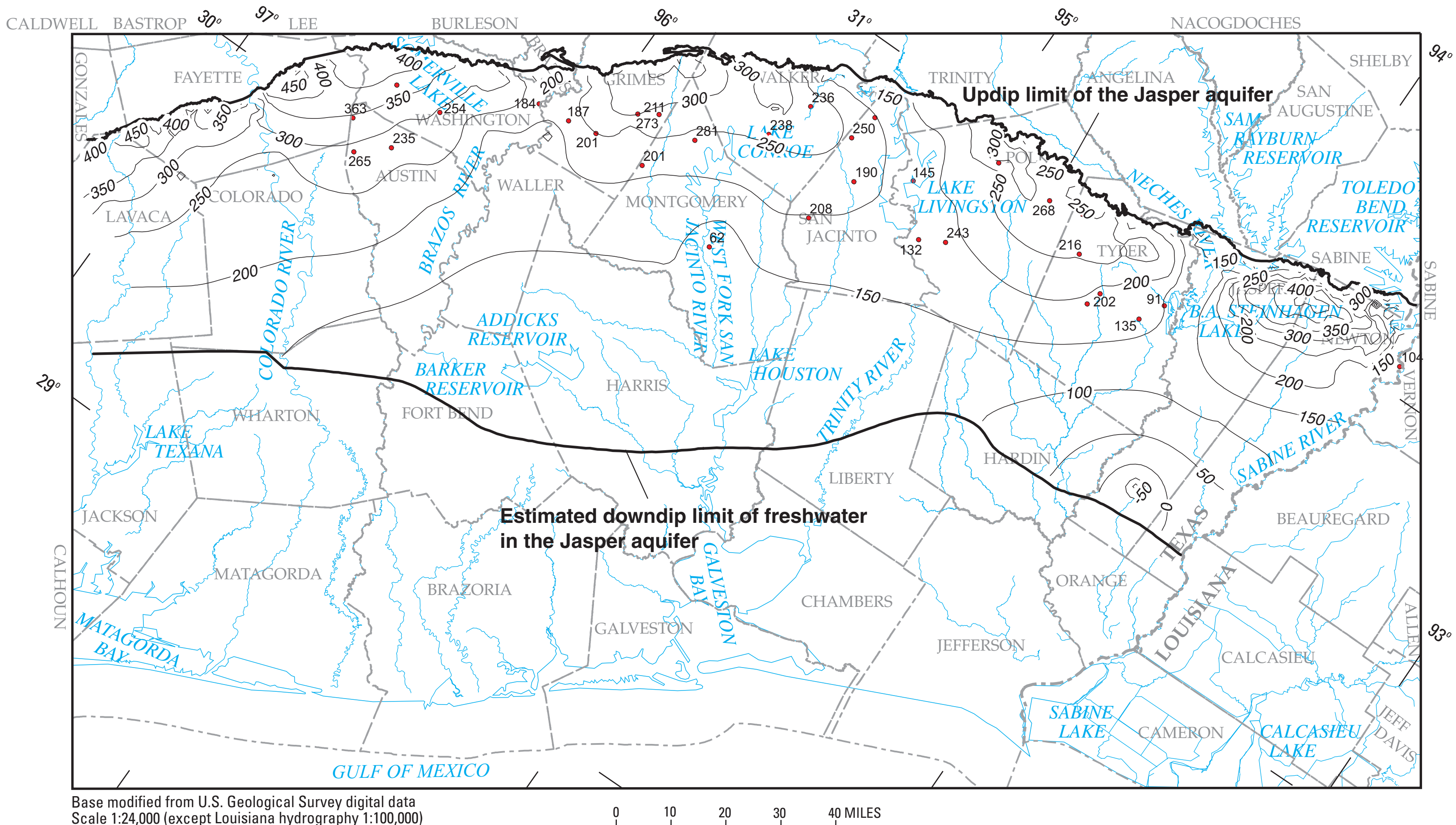
EXPLANATION

-150 **Simulated potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Interval 50 feet. Datum is NGVD 29

-150 **Measured potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Interval 50 feet. Datum is NGVD 29

155 **Data point**—Well in which water-level measurement was made. Number is water-level altitude (shown in areas not having published contours)

Figure 49. Simulated and measured 1977 potentiometric surfaces of the Evangeline aquifer and 1977 water-level measurements from wells screened in the Evangeline aquifer (modified from Gabrysch, 1979) in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

—-150— **Potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Interval 50 feet. Datum is NGVD

211 **Data point**—Well in which water-level measurement was made. Number is water-level altitude (shown in areas not having published contours)

Figure 50. Simulated 1977 potentiometric surface of the Jasper aquifer and 1977 water-level measurements from wells screened in the Jasper aquifer in the Ground-Water Availability Model area.

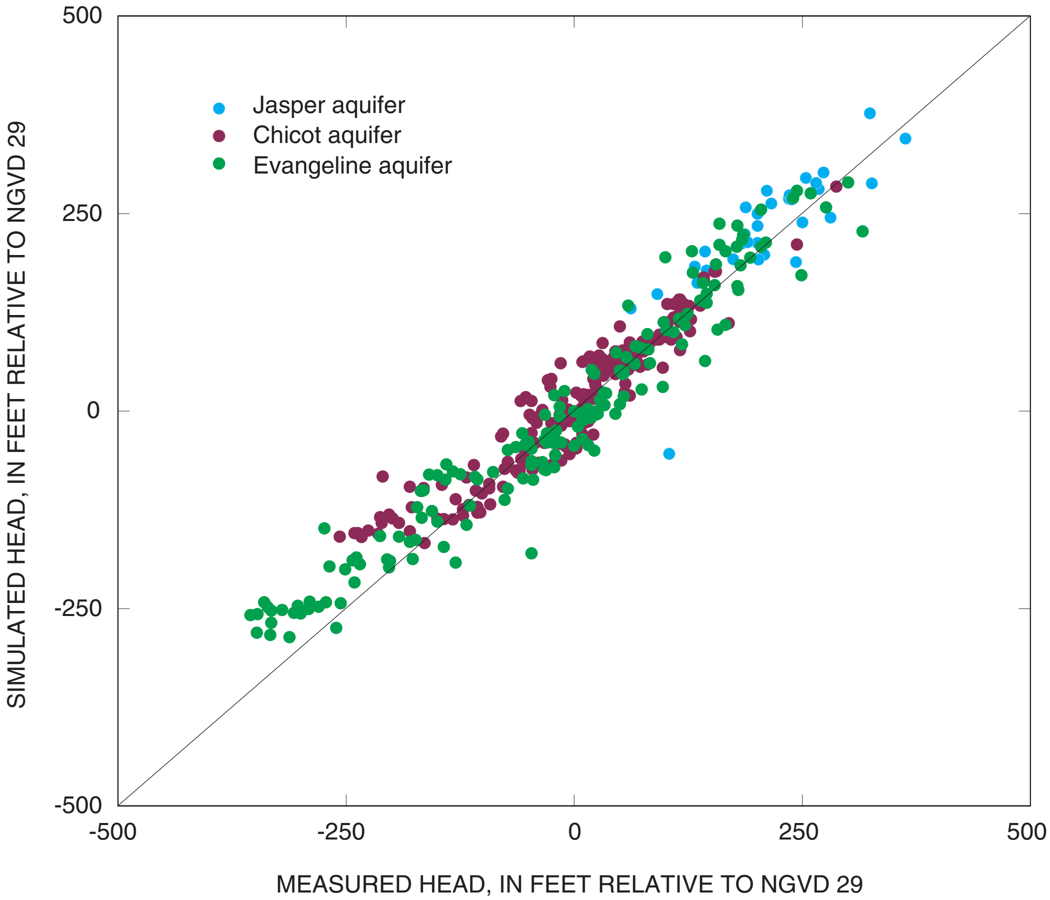


Figure 51. Relations between simulated and measured 1977 heads for the Chicot, Evangeline, and Jasper aquifers in the Ground-Water Availability Model.

the withdrawal data for the area for 1977 underestimate the actual ground-water withdrawals for that period.

In the Evadale-Beaumont area, the simulated cone of depression in the Chicot aquifer for 2000 is not appreciably different from that for 1977, but the 2000 cone in the Evangeline aquifer is larger and about 150 ft deeper (from -100 to -250 ft relative to NGVD 29) than the cone of 1977. One measured head in the cone in 2000 was 180 ft below NGVD 29. Simulated withdrawals in the area increased about 85 percent between 1977 and 2000.

The RMS errors for the three aquifer potentiometric surfaces for 2000 were about 31 ft for the Chicot aquifer, about 40 ft for the Evangeline aquifer, and about 34 ft for the Jasper aquifer (table 2). The RMS errors are about 8, 6, and 11 percent, respectively, of the total range in measured heads for the respective aquifers.

Graphical comparison of simulated and measured 2000 heads for the Chicot aquifer (fig. 58) shows some bias toward simulated heads less than measured heads from the middle to lower range of head values. For the Evangeline aquifer (fig. 58), little bias toward simulated heads greater than or less than measured heads is evident; and for the Jasper aquifer (fig. 58), as for 1977 heads, more simulated heads are greater than measured heads throughout the entire range of head values.

The distribution of 2000 head residuals for the Chicot aquifer (fig. 59) shows simulated heads less than measured heads over most of the GAM area except for areas along the updip limit of the aquifer (areas of higher heads), which is consistent with the graph of figure 58. For the Evangeline aquifer (fig. 60), the GAM area is more evenly divided between areas of simulated heads less than and greater than measured heads; however, except for a strip along the Sabine River, areas of simulated heads greater than measured heads tend to be in the northwestern part of the GAM area, and areas of simulated heads less than measured heads tend to be in the southwestern part of the GAM area. For the Jasper aquifer (fig. 61), simulated 2000 heads tend to be greater than measured heads over most of the GAM area, which is consistent with the graph of figure 58.

Assuming that ground water flows downgradient and perpendicular to equipotential lines, predevelopment potentio-

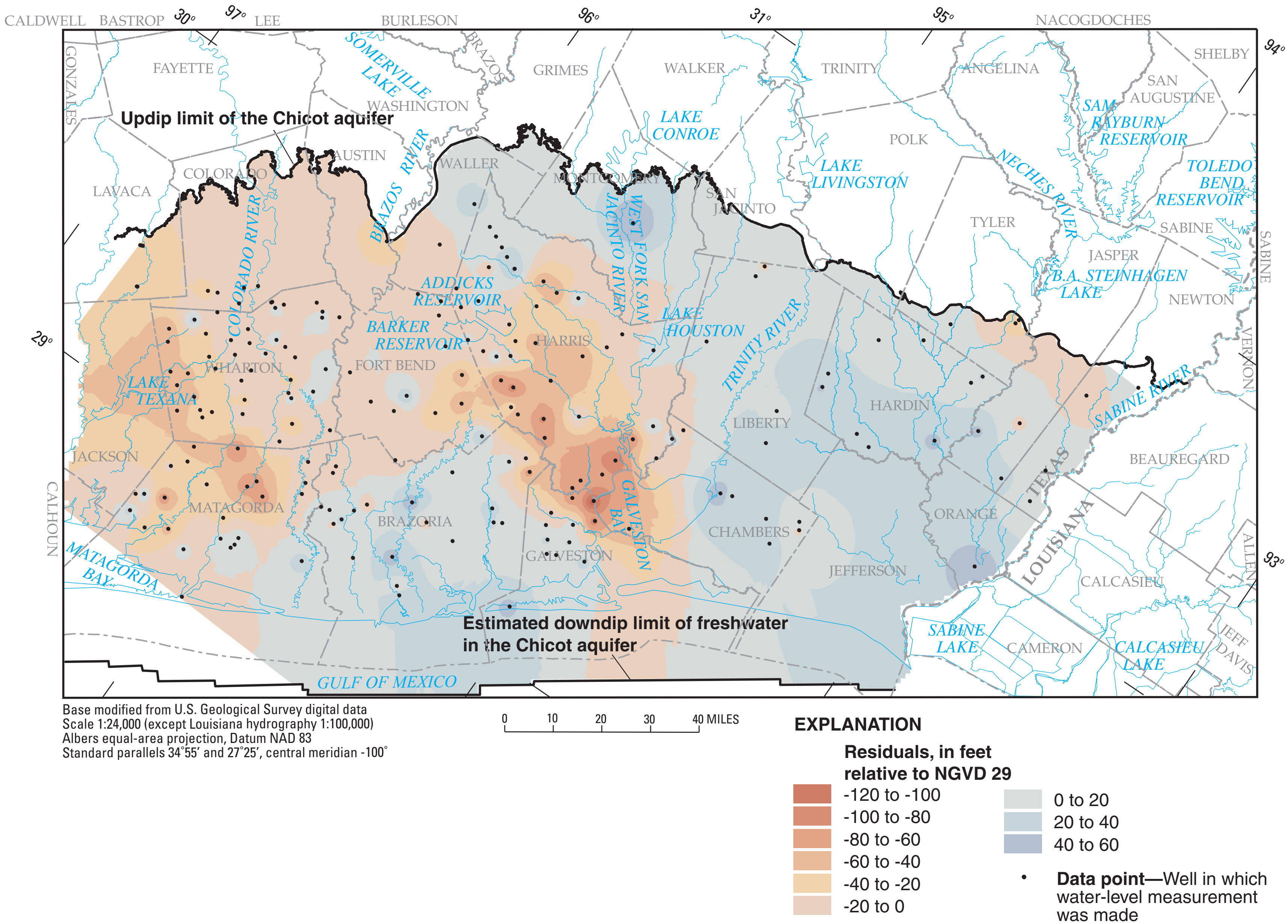
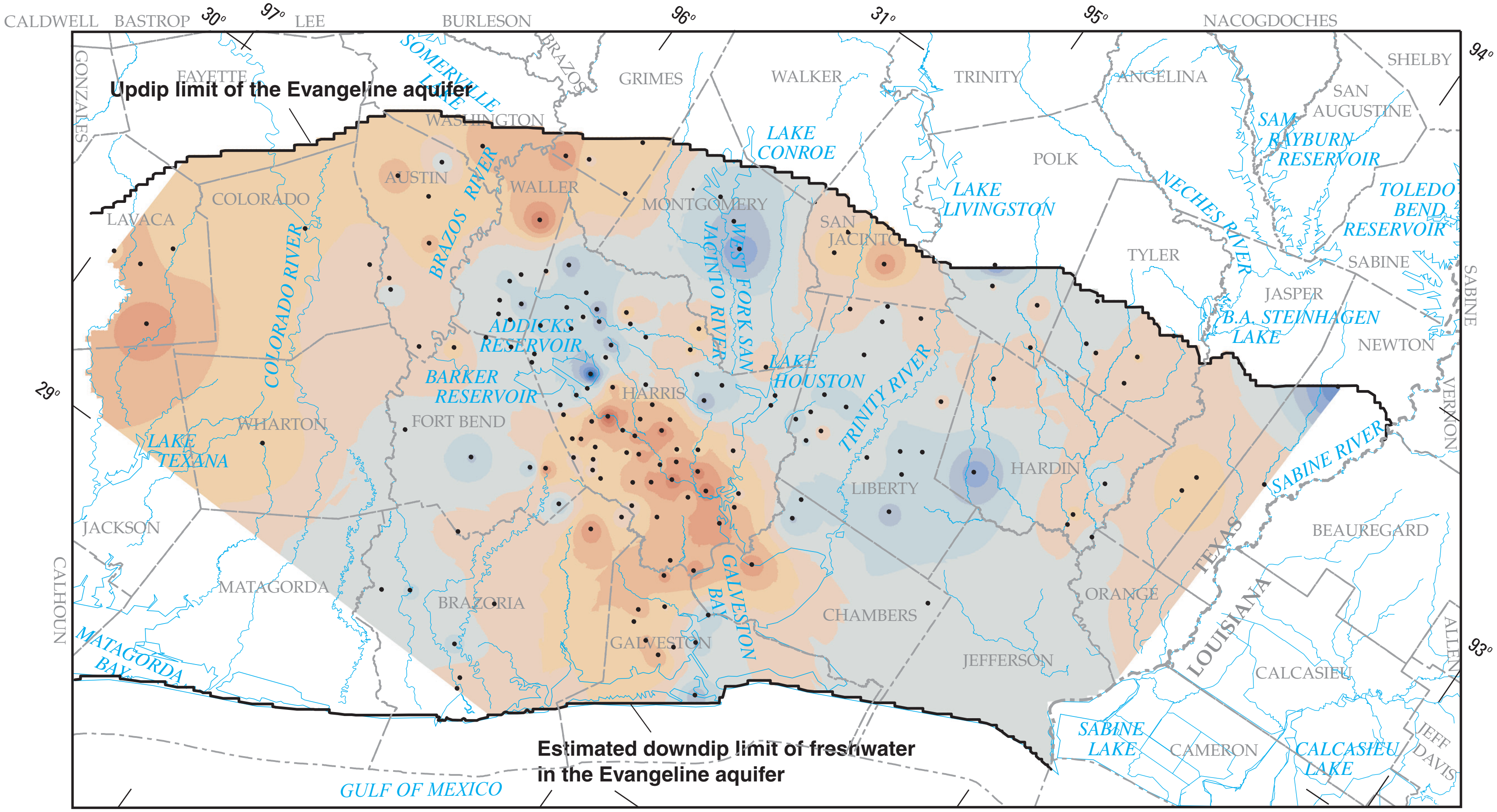


Figure 52. Distribution of water-level (head) residuals (measured minus simulated heads) for the Chicot aquifer, 1977, in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

0 10 20 30 40 MILES

EXPLANATION

Residuals, in feet relative to NGVD 29

-140 to -120
-120 to -100
-100 to -80
-80 to -60
-60 to -40
-40 to -20
-20 to 0
0 to 20
20 to 40
40 to 60
60 to 80
80 to 100
100 to 120
120 to 140

• **Data point**—Well in which water-level measurement was made

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 53. Distribution of water-level (head) residuals (measured minus simulated heads) for the Evangeline aquifer, 1977, in the Ground-Water Availability Model area.

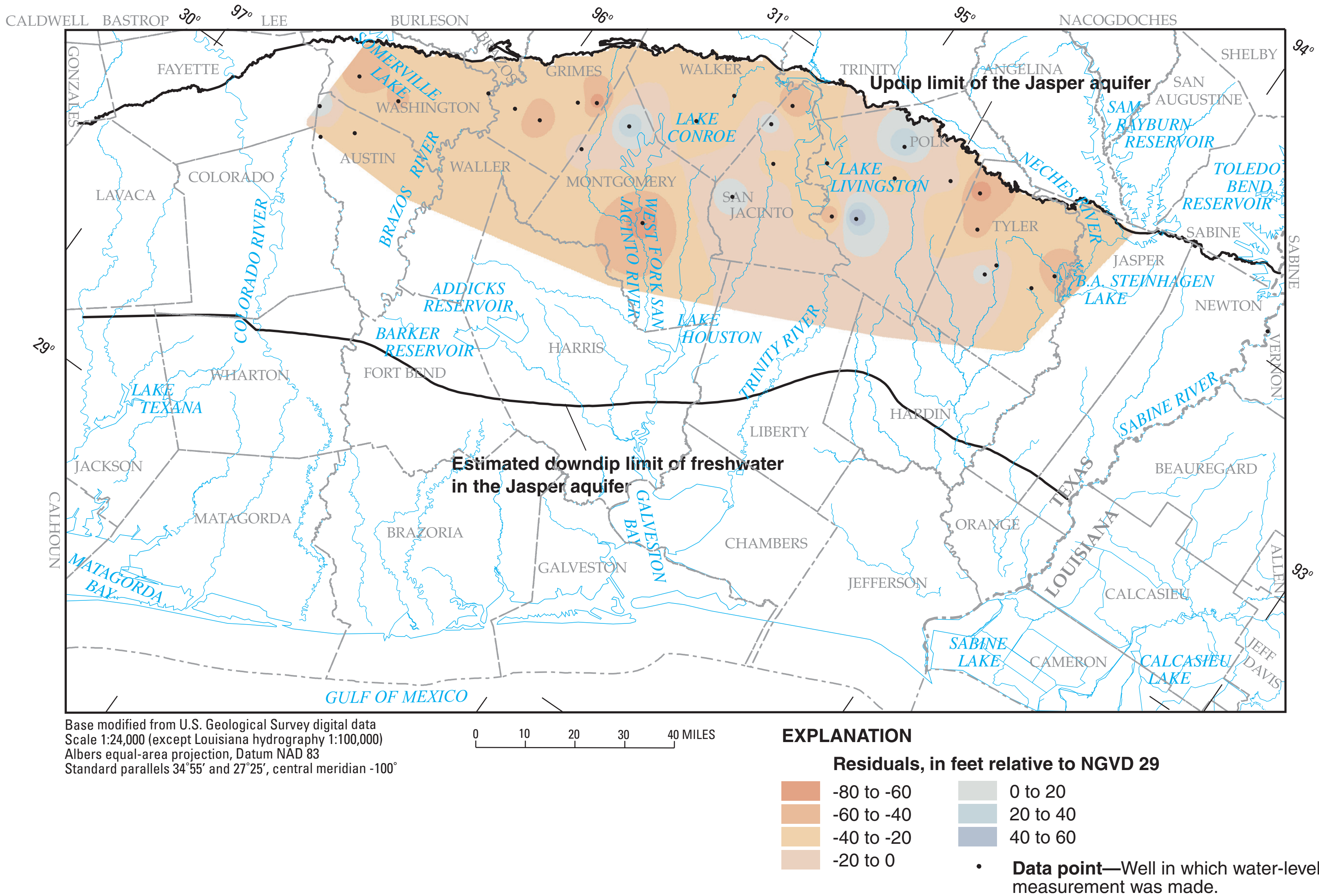
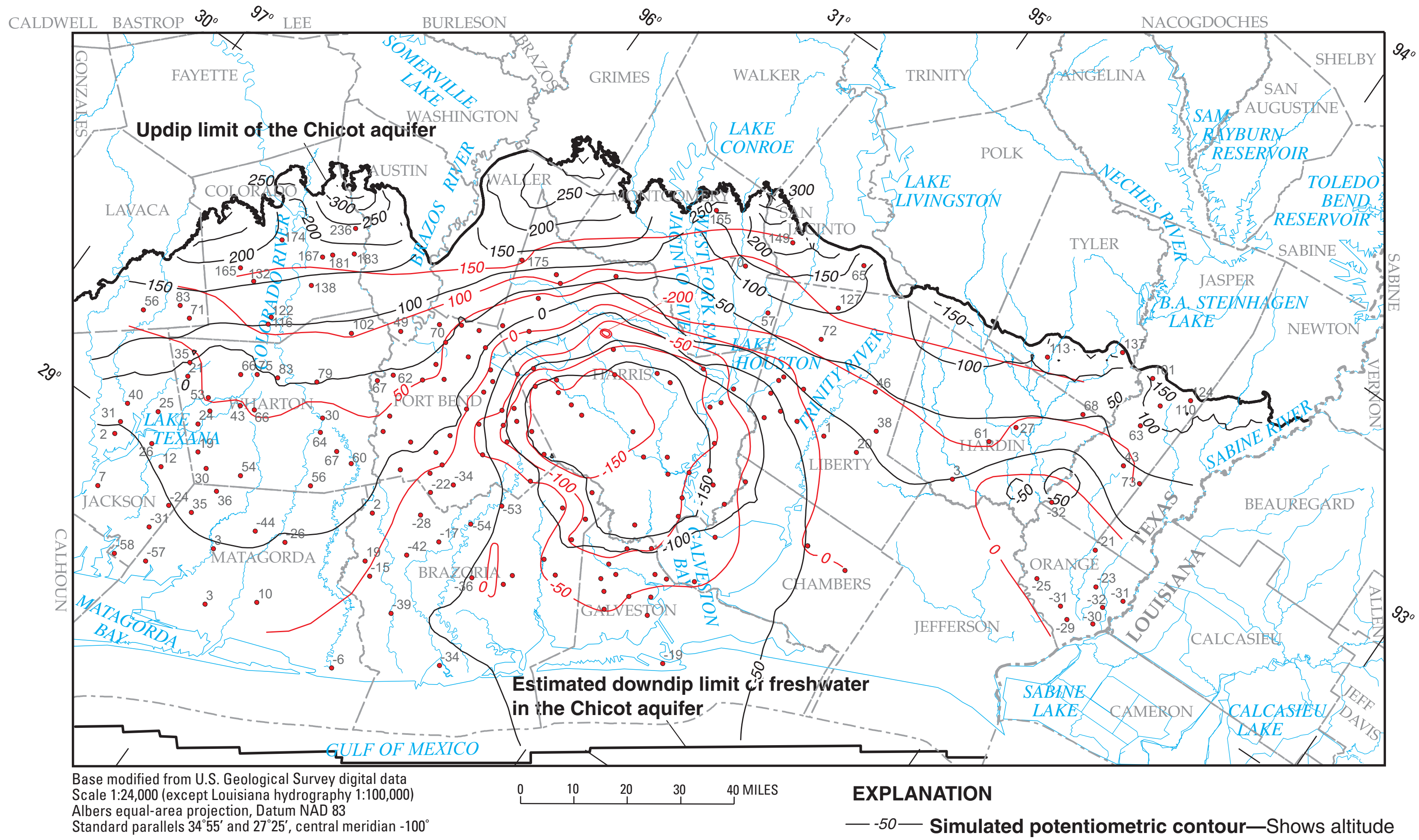


Figure 54. Distribution of water-level (head) residuals (measured minus simulated heads) for the Jasper aquifer, 1977, in the Ground-Water Availability Model area.



EXPLANATION

—— -50 —— **Simulated potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Interval 50 feet. Datum is NGVD 29

—— -50 —— **Measured potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Interval 50 feet. Datum is NGVD 29

-17 **Data point**—Well in which water-level measurement was made. Number is water-level altitude (shown in areas not having published water-level contours)

Figure 55. Simulated and measured 2000 potentiometric surfaces of the Chicot aquifer and 2000 water-level measurements from wells screened in the Chicot aquifer (modified from Coplin and Santos, 2000) in the Ground-Water Availability Model area.

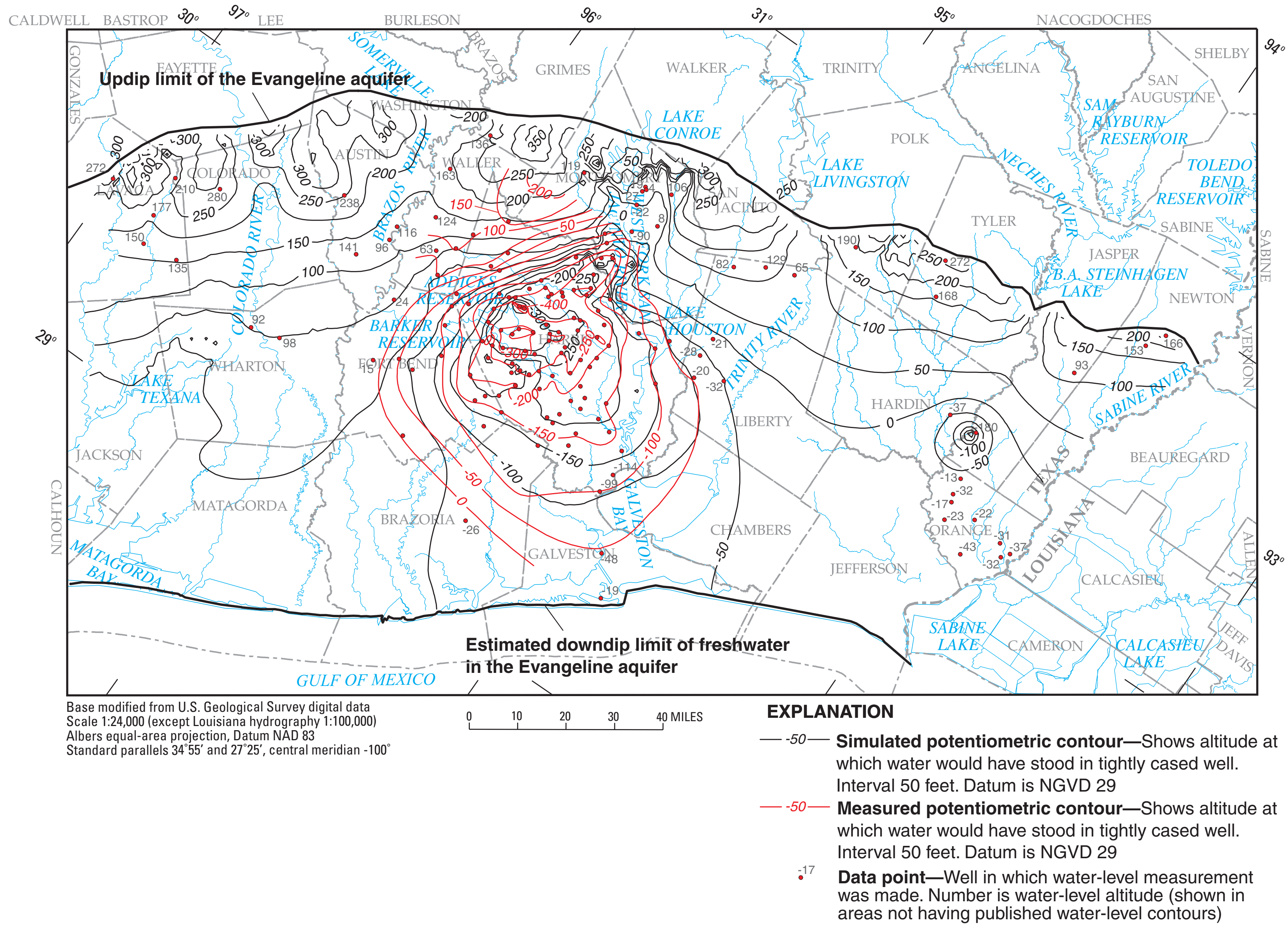
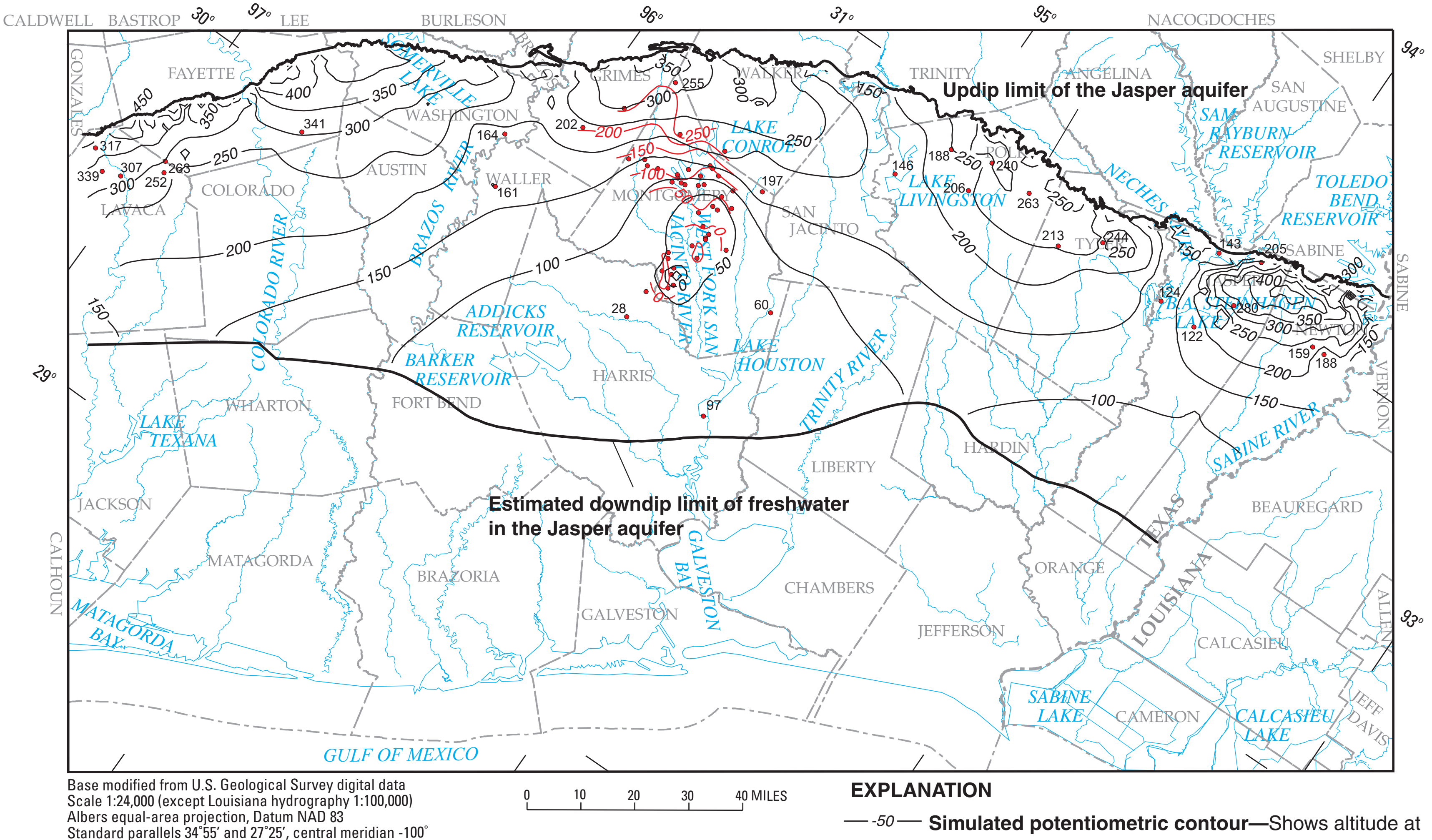


Figure 56. Simulated and measured 2000 potentiometric surfaces of the Evangeline aquifer and 2000 water-level measurements from wells screened in the Evangeline aquifer (modified from Coplin and Santos, 2000) in the Ground-Water Availability Model area.



EXPLANATION

-50 **Simulated potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Interval 50 feet. Datum is NGVD 29

-50 **Measured potentiometric contour**—Shows altitude at which water would have stood in tightly cased well. Interval 50 feet. Datum is NGVD 29

255 **Data point**—Well in which water-level measurement was made. Number is water-level altitude (shown in areas not having published water-level contours)

Figure 57. Simulated and measured 2000 potentiometric surfaces of the Jasper aquifer and 2000 water-level measurements from wells screened in the Jasper aquifer (modified from Coplin and Santos, 2000) in the Ground-Water Availability Model area.

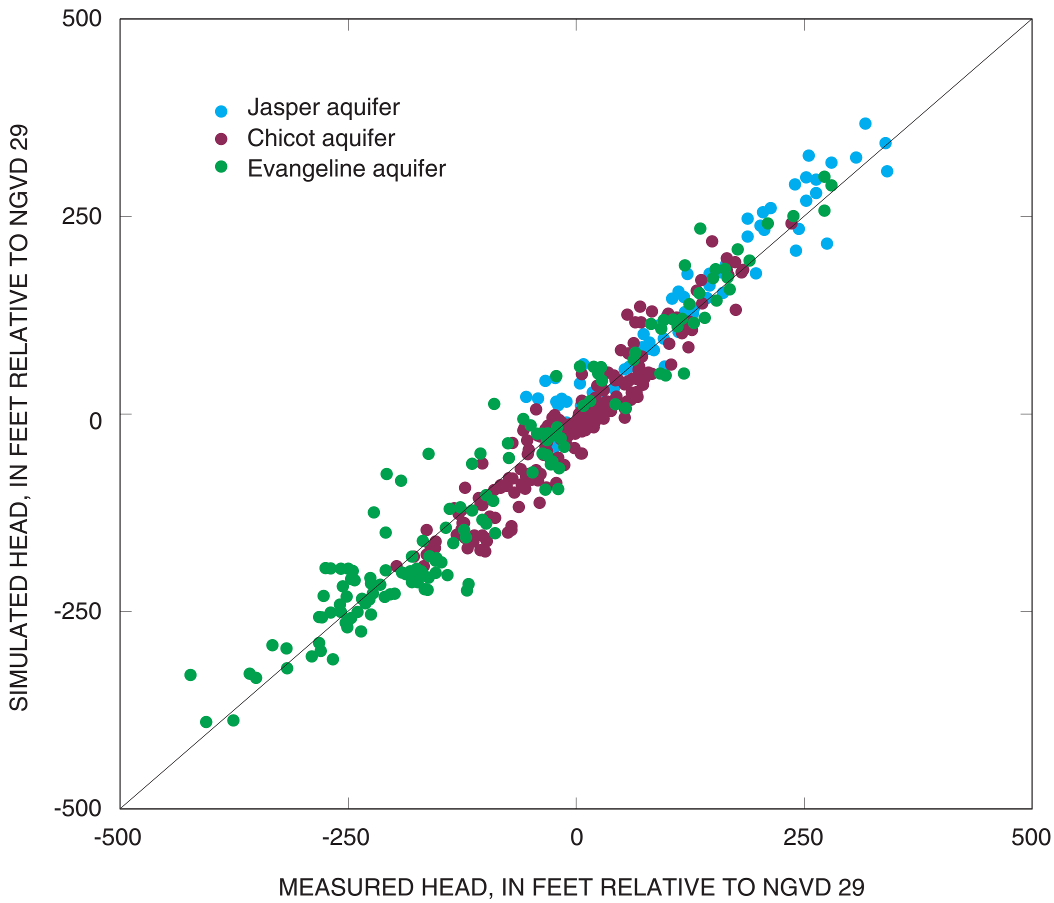


Figure 58. Relations between simulated and measured 2000 heads for the Chicot, Evangeline, and Jasper aquifers in the Ground-Water Availability Model.

metric surfaces of the three aquifers (figs. 62–64) confirm the generalized conceptual model of the natural ground-water-flow system: Recharge enters the system in topographically high outcrops of the hydrogeologic units in the northwestern part of the GAM area, flows either relatively short distances to discharge into topographically lower stream areas, or longer distances southeastward, through deeper zones, where it is discharged by upward leakage in topographically low areas near the coast.

Simulated and Measured Hydrographs

The locations of wells for which long-term hydrographs are available primarily in the Houston area, the coastal irrigation area, and selected counties away from those areas of withdrawal, including two likely in the Evadale-Beaumont withdrawal area, are shown in figure 65. The simulated and measured hydrographs for the Chicot aquifer in Galveston and Harris Counties (fig. 66) match closely relative to the ranges of change. The Galveston County hydrographs (fig. 66a, b) reflect generally declining heads through the mid-1970s followed by rises associated with decreased withdrawals. The simulated and measured hydrographs for the Evangeline aquifer in Harris County (fig. 67) also match closely relative to the ranges of change. The simulated and measured hydrographs for the Jasper aquifer in Harris and Montgomery Counties (fig. 68) match slightly less closely than those for the Chicot and Evangeline aquifers in the Houston area, but the trends in simulated and measured heads are similar.

For the Chicot aquifer in Wharton and Matagorda Counties (the primary source of withdrawals in the coastal irrigation area), the simulated and measured hydrographs (fig. 69a–c) match less closely than those for the Chicot and Evangeline aquifers in the Houston area; simulated heads generally are greater than measured heads by several tens of feet. However, the trends in simulated heads generally match those of measured heads. The authors acknowledge that the calibration is less reliable in the coastal irrigation area than in the Houston area.

Away from the Houston and coastal irrigation areas of withdrawal, the simulated and measured hydrographs for a

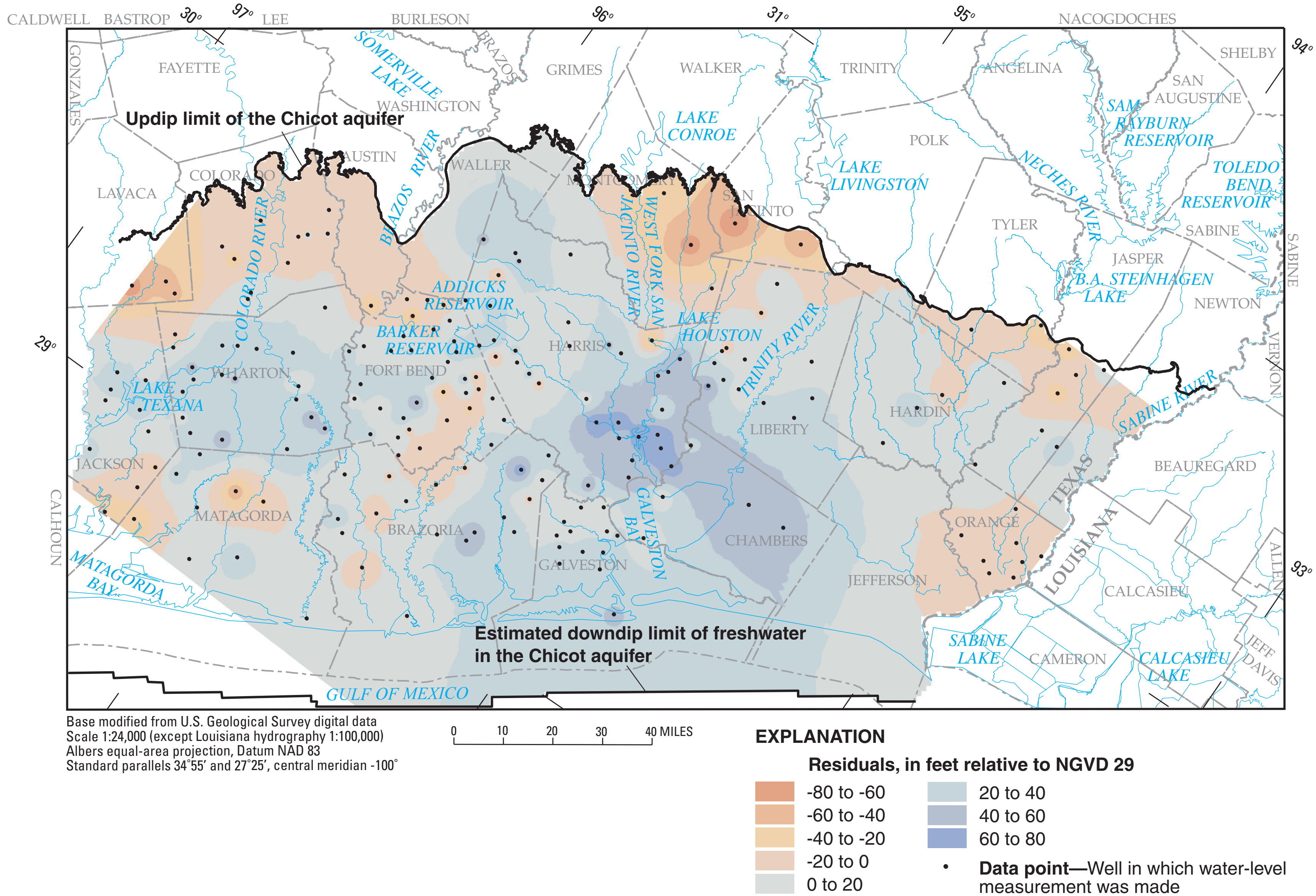
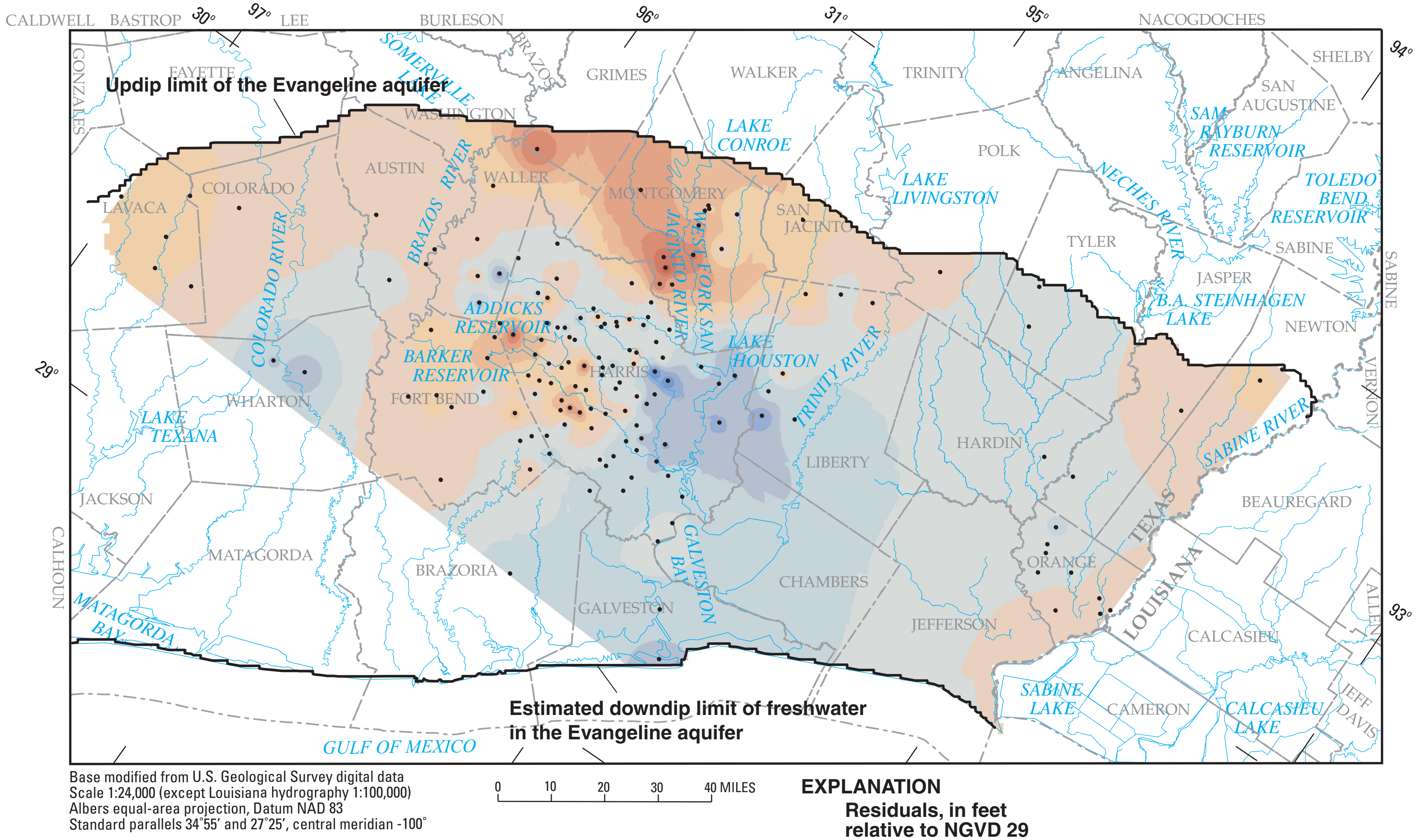


Figure 59. Distribution of water-level (head) residuals (measured minus simulated heads) for the Chicot aquifer, 2000, in the Ground-Water Availability Model area.



EXPLANATION

Residuals, in feet relative to NGVD 29

-140 to -120	0 to 20
-120 to -100	20 to 40
-100 to -80	40 to 60
-80 to -60	60 to 80
-60 to -40	80 to 100
-40 to -20	100 to 120
-20 to 0	• **Data point**—Well in which water-level measurement was made

Note: Not all subranges of the feature, as listed in the explanation for completeness, are visible on the map because the areas within some subranges (typically at the extremes of the range) are too small at the scale of the map.

Figure 60. Distribution of water-level (head) residuals (measured minus simulated heads) for the Evangeline aquifer, 2000, in the Ground-Water Availability Model area.

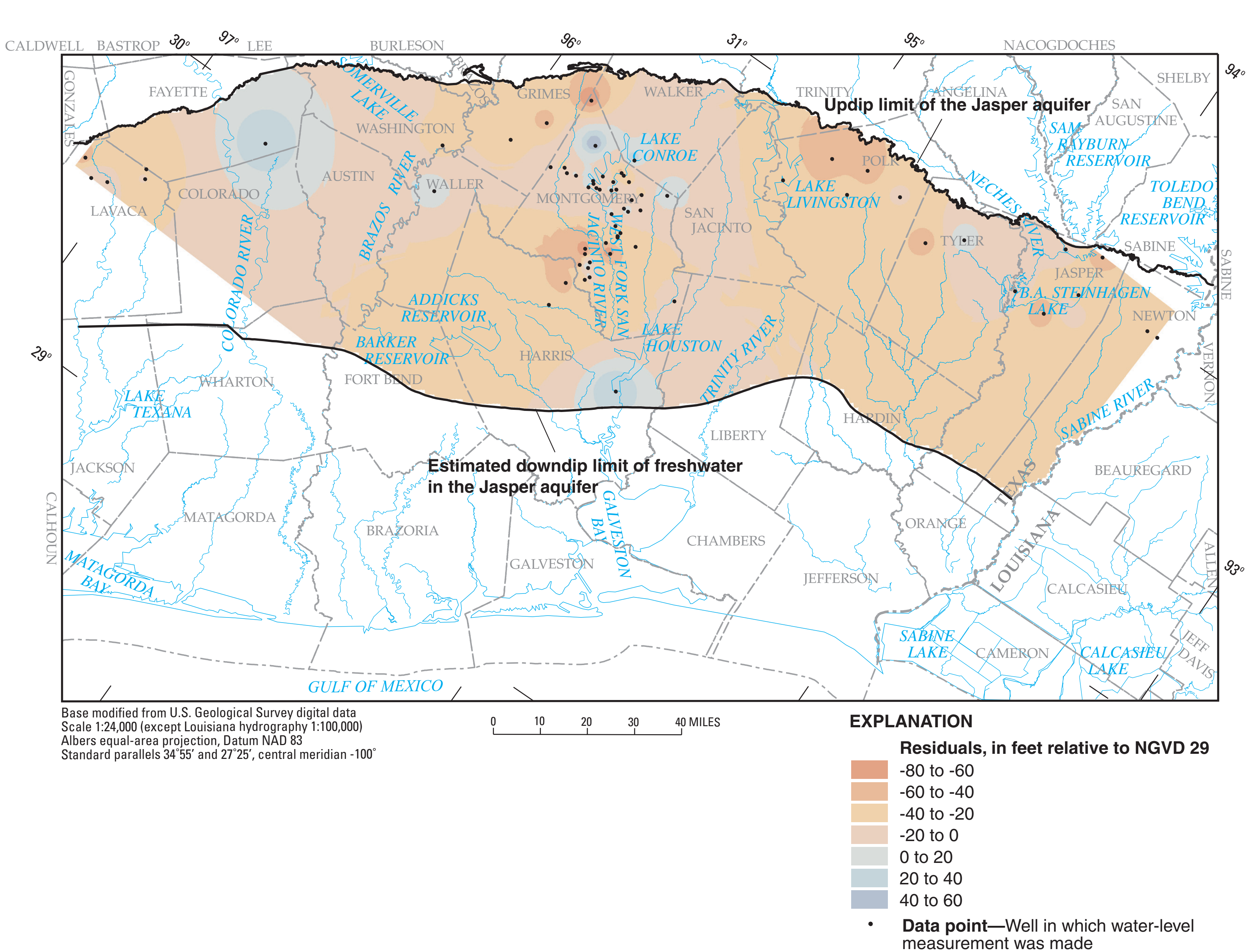
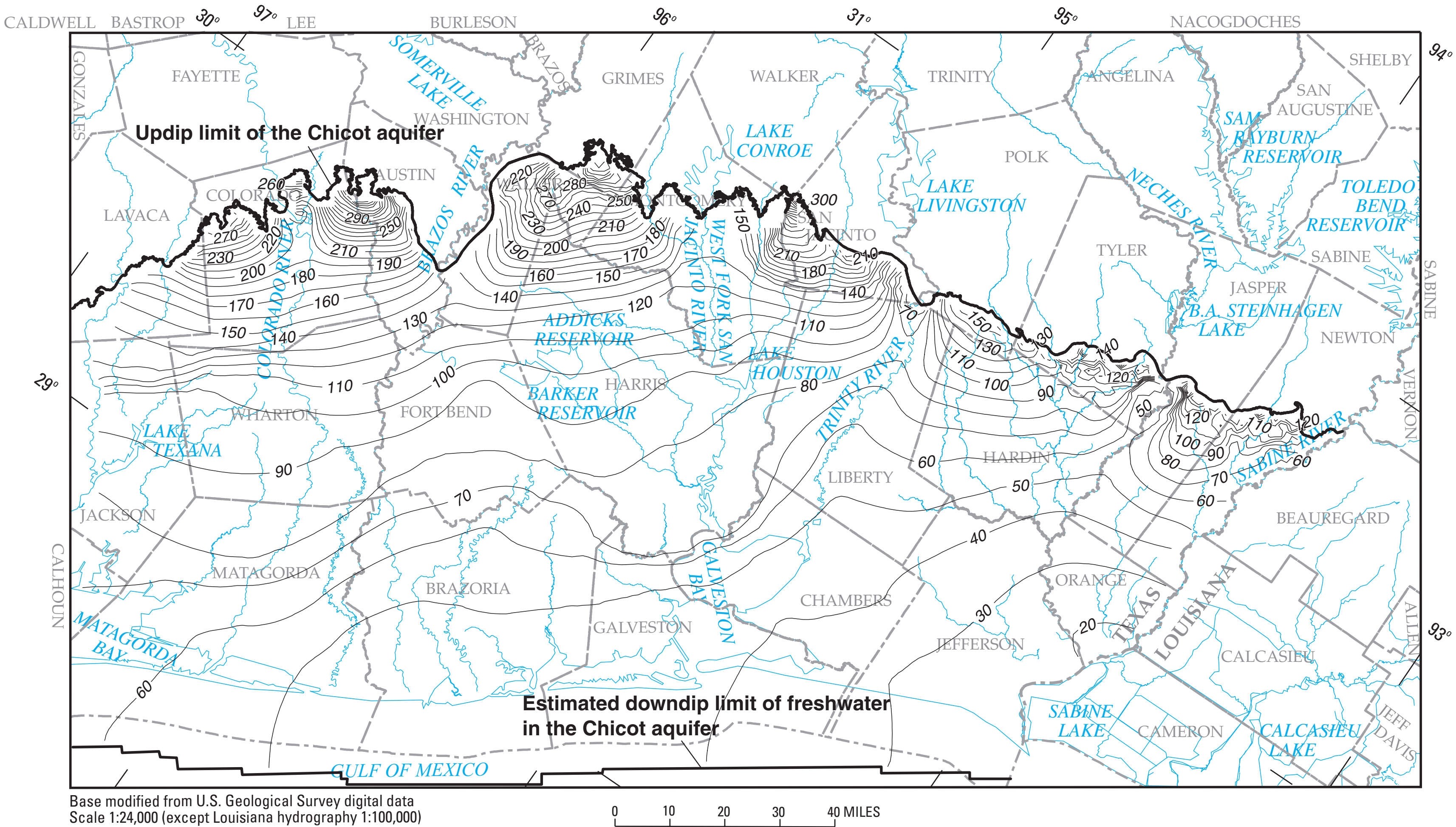


Figure 61. Distribution of water-level (head) residuals (measured minus simulated heads) for the Jasper aquifer, 2000, in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

—210— **Potentiometric contour—**Shows altitude at which water would have stood in tightly cased well. Interval 10 feet. Datum is NGVD 29

Figure 62. Simulated predevelopment potentiometric surface of the Chicot aquifer in the Ground-Water Availability Model area.

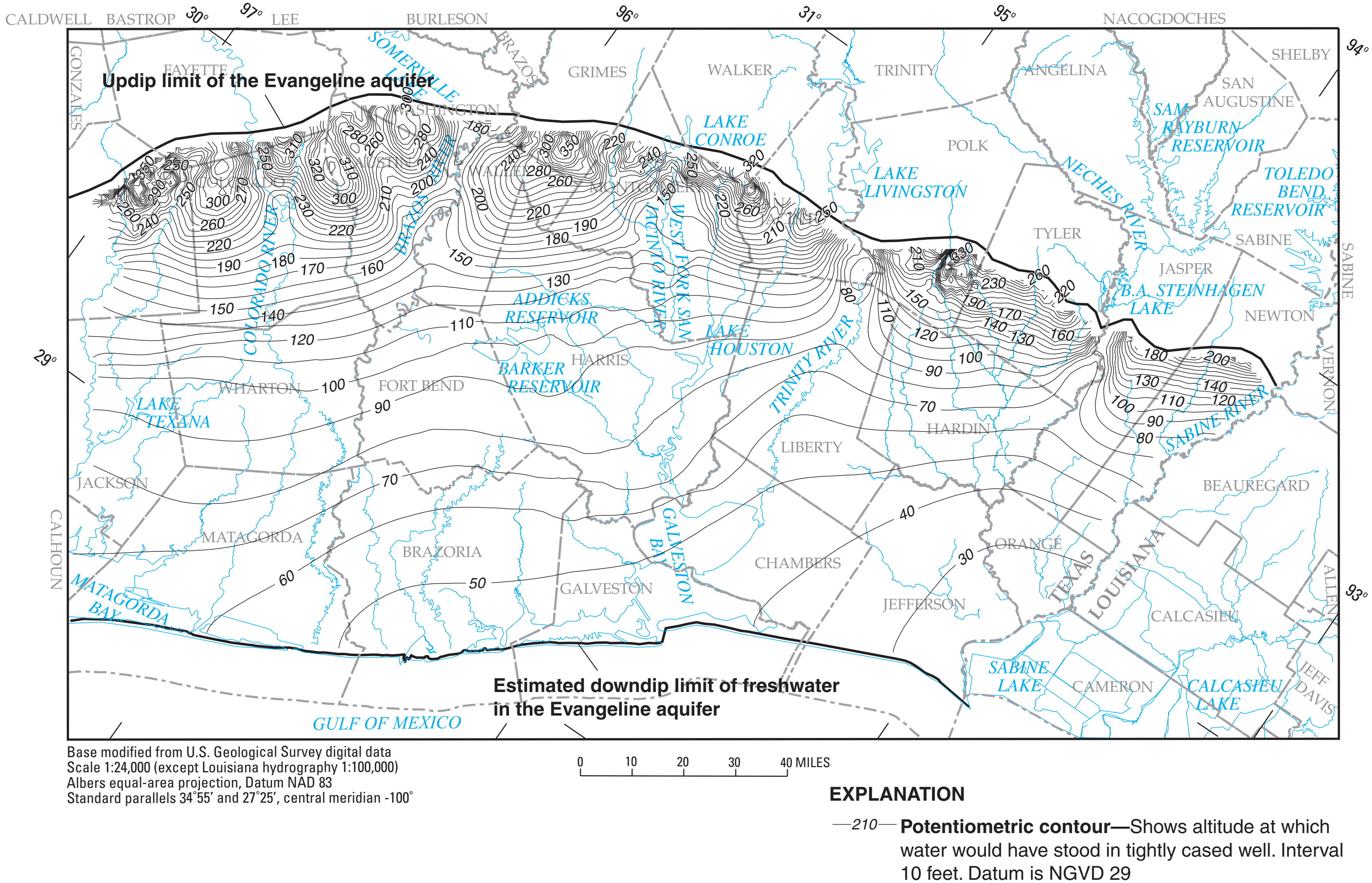
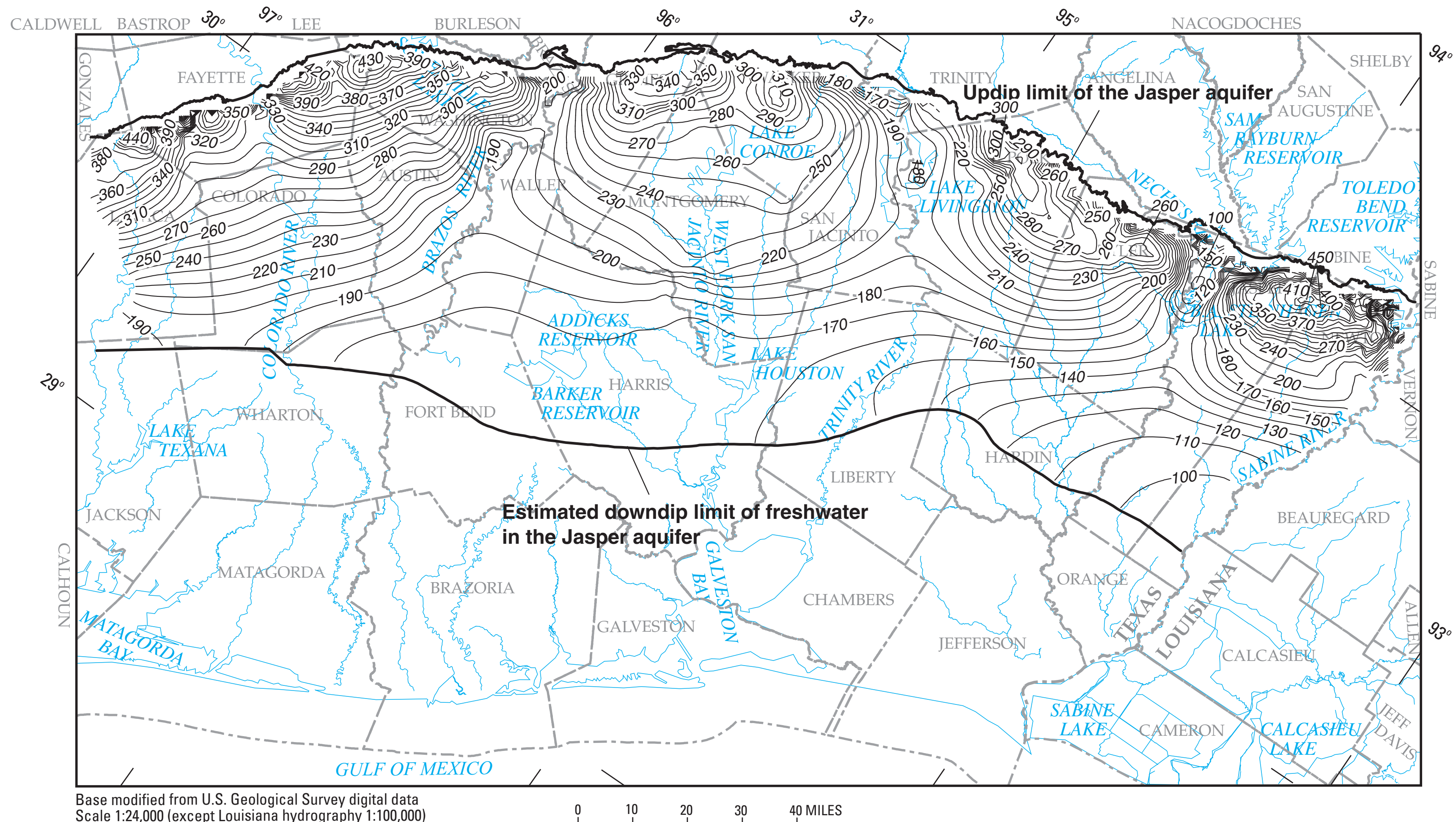


Figure 63. Simulated predevelopment potentiometric surface of the Evangeline aquifer in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

—*110*— **Potentiometric contour—**Shows altitude at which water would have stood in tightly cased well. Interval 10 feet. Datum is NGVD 29

Figure 64. Simulated predevelopment potentiometric surface of the Jasper aquifer in the Ground-Water Availability Model area.

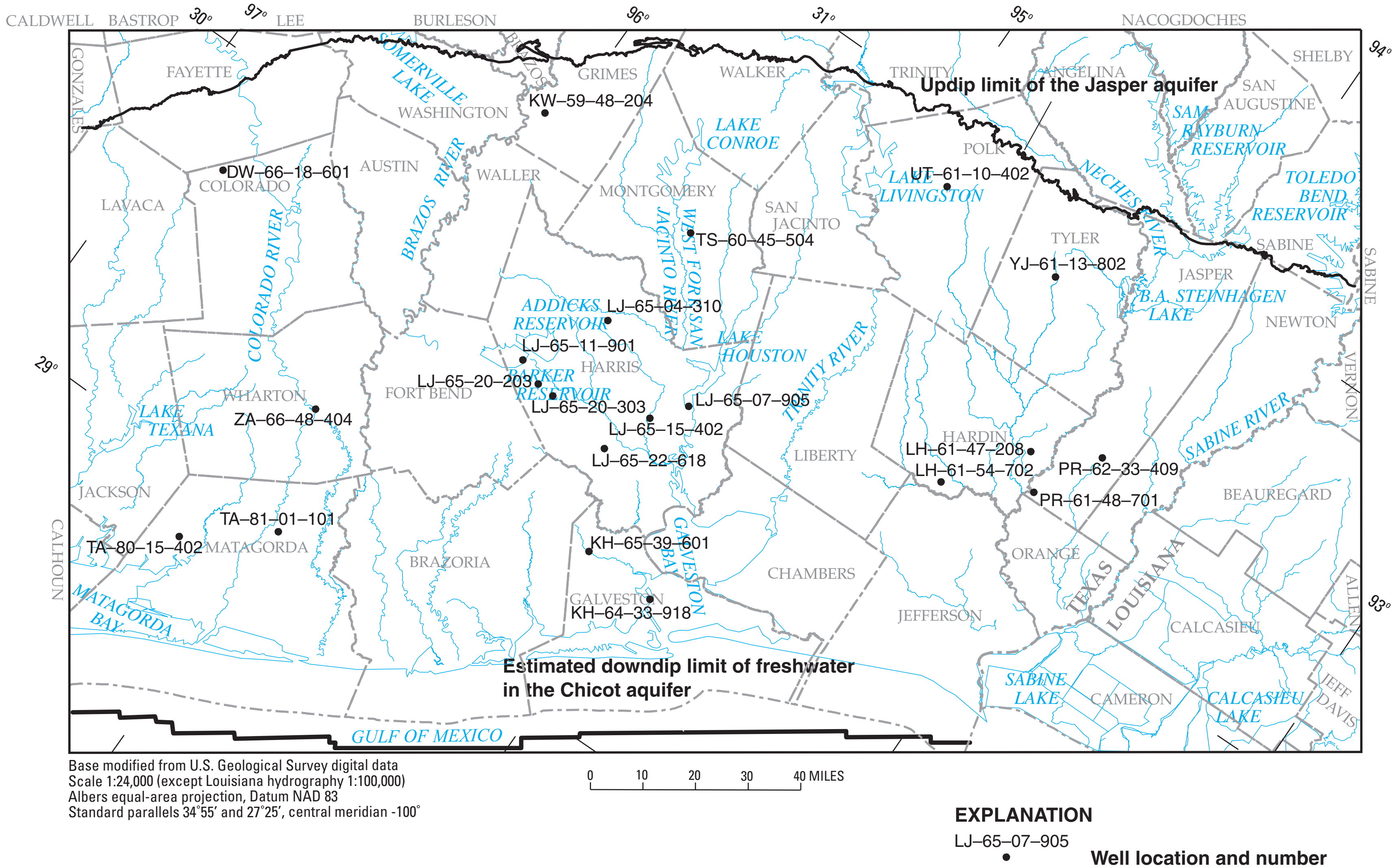


Figure 65. Locations of wells with hydrographs in the Ground-Water Availability Model area.

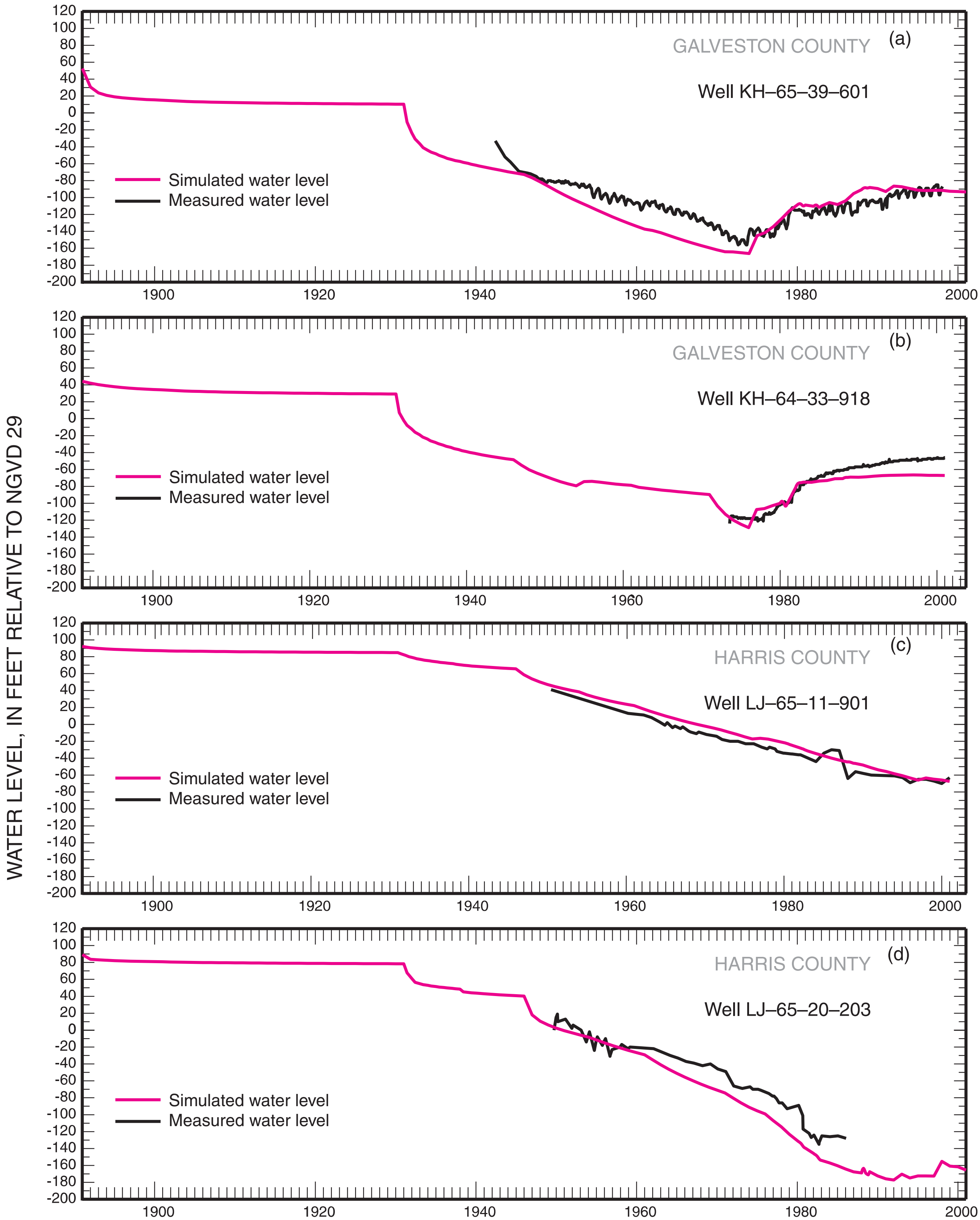


Figure 66. Hydrographs showing simulated and measured water levels in selected observation wells screened in the Chicot aquifer in Galveston and Harris Counties in the Ground-Water Availability Model area.

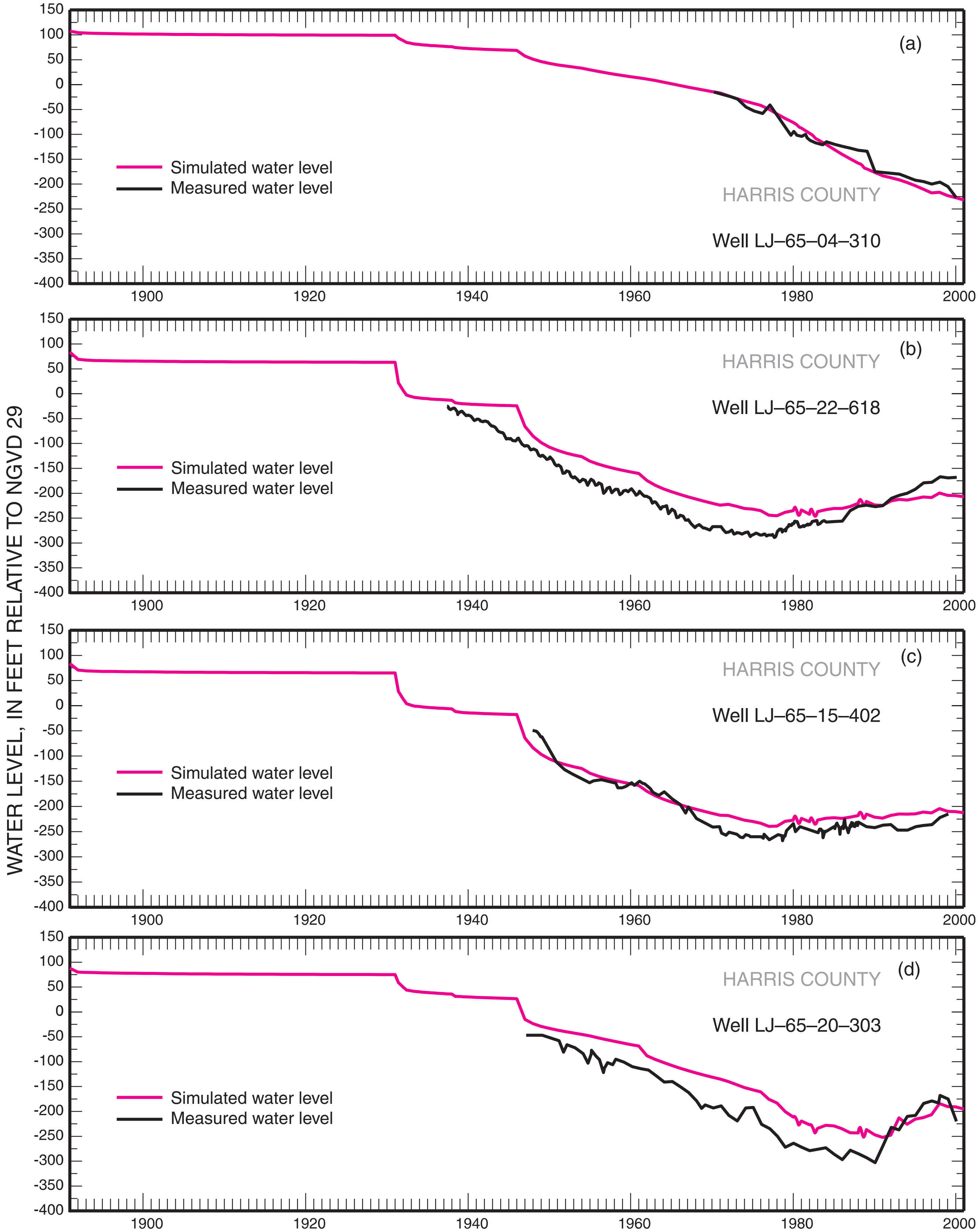


Figure 67. Hydrographs showing simulated and measured water levels in selected observation wells screened in the Evangeline aquifer in Harris County in the Ground-Water Availability Model area.

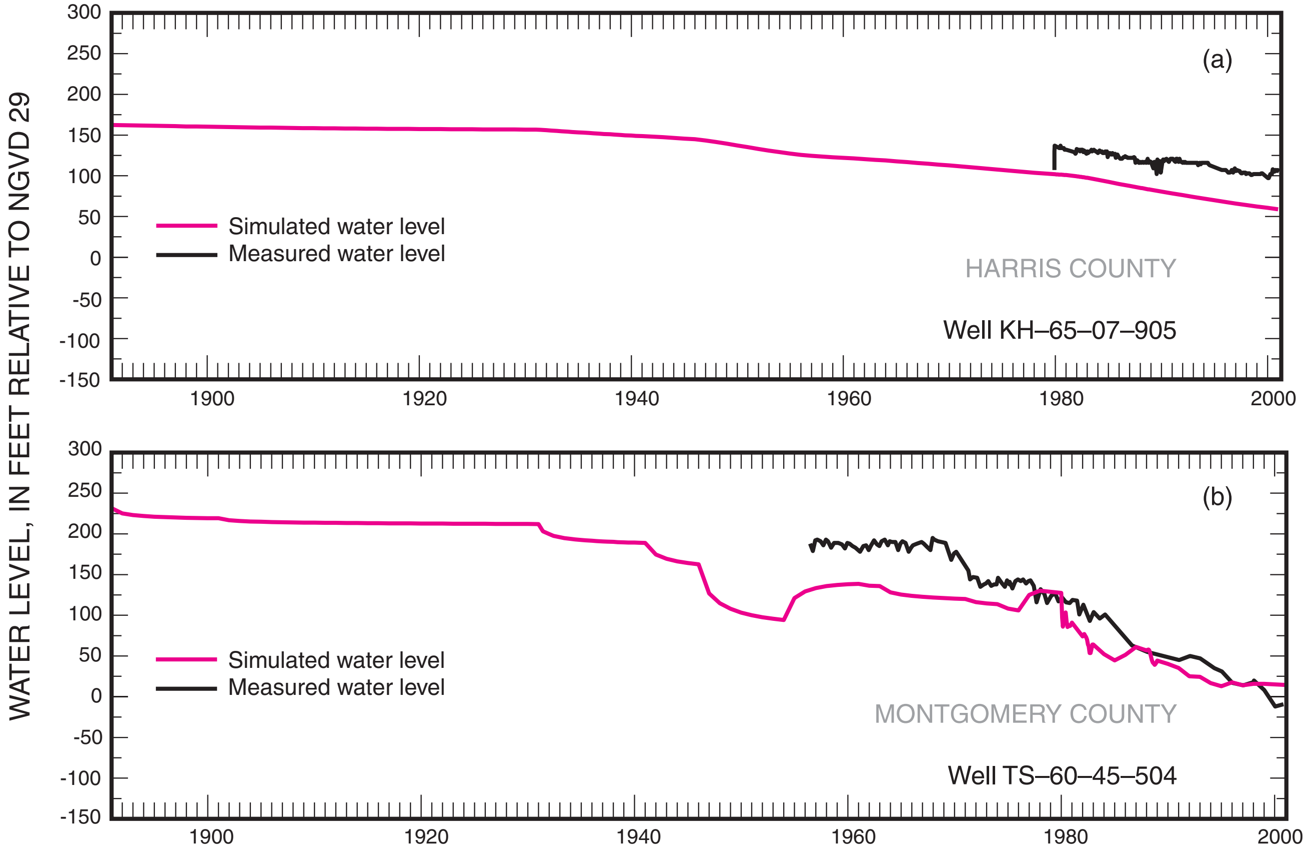


Figure 68. Hydrographs showing simulated and measured water levels in selected observation wells screened in the Jasper aquifer in Harris and Montgomery Counties in the Ground-Water Availability Model area.

Chicot aquifer well in Hardin County (fig. 69d) generally match within 20 ft, but the simulated heads are less than the measured heads throughout the coincident period. The simulated and measured hydrographs for an Evangeline aquifer well in Hardin County (fig. 70a), which appear to be within the area of influence of withdrawals in the Evadale-Beaumont area, match closely relative to the range of change. The simulated and measured hydrographs for one of the two Evangeline aquifer wells in Jasper County that also appear to be within the area of influence of Evadale-Beaumont withdrawals (fig. 70b) are relatively close, but an upward trend during the 1980s in the measured hydrograph was not simulated. Neither of the hydrographs from the two Evangeline aquifer wells probably within the area of influence of Evadale-Beaumont withdrawals shows heads substantially affected by those withdrawals. Another Evangeline aquifer simulated/measured hydrograph pair (fig. 70c) shows matching trends, but simulated heads are greater than measured heads by 30 to 40 ft. Four Jasper aquifer simulated/measured hydrograph pairs in Tyler, Polk, Grimes, and Colorado Counties (fig. 71), spanning the Jasper aquifer outcrop, match with varying degrees of closeness. Except for the Colorado County hydrographs, which probably are affected by irrigation withdrawals, the hydrographs are relatively flat.

Simulated and Estimated Water-Budget Components

Simulated recharge and discharge in outcrops of the hydrogeologic units, vertical leakage between units, changes in storage, and withdrawals for 1977 are summarized in figure 72. The diagram shows a net recharge of 555 ft^3/s (0.40 in/yr) in the Chicot aquifer outcrop, 19 ft^3/s (0.12 in/yr) in the Evangeline aquifer outcrop, negligible net recharge in the Burkeville confining unit outcrop, and 14 ft^3/s (0.06 in/yr) in the Jasper aquifer outcrop. For the entire system, the simulated net outcrop recharge for 1977 is 588 ft^3/s (0.32 in/yr). In terms of a water balance (within 5 ft^3/s) for the entire system in 1977, 757 ft^3/s of recharge plus 1,082 ft^3/s from depletion of sand storage (742 ft^3/s) and inelastic compaction of clays (340 ft^3/s) is offset by 169 ft^3/s of natural discharge and 1,670 ft^3/s (1,080 Mgal/d) of withdrawals. Thus in 1977, net recharge supplied about 35 percent of withdrawals, depletion of sand storage about 45 percent, and inelastic compaction of clays about 20 percent. Expressed as a percentage of an estimated 48 in/yr average rainfall over the entire aquifer-system area, outcrop recharge of 757 ft^3/s to all units is about 0.9 percent.

Simulated recharge and discharge in outcrops of the hydrogeologic units, vertical leakage between units, changes in

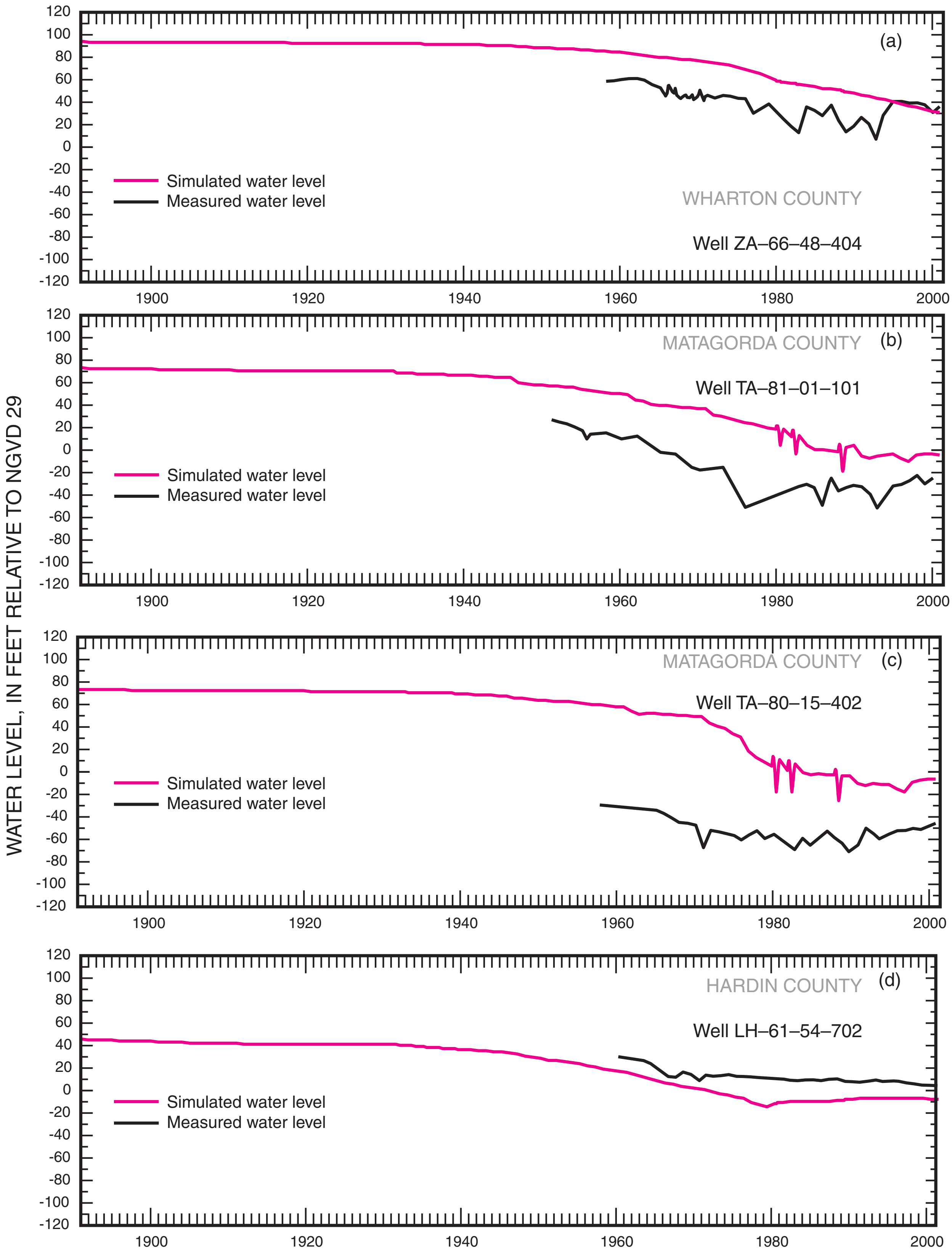


Figure 69. Hydrographs showing simulated and measured water levels in selected observation wells screened in the Chicot aquifer in Wharton, Matagorda, and Hardin Counties in the Ground-Water Availability Model area.

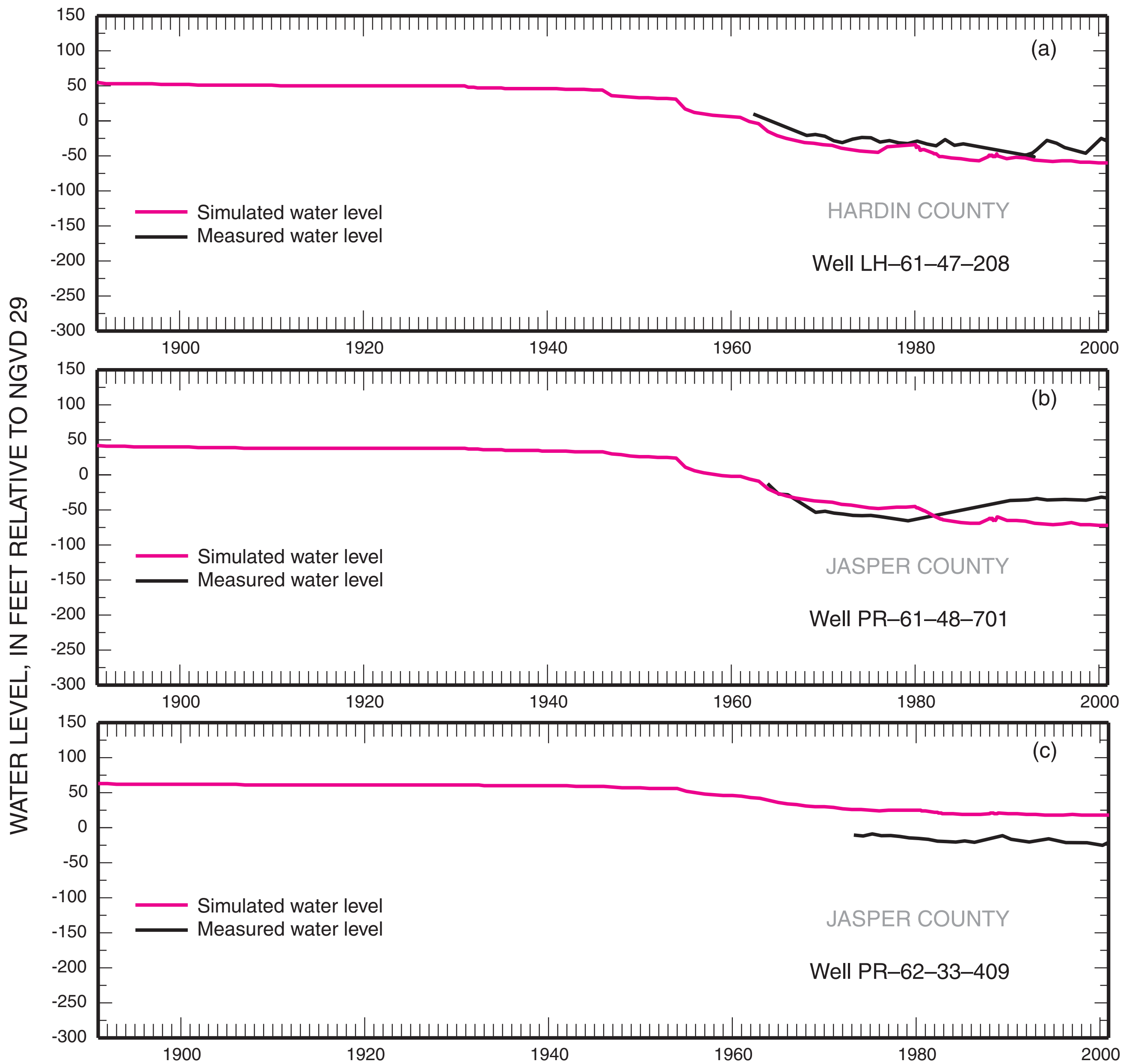


Figure 70. Hydrographs showing simulated and measured water levels in selected observation wells screened in the Evangeline aquifer in Jasper and Hardin Counties in the Ground-Water Availability Model area.

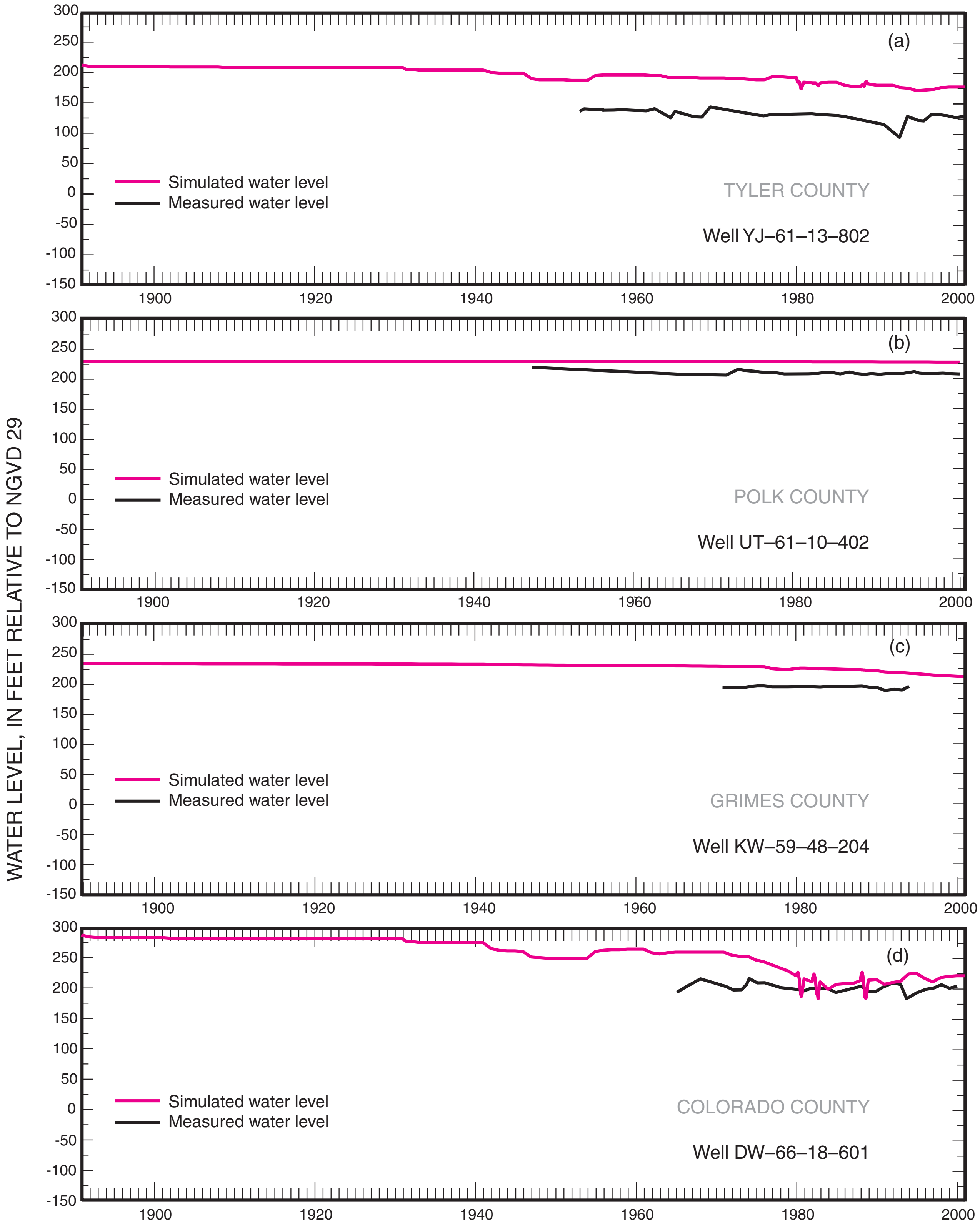
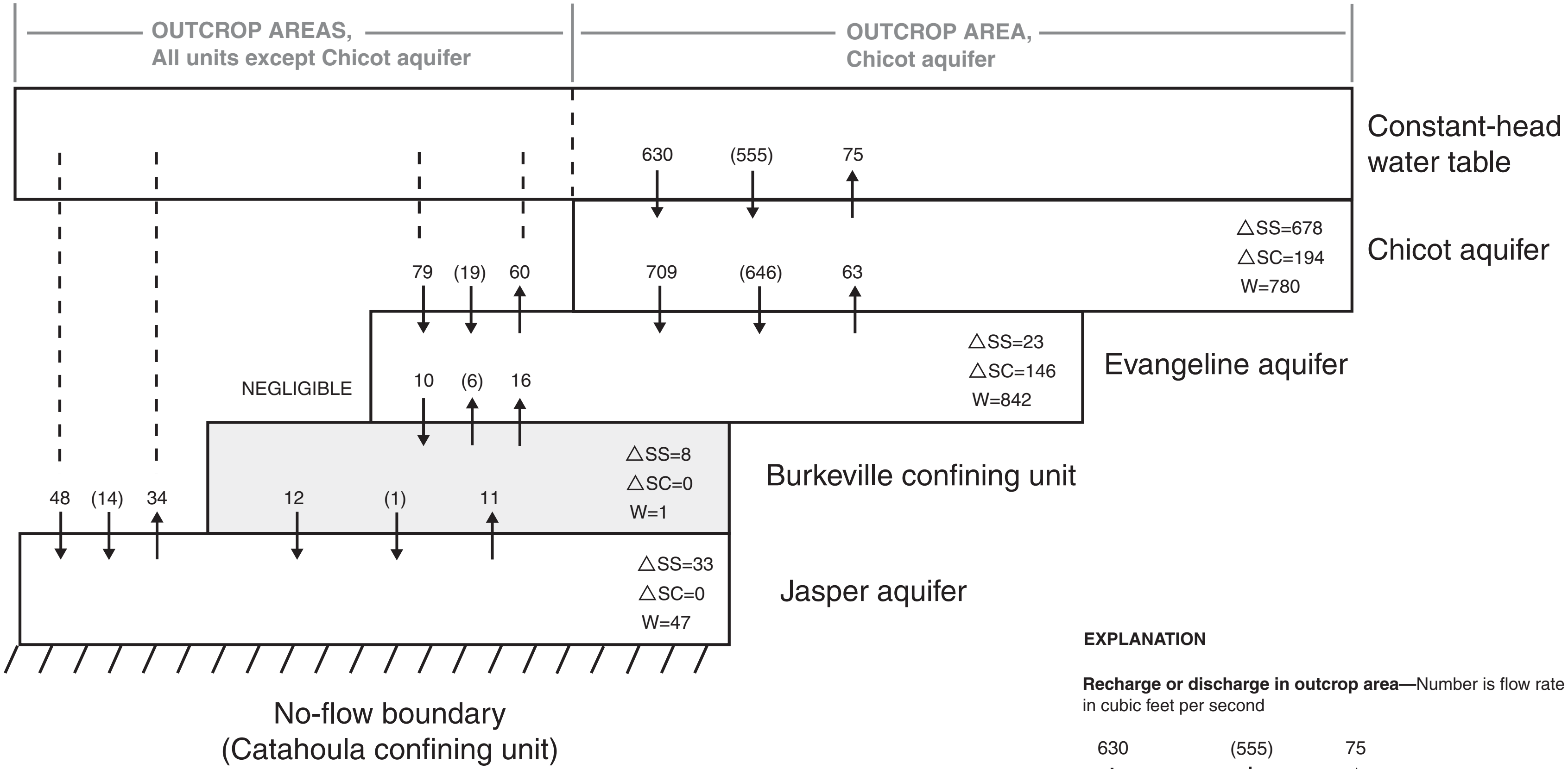


Figure 71. Hydrographs showing simulated and measured water levels in selected observation wells screened in the Jasper aquifer in Tyler, Polk, Grimes, and Colorado Counties in the Ground-Water Availability Model area.



EXPLANATION

Recharge or discharge in outcrop area—Number is flow rate in cubic feet per second

630 RECHARGE (555) NET 75 NATURAL DISCHARGE

Leakage through bottom of hydrogeologic unit—Number is flow rate in cubic feet per second

709 DOWNWARD (646) NET 63 UPWARD

△SS Net rate of release of water from sand storage, in cubic feet per second

△SC Net rate of release of water from inelastic clay compaction, in cubic feet per second

W Withdrawal rate, in cubic feet per second

WATER BALANCE

RECHARGE + △SS + △SC = NATURAL DISCHARGE + W

757 + 742 + 340 = 169 + 1,670

1,839 = 1,839

Figure 72. Simulated 1977 water-budget components of the hydrogeologic units in the Ground-Water Availability Model area.

storage, and withdrawals for 2000 are summarized in figure 73. The diagram shows a net recharge of 769 ft^3/s (0.55 in/yr) in the Chicot aquifer outcrop, 18 ft^3/s (0.11 in/yr) in the Evangeline aquifer outcrop, negligible net recharge in the Burkeville confining unit outcrop, and 17 ft^3/s (0.07 in/yr) in the Jasper aquifer outcrop. For the entire system, the simulated net outcrop recharge for 2000 is 804 ft^3/s (0.43 in/yr). In terms of a water balance (within 5 ft^3/s) for the entire system in 2000, 965 ft^3/s of recharge plus 516 ft^3/s from depletion of sand storage (410 ft^3/s) and inelastic compaction of clays (106 ft^3/s) is offset by 161 ft^3/s of natural discharge and 1,322 ft^3/s (854 Mgal/d) of withdrawals. Thus in 2000, net recharge supplied 61 percent of withdrawals, depletion of sand storage 31 percent, and inelastic compaction of clays 8 percent. Expressed as a percentage of an estimated 48 in/yr average rainfall over the entire aquifer-system area, outcrop recharge of 965 ft^3/s to all units is about 1.1 percent.

The most notable differences between the simulated water-budget components of 1977 and 2000, besides the fact withdrawals were about 21 percent less in 2000, are the increase in the percentage of withdrawals supplied by recharge and the decrease in the percentage of water supplied by depletion of storage and inelastic compaction of clays between 1977 and 2000. In the intervening 23 years, the amount of water available from storage and clay compaction decreased about 52 percent (from 1,082 to 516 ft^3/s). To sustain withdrawals, additional recharge was induced (recharge increased from 757 to 965 ft^3/s), and a small amount of natural discharge was captured (natural discharge decreased from 169 to 161 ft^3/s).

The simulated recharge rates for the GAM for 1977 and 2000 appear to be generally comparable to estimates of recharge rates from previous studies involving all or parts of the Gulf Coast aquifer system, as discussed in the "Ground-Water-Flow Conditions, Recharge, and Discharge" section of this report. Estimates of total recharge from two county studies (Sandeen, 1972; Loskot and others, 1982) were about 1.2 in/yr. However as previously discussed, finite-difference models on average simulate less than total recharge. A more comparable simulation study is that of Ryder and Ardis (2002). Although the model of that study encompassed the entire Gulf Coast aquifer system in Texas, recharge and discharge were simulated the same way as in the GAM. The Ryder and Ardis (2002) simulation for 1982 resulted in a net recharge rate of 0.25 in/yr; the GAM simulations for 1977 and 2000 resulted in net recharge rates of 0.32 and 0.43 in/yr, respectively. The recharge rates of Ryder and Ardis (2002) are expected to be smaller than the recharge rates of the GAM because of the difference in model scale (grid-block size of 10 mi^2 in that earlier model and 1 mi^2 in the GAM).

Simulated predevelopment recharge and discharge in outcrops of the hydrogeologic units and vertical leakage between units are summarized in figure 74. The diagram shows total recharge (and total discharge, as the system is steady state so there is no change in storage) of about 307 ft^3/s (0.17 in/yr). About two-thirds of the simulated flow in the predevelopment system occurs in the Chicot aquifer (195 ft^3/s recharge and 201 ft^3/s discharge). The nearly equal rates of recharge and discharge in the outcrop of the Evangeline aquifer (67 ft^3/s recharge and 70 ft^3/s discharge) are about the same as the rates of vertical leakage between the Evangeline and Chicot aquifers in the Evangeline subcrop (72 ft^3/s downward and 72 ft^3/s upward). Simulated Jasper aquifer outcrop recharge (45 ft^3/s) is about 30 percent more than outcrop discharge (35 ft^3/s); the difference is accounted for by greater upward leakage from the Jasper aquifer to the Burkeville confining unit and Evangeline aquifer in the subcrop.

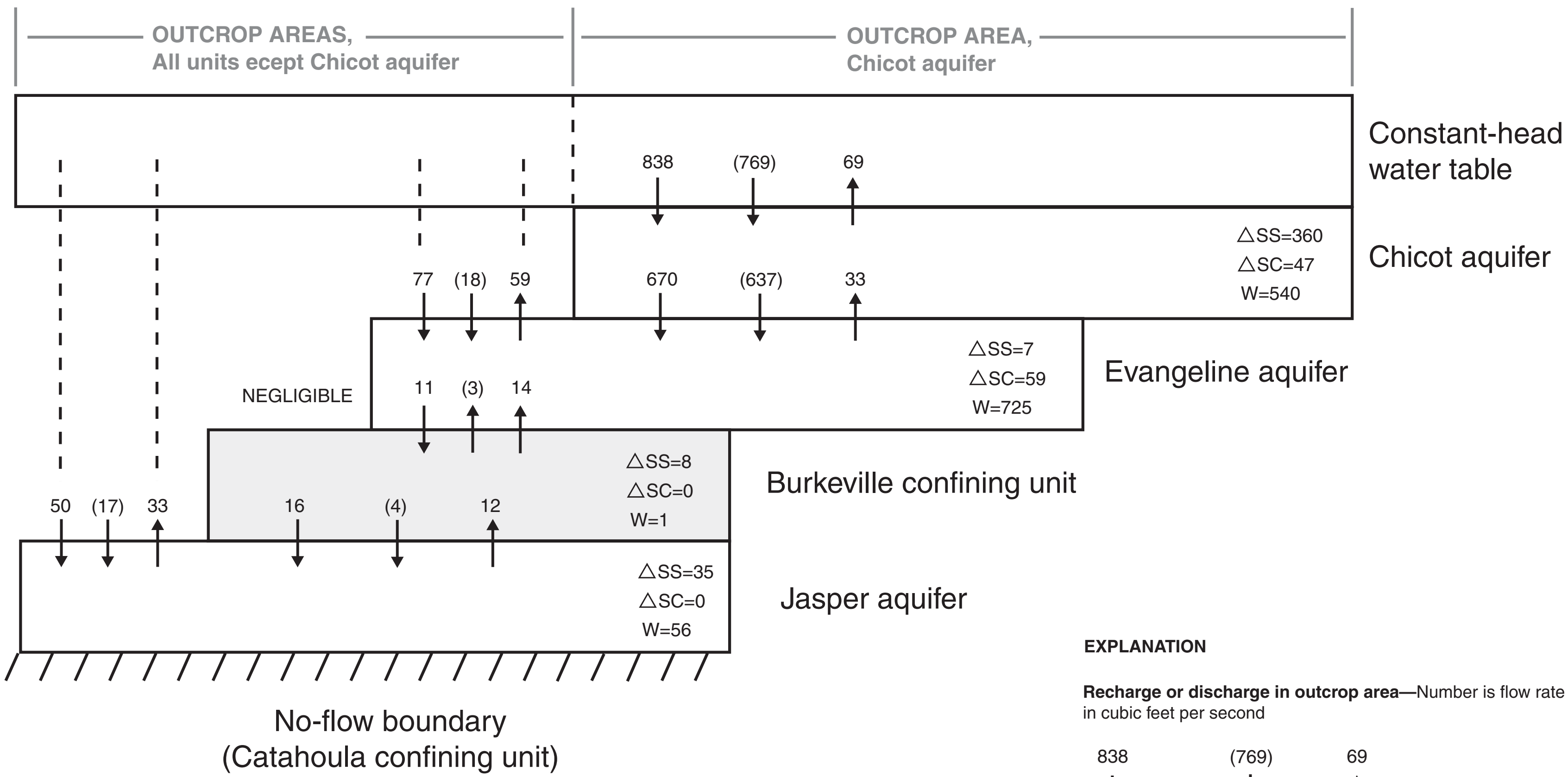
The predevelopment recharge rate over the GAM outcrop area is equivalent to 0.17 in/yr. For comparison, the simulated predevelopment recharge rate over the outcrop area of the Gulf Coast aquifer system in Texas (Ryder and Ardis, 2002) is 0.12 in/yr.

The most notable differences between the simulated water-budget components of the predevelopment and postdevelopment aquifer systems are the relatively large increases in recharge, decreases in natural discharge, and downward leakage from the Chicot to the Evangeline aquifer in 1977 and 2000 compared with predevelopment. Predevelopment recharge of about 307 ft^3/s increased to 757 ft^3/s (to help sustain withdrawals of 1,670 ft^3/s) for 1977 and increased to 965 ft^3/s (to help sustain withdrawals of 1,322 ft^3/s) for 2000. Predevelopment discharge of about 307 ft^3/s decreased to 169 ft^3/s for 1977 and 161 ft^3/s for 2000. One notable similarity among all three water budgets (figs. 72–74) is that major fractions of the simulated recharge in the Evangeline and Jasper aquifers discharge naturally in the respective outcrops rather than flowing to deeper, downdip parts of the aquifers.

Simulated and Measured Land-Surface Subsidence

Simulated and measured land-surface subsidence from predevelopment to near present day are shown in figures 75 and 76. The match in the Harris-Galveston-Fort Bend County area, where compaction of subsurface material and thus subsidence has been monitored continuously since the 1970s, is close. As much as 10 ft of subsidence has occurred in southeastern Harris County near the northern end of Galveston Bay. A larger geographic area encompassing the maximum land-surface-subsidence area and much of central to southeastern Harris County has subsided at least 6 ft.

Away from the Harris-Galveston-Fort Bend County area, simulated subsidence of as much as 3 ft is shown in the Evadale-Beaumont withdrawal area in southwestern Jasper County (fig. 75). No measured subsidence in that area is shown for comparison because no recent (near 2000) measurements are available. Wesselman (1967, p. 58) hypothesizes on the basis of differential subsidence and head decline measured at Evadale between 1955 and 1963 that a ratio of 0.912 ft of subsidence per 100 ft of head decline might be applicable in the area. Applying that ratio to simulated maximum head decline of about 300 ft in the Evangeline aquifer (difference between simulated Evangeline aquifer predevelopment head [fig. 63] and 2000 head



EXPLANATION

Recharge or discharge in outcrop area—Number is flow rate in cubic feet per second

838 ↓ RECHARGE (769) ↓ NET 69 ↑ NATURAL DISCHARGE

Leakage through bottom of hydrogeologic unit—Number is flow rate in cubic feet per second

670 ↓ DOWNWARD (637) ↓ NET 33 ↑ UPWARD

△SS Net rate of release of water from sand storage, in cubic feet per second

△SC Net rate of release of water from inelastic clay compaction, in cubic feet per second

W Withdrawal rate, in cubic feet per second

WATER BALANCE[1]

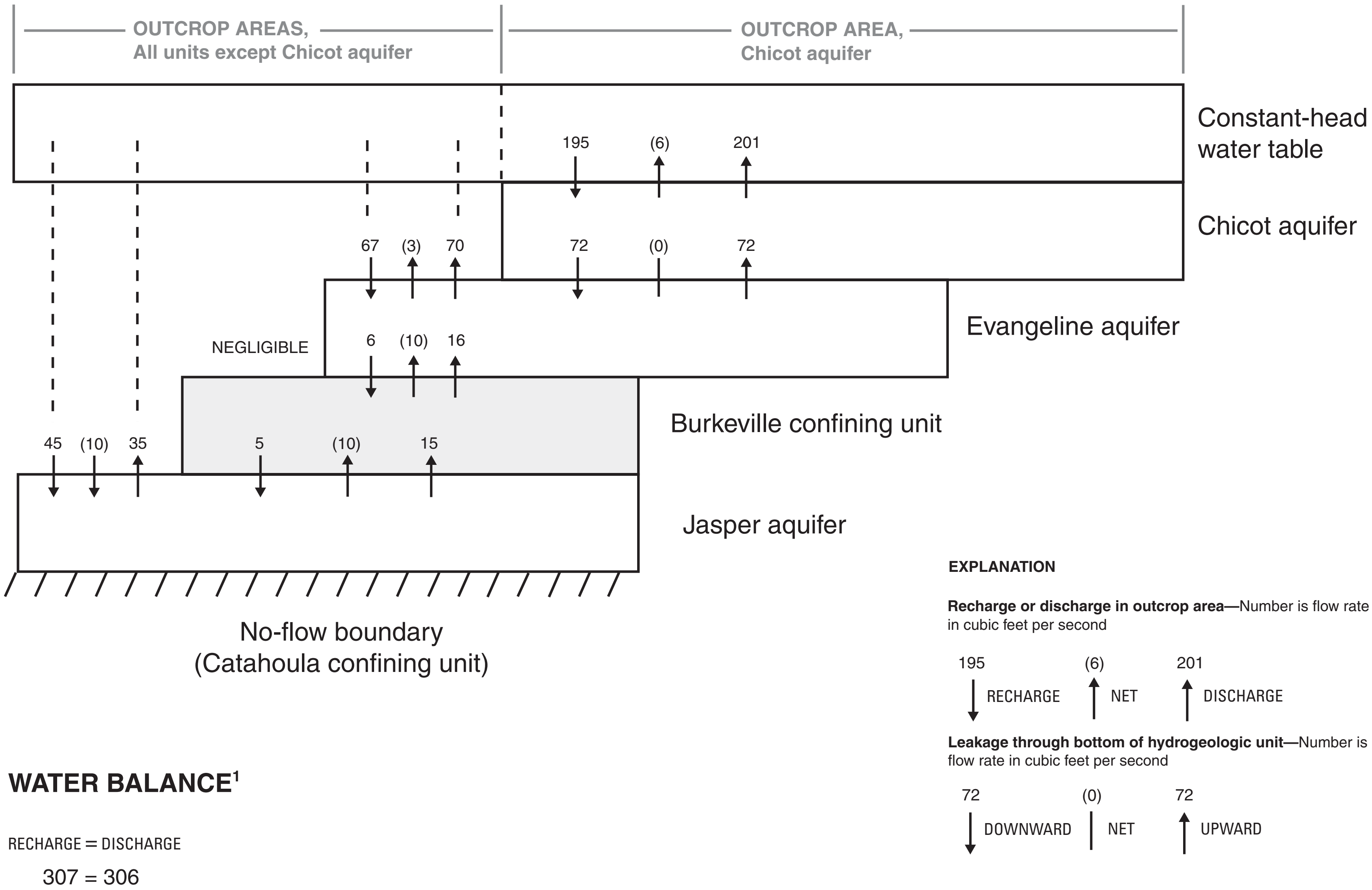
RECHARGE + △SS + △SC = NATURAL DISCHARGE + W

965 + 410 + 106 = 161 + 1,322

1,481 = 1,483

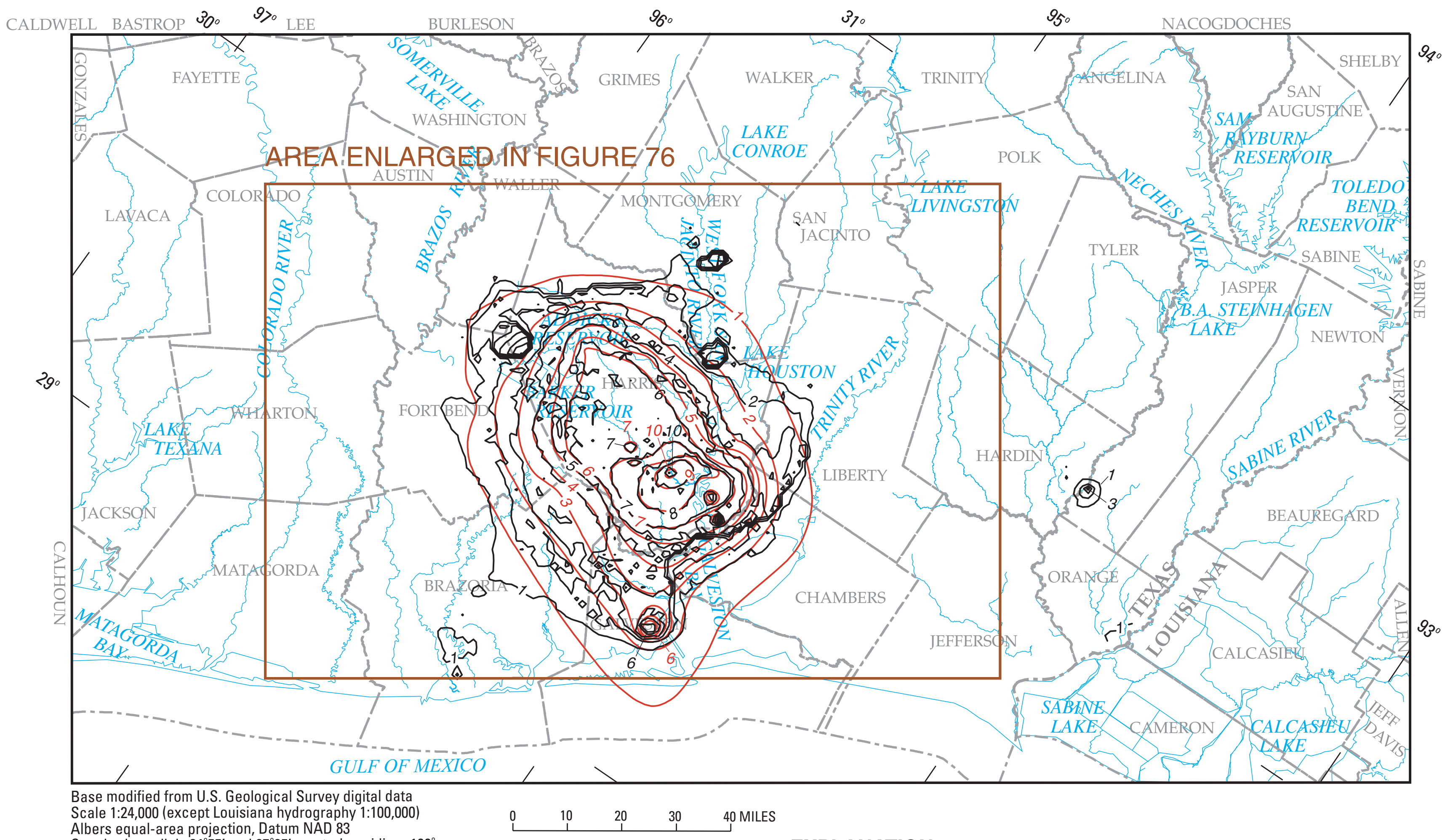
[1] Inexact because of rounding.

Figure 73. Simulated 2000 water-budget components of the hydrogeologic units in the Ground-Water Availability Model area.



[1] Inexact because of rounding.

Figure 74. Simulated predevelopment water-budget components of the hydrogeologic units in the Ground-Water Availability Model area.



Base modified from U.S. Geological Survey digital data
Scale 1:24,000 (except Louisiana hydrography 1:100,000)
Albers equal-area projection, Datum NAD 83
Standard parallels 34°55′ and 27°25′, central meridian -100°

EXPLANATION

Land-surface subsidence, in feet

—5— 1891–2000 Simulated contour—Interval 1 foot

—5— 1906–95 Measured contour—Interval 1 foot

Figure 75. Simulated and measured 2000 land-surface subsidence in the Ground-Water Availability Model area.

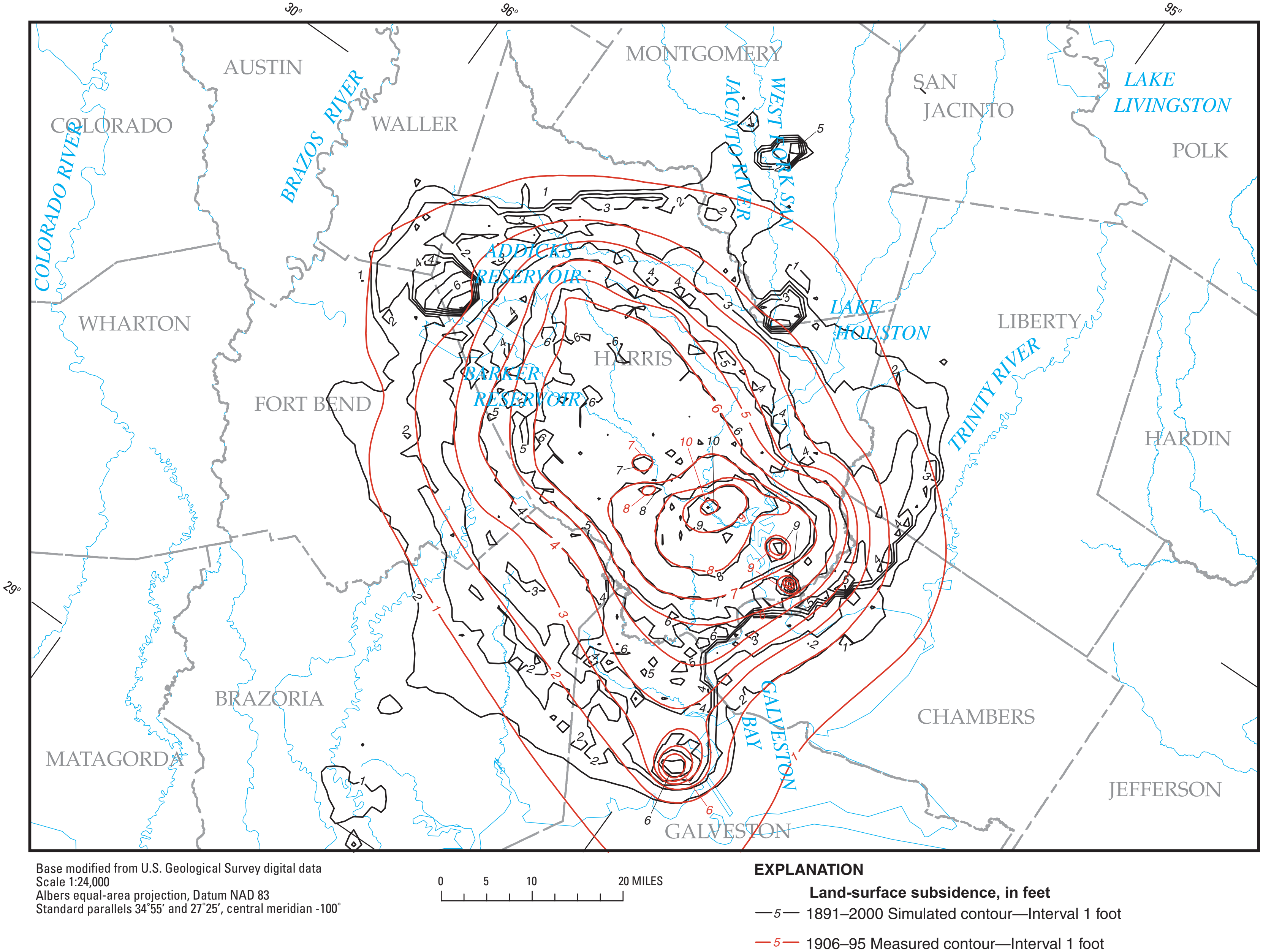


Figure 76. Simulated and measured 2000 land-surface subsidence in the Houston area of the Ground-Water Availability Model area.

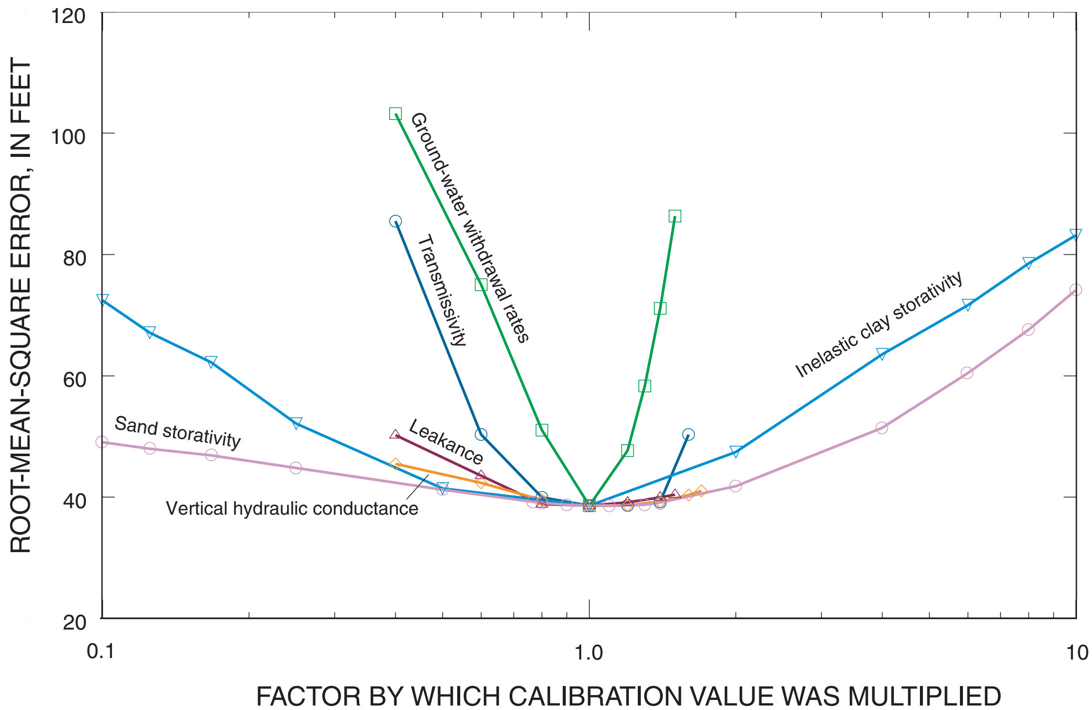


Figure 77. Sensitivity of calibrated-model responses to changes in selected model input data.

[fig. 56]) yields 2.74 ft of subsidence, a close match to the simulated 3 ft of subsidence in the area.

No simulated subsidence is shown in the coastal irrigation area centered in southern Wharton County. As in the Evadale-Beaumont withdrawal area, no recent subsidence measurements are available for the rice irrigation area. However, Carr and others (1985, fig. 37) show measured subsidence of as much as 1.5 ft in two small areas near the confluence of Wharton, Jackson, and Matagorda Counties.

Sensitivity Analysis

The sensitivity of calibrated-model responses to changes in input data (the aquifer properties that control flow, recharge, discharge, and storage, plus withdrawals) was evaluated to help assess model reliability. The values of selected model input data were varied individually over ranges that reflect plausible uncertainty (potential lack of accuracy in the estimated values) in a series of simulations to show the effects of the uncertainty on simulated heads. As sand storativity and inelastic clay storativity could range over several orders of magnitude, those properties were varied over a larger range than other properties. The effects of those changes on simulated 2000 heads were measured in terms of increases in RMS error (fig. 77).

This analysis has implications if the GAM is used for prediction of aquifer responses to future stresses. For example, the plots in figure 77 show that accurate estimates of transmissivity and withdrawals are substantially more important to reliable predictions of heads than accurate estimates of leakance or vertical hydraulic conductance.

Model Limitations

Several factors limit, or detract from, the ability of the GAM to reliably predict aquifer responses to future conditions. The GAM, like any model, is a simplification of the real aquifer system it simulates. As Brooks and others (1994) explain, simplification is necessary not only to make the problem tractable, but also because the structure, properties, and boundaries of and stresses on the aquifer system can never be fully known. Simplifications involve assumptions about the real system and the way it functions. Knowledge (or lack of knowledge) of the system is reflected in the quality and quantity of input data. The scale of the model, which is associated with the necessity to discretize a continuous system in space, also affects the ability of a model to produce reliable results.

Assumptions

A basic assumption is that the hydrogeologic units of the aquifer system can be adequately represented by four discrete layers, a simplification because in the real system, the change from one aquifer to another with depth likely is transitional rather than abrupt. Other assumptions pertain to the boundary conditions. The conceptualization of the downdip boundaries of each hydrogeologic unit as the downdip limit of freshwater flow probably is realistic—salinity increases and flow becomes increasingly sluggish with distance downdip in each unit; however, the simplifying assumption that the downdip limit of freshwater flow in each unit is a sharp interface across which no flow occurs, the position of which is known and static over time, is more tenuous, as was discussed in the section on lateral boundaries. The assumption of the southwestern and northeastern aquifer-system boundaries as no-flow, coincident with the Lavaca and Sabine Rivers, respectively, is not entirely realistic. Although those streams likely represent effective ground-water-flow divides in the shallow subsurface, the vertical extent of their influence on ground-water flow is unknown. However, those lateral boundaries are far enough from areas of major withdrawals so that they likely have negligible influence on the simulated response to withdrawals. The base of the aquifer system (base of the Jasper aquifer) is assumed to be a no-flow boundary, although in the real system, a relatively small amount of water probably flows between the Jasper aquifer and the underlying Catahoula confining unit.

Another assumption is that in areas of large withdrawals and substantial declines in the potentiometric surface of an aquifer, the overlying water table has not declined in response to increased downward gradients; water-table heads are held constant during simulations. If this assumption is not valid, then more recharge than actually occurs in the real system could be simulated in such areas, which also could result in simulated heads higher than actual heads. Although the validity of this assumption has not been studied, the authors believe annual rainfall is sufficient to keep any long-term water-table declines to a minimum.

As noted in the section on land-surface subsidence and storage in clays, assuming a constant-head water table also means constant geostatic pressure, which in turn makes changes in effective stress a function only of changes in head. If the assumption of a constant water table were not valid and the water table in the real system were to decline appreciably, then the model could overestimate effective stress and thus overestimate compaction (subsidence).

Also pertaining to the simulation of land-surface subsidence, the assumption is made that head changes within a model time step in the aquifer sands are the same as those in the interbedded clays; in other words, head changes in the clays do not lag those in the sands. If simulated time steps are too short to allow for dissipation of all excess pore pressure in the clays of the real system, then the amount of water released by the clays in the simulated system will be unrealistically large for the time step. Leake and Prudic (1991, p. 7) provide an equation for the upper limit on the time required for excess pore pressure in the real system to dissipate on the basis of interbedded clay properties, which can be compared to the length of model time steps. Computations for the interbedded clays in the aquifer system indicate excess pore pressure will dissipate in about 300 days. Thus the 1-year model time steps that were applied for all of the transient period except 1980, 1982, and 1988 appear to be adequate, but the 1-month time steps during those 3 years probably are not; which implies that the simulated amount of water released by the clays for each of those 3 years probably is greater than the actual amount.

Input Data

Associated with each of the input datasets is a level of uncertainty and a degree of bias, neither of which is quantitatively known. The uncertainty arises from the fact that point measurements or estimates of the input data represent regions around the points. The bias originates from the facts that some properties are better known than others and individual properties are better known in some areas than others (data points commonly are concentrated in some areas, sparse in others). The result is that the optimum (but non-unique) distributions of input data arrived at through calibration, or history matching, are distributions of effective properties, not actual properties. That is, the set of property distributions for the calibrated model is one of potentially many plausible sets that would allow simulated heads, subsidence, and water-budget components to reasonably match those of the real system under selected conditions.

In all likelihood, the property distributions reflect the order of magnitude of the real-system properties, but not the true distributions of the real-system properties. For example, the simulated distributions of transmissivity of the Chicot, Evangeline, and Jasper aquifers (figs. 34, 35, 37), while generally of the correct orders of magnitude, show larger values and generally more "definition" in areas coincident with large withdrawals. The distributions reflect the availability of more historical information for those areas and thus more attention to those areas during calibration. It is likely that if comparable ground-water development, subsurface information, head data, and calibration attention were focused on the system in other parts of the GAM area, the distributions of transmissivity in those areas would reflect that situation and be different from the distributions of figures 34, 35, and 37.

What can be said about the distributions of aquifer-system properties after calibration is that, collectively, they are one set of probably many sets of input data that allows the model to reasonably reproduce selected historical heads, subsidence, and flows. This implies that the reliability of the model for predictive simulation is uncertain.

Scale of Application

The GAM is a regional-scale model, and as such it is intended for regional-scale rather than local-scale analyses. Discretization of the GAM area into 1-mi^2 grid blocks in which aquifer properties and conditions are assumed to be averages over the area of each grid block precludes site-specific analyses. For example, the simulated head in a grid block encompassing one or more pumping wells will represent an average head in the actual grid-block area rather than the head at or near the pumping well, which is much lower. An implication of simulated areal average heads is that, for calibration, comparison of simulated heads to measured heads might not always be a comparison of like data. Although care is taken to ensure that static (non-pumping) water-level data are collected, undoubtedly some measured heads are influenced by nearby pumping or for other reasons are not representative of an average head in the grid-block area.

Another scale-related issue—the "scale problem" as defined by Johnston (1999)—was described in the "Ground-Water-Flow Conditions, Recharge, and Discharge" section. Because flow that enters and exits the real system within the area encompassed by a single grid block cannot be simulated except by superposition of sources or sinks, which would be impractical over a regional area, the model does not simulate total recharge (and thus total [real-system] ground-water flow). The fraction of total flow simulated is unknown, but the fraction of total flow simulated decreases as the grid-block size increases. What this implies is that any simulated components of flow not explicitly specified (for example, natural recharge and discharge) will be less than their real-system counterparts. Explicitly specified components (for example, withdrawals) are based on measured or estimated real-system data and therefore will more closely match real-system magnitudes.

Summary

The northern part of the Gulf Coast aquifer system in Texas, which includes the Chicot, Evangeline, and Jasper aquifers, supplies most of the water used for industrial, municipal, agricultural, and commercial purposes for an approximately 25,000-mi^2 area that includes the Beaumont and Houston metropolitan areas. The area has an abundant amount of potable ground water, but withdrawals of large quantities of ground water have resulted in potentiometric-surface declines in the Chicot, Evangeline, and Jasper aquifers and land-surface subsidence from depressurization and compaction of clay layers interbedded in the aquifer sediments. The study that generated this report, done in cooperation with the Texas Water Development Board (TWDB) and the Harris-Galveston Coastal Subsidence District as a part of the TWDB Ground-Water Availability Modeling (or Model) (GAM) program, was designed to develop and test a ground-water-flow model of the northern part of the Gulf Coast aquifer system in Texas that water-resource managers can use as a tool to address future ground-water-availability issues.

The GAM area, which encompasses the northern part of the Gulf Coast aquifer system in Texas, is a gently sloping coastal plain that includes all or parts of 38 counties in Texas, all or parts of six regional water-planning groups, two subsidence districts, and parts of four natural subregions. Land-surface altitudes are topographically highest along the northwestern boundary. The major river basins in the GAM area are the Brazos, Colorado, Lavaca, Sabine, San Jacinto, and Trinity. Average annual rainfall over the GAM area is about 48 in.

In a generalized conceptual model of the aquifer system, water enters the ground-water-flow system in topographically high outcrops of the hydrogeologic units in the northwestern part of the system. Much of the water that infiltrates to the saturated zone flows relatively short distances through shallow zones and discharges to streams; the remainder of the water flows to intermediate and deep zones of the system southeastward of the outcrop areas where it is discharged by wells and by upward leakage in topographically low areas near the coast. Near the coast and at depth, saline water is present, which causes freshwater not captured by wells to be redirected upward as diffuse leakage and ultimately discharged to coastal water bodies.

From land surface downward, the Chicot aquifer, the Evangeline aquifer, the Burkeville confining unit, the Jasper aquifer, and the Catahoula confining unit are the hydrogeologic units of the Gulf Coast aquifer system. (The Catahoula confining unit is assumed to be a no-flow base of the system for simulation.) From several previous studies, transmissivity of the Chicot aquifer ranges from about 3,000 to 68,000 ft^2/d, the Evangeline aquifer about 2,000 to 15,000 ft^2/d, and the Jasper aquifer about 1,000 to 35,000 ft^2/d. Storativity of the Chicot aquifer ranges from about 4×10^{-4} to 0.1, the Evangeline aquifer about 5×10^{-4} to 0.1; the few available Jasper aquifer storativities are within those ranges.

The uppermost parts of the aquifer system, which include outcrop areas, are under water-table conditions. As depth increases in the aquifer system and interbedded sand and clay accumulate, water-table conditions evolve into confined conditions. Thus the lowermost parts of the aquifer system are under confined conditions. The middle parts of the aquifer system therefore are under semiconfined conditions.

Recharge rates from previous studies range from a fraction of an inch per year to as much as 7 in/yr, depending on the type of study (field or simulation) and whether the term "recharge" refers to total recharge (all of the precipitation that infiltrates the subsurface) or some fraction of total recharge (that which does not discharge rapidly to streams but flows to deeper parts of the system). Simulation studies tend to yield smaller estimates of recharge because some of the flow that enters and exits the physical system within the area encompassed by a single model grid block cannot be simulated except by superimposing sources or sinks.

Three principal areas of concentrated ground-water withdrawals from the system are in the GAM area: the largest, in terms of water withdrawn, is Harris and Galveston Counties (the Houston area). Withdrawals began there at the end of the 19th century. The long-term rate of increase in withdrawals was changed in the late 1970s by regulation of withdrawals by the newly created Harris-Galveston Coastal Subsidence District. Withdrawals of about 450 Mgal/d in 1976 increased moderately to about 463 Mgal/d by 1996. The second principal area of withdrawals is the coastal irrigation area centered in Wharton and

Jackson Counties. Most of the irrigation withdrawals are from the Chicot aquifer for rice. By the late 1960s, irrigation withdrawals in Wharton County, which historically account for about 70 to 80 percent of the irrigation total for the area, had reached 172 Mgal/d. The rate of irrigation withdrawal in Wharton County decreased during most of the ensuing years but increased to about 183 Mgal/d in 2000. The third principal area of withdrawals is the Evadale-Beaumont area. Industrial withdrawals are associated with wood-pulp processing at Evadale in southwestern Jasper County, and public-supply withdrawals are from the Beaumont well field in southeastern Hardin County. The combined Evadale-Beaumont withdrawals from the Chicot and Evangeline aquifers for 1977 were about 24 Mgal/d; by 2000, the rate had increased to about 44 Mgal/d.

Before appreciable ground-water withdrawals from the system in the GAM area, the potentiometric surfaces in the confined parts of the aquifers were higher than land surface in places. Ground-water development has caused substantial (as much as 350 ft) declines of the potentiometric surfaces of the aquifers (and subsequent land-surface subsidence) primarily in the Houston area. Although appreciable amounts of water have been withdrawn from the Chicot aquifer in the coastal irrigation area for decades, relatively little long-term drawdown (tens of feet) has occurred there. Rice-irrigation return flow and withdrawals from relatively shallow zones (under water-table conditions) that are readily recharged probably have helped to lessen long-term water-level declines in the area. No recent synoptic water-level measurements are available in the Evadale-Beaumont area, but one estimate of drawdown in the Evangeline aquifer centered at Evadale in 1982 was 150 to 200 ft.

Head declines associated with withdrawals in the Chicot and Evangeline aquifers, predominantly in the Houston area, have caused compaction of interbedded clays, which has resulted in substantial land-surface subsidence. More than 10 ft of land-surface subsidence has been documented in the Baytown and Houston Ship Channel area in southwestern Harris County. Subsidence of smaller but still destructive magnitudes has occurred in places throughout most of Harris County and to a lesser extent in parts of Galveston and Fort Bend Counties.

The U.S. Geological Survey MODFLOW finite-difference model was used to simulate ground-water flow and land-surface subsidence. The model comprises four layers, one for each of the hydrogeologic units of the aquifer system except the Catahoula confining unit, the no-flow base of the system. Each layer consists of 137 rows and 245 columns of uniformly spaced grid blocks, each block representing 1 mi^2.

The northwestern boundaries of the three aquifers and the Burkeville confining unit are the northwestern extent of the updip outcrop sediments for each unit. The downdip limit of freshwater (dissolved solids concentration of 10,000 mg/L) was chosen as the southeastern boundary of flow in each hydrogeologic unit. The southwestern-northeastern lateral boundaries of the hydrogeologic units were selected to coincide with ground-water-flow divides associated with major streams—the Lavaca River to the southwest and the Sabine River to the northeast.

The MODFLOW general-head boundary package was used to simulate recharge and discharge in the outcrops of the Chicot, Evangeline, and Jasper aquifers and the Burkeville confining unit. This package allows the water table of an aquifer system to function as a head-dependent flux (flow per unit area) boundary. Flow between streams and the aquifer system was not explicitly simulated by imposing sinks along streams in the model. The rationale for this decision is that the general-head boundary package, assuming the model is adequately calibrated, would account for stream discharge to the level of accuracy that such discharge is known.

The initial values of hydraulic properties associated with ground-water flow were selected on the basis of findings of numerous previous studies and hydrologic judgment. Simulations were made under transient conditions from 1891 through 2000 for 68 withdrawal (stress) periods of variable, but mostly annual, length. Historical ground-water-withdrawal data—municipal, manufacturing, mining, power generation, livestock, irrigation, and county-other—were compiled from numerous sources and distributed to the appropriate model layers and grid blocks by various methods.

Simulation of land-surface subsidence (actually, compaction of clays) and release of water from storage in the clays of the Chicot and Evangeline aquifers was accomplished using the Interbed-Storage Package designed for use with the MODFLOW model. Subsidence and compaction of clays in the Jasper aquifer and the Burkeville confining unit were not simulated because the sediments of those units are more consolidated relative to the sediments of the Chicot and Evangeline aquifers and less head decline has occurred in those units.

The GAM was calibrated by trial-and-error adjustment of selected model input data (the aquifer properties that control water flow, recharge, discharge, and storage) in a series of transient simulations until the model output (potentiometric surfaces, land-surface subsidence, selected water-budget components) reasonably reproduced field measured (or estimated) aquifer responses. The calibration objective was to minimize the differences between simulated and measured aquifer responses. Model calibration comprised four elements: The first was qualitative comparison of simulated and measured potentiometric surfaces in the aquifers for 1977 and 2000; and quantitative comparison of simulated and measured potentiometric surfaces by computation and areal distribution of the RMS error between simulated and measured heads. The second calibration element was comparison of simulated and measured hydrographs from wells in the aquifers in the Houston area, the coastal irrigation area, and selected counties away from those areas of withdrawal. The third calibration element was comparison of simulated water-budget components—primarily recharge and discharge—to estimates of physically reasonable ranges of actual water-budget components; and comparison of simulated distributions of recharge and discharge in the outcrops of aquifers to estimates of physically reasonable distributions based on knowledge of the hydrology of the aquifer system. The fourth calibration element was comparison of simulated land-surface subsidence from predevelopment to

2000 to measured land-surface subsidence from 1906 through 1995.

For the Chicot aquifer, transmissivities after calibration range from negligible to about 77,000 ft^2/d. For the Evangeline aquifer, transmissivities range from negligible to about 43,000 ft^2/d. Transmissivities near the maximums for both aquifers occur in only a few grid blocks. Transmissivities of the Burkeville confining unit (unadjusted from initial values during calibration) are very small (maximum about 8 ft^2/d). For the Jasper aquifer, transmissivities range from negligible to about 14,500 ft^2/d.

Storativities of the Chicot and Evangeline aquifers (1×10^{-4} to 0.2 and 4×10^{-5} to 0.2, respectively) reflect aquifer conditions from confined to semiconfined to water table. Chicot aquifer storativities generally are largest in the updip, outcrop areas where water-table conditions prevail. Storativities of the Burkeville confining unit and the Jasper aquifer (1×10^{-5} to 5×10^{-2} and 2×10^{-5} to 0.2, respectively) also are generally largest in the updip, outcrop areas where water-table conditions prevail.

The simulated potentiometric surfaces of the Chicot, Evangeline, and Jasper aquifers for 1977 show general agreement with the measured potentiometric surfaces (or with measured point head data in areas where data are sparse). The RMS errors for the aquifer potentiometric surfaces, which reflect the average difference between 1977 simulated and measured heads, were about 34 ft for the Chicot aquifer, about 43 ft for the Evangeline aquifer, and about 47 ft for the Jasper aquifer. The RMS errors are about 7, 8, and 17 percent, respectively, of the total range in measured heads for the respective aquifers.

The simulated potentiometric surfaces of the Chicot, Evangeline, and Jasper aquifers for 2000 also show general agreement with the measured potentiometric surfaces (or with measured point head data in areas where data are sparse). The simulated and measured 2000 Chicot and Evangeline potentiometric surfaces, compared with those for 1977, show substantial shifts to the northwest in the major cones of depression in the Houston area, which reflect shifts northwestward of the centers of withdrawals during 1977–2000. The measured 2000 Chicot aquifer potentiometric surface also shows about 100 ft of recovery in the major cone of depression, which is consistent with the overall reduction in withdrawals from the system during 1977–2000. In the Evadale-Beaumont area, the simulated 2000 cone of depression in the Evangeline aquifer is larger and about 150 ft deeper than the cone of 1977. Simulated withdrawals in the area increased about 85 percent between 1977 and 2000. The RMS errors for the three aquifer potentiometric surfaces for 2000 were about 31 ft for the Chicot aquifer, about 40 ft for the Evangeline aquifer, and about 34 ft for the Jasper aquifer. The RMS errors are about 8, 6, and 11 percent, respectively, of the total range in measured heads for the respective aquifers.

Simulated and measured hydrographs for the three aquifers in the Houston area (10 hydrograph pairs) match closely relative to the ranges of change; those from the Chicot aquifer in the coastal irrigation area (three hydrograph pairs) match less closely; and those for the aquifers away from the Houston and coastal irrigation areas (eight hydrograph pairs, including two in the Evadale-Beaumont withdrawal area) match with varying degrees of closeness. For hydrographs in which the match between simulated and measured heads is less close than others, the trends in simulated and measured heads generally are similar.

For calibrated 1977 conditions, simulated net recharge is 555 ft^3/s (0.40 in/yr) in the Chicot aquifer outcrop, 19 ft^3/s (0.12 in/yr) in the Evangeline aquifer outcrop, negligible in the Burkeville confining unit outcrop, and 14 ft^3/s (0.06 in/yr) in the Jasper aquifer outcrop. In terms of a water balance (within 5 ft^3/s) for the entire system in 1977, 757 ft^3/s of recharge plus 1,082 ft^3/s from depletion of sand storage (742 ft^3/s) and inelastic compaction of clays (340 ft^3/s) is offset by 169 ft^3/s of natural discharge and 1,670 ft^3/s (1,080 Mgal/d) of withdrawals. Thus in 1977, net recharge supplied about 35 percent of withdrawals, depletion of sand storage about 45 percent, and inelastic compaction of clays about 20 percent.

For calibrated 2000 conditions, simulated net recharge is 769 ft^3/s (0.55 in/yr) in the Chicot aquifer outcrop, 18 ft^3/s (0.11 in/yr) in the Evangeline aquifer outcrop, negligible in the Burkeville confining unit outcrop, and 17 ft^3/s (0.07 in/yr) in the Jasper aquifer outcrop. In terms of a water balance (within 5 ft^3/s) for the entire system in 2000, 965 ft^3/s of recharge plus 516 ft^3/s from depletion of sand storage (410 ft^3/s) and inelastic compaction of clays (106 ft^3/s) is offset by 161 ft^3/s of natural discharge and 1,322 ft^3/s (854 Mgal/d) of withdrawals. Thus in 2000, net recharge supplied 61 percent of withdrawals, depletion of sand storage 31 percent, and inelastic compaction of clays 8 percent.

The most notable differences between the simulated water-budget components of 1977 and 2000, besides the fact withdrawals were about 21 percent less in 2000, are the increase in the percentage of withdrawals supplied by recharge and the decrease in the percentage of water supplied by depletion of storage and inelastic compaction of clays between 1977 and 2000. The simulated recharge rates for the GAM for 1977 and 2000 appear to be generally comparable to estimates of recharge rates from previous studies involving all or parts of the Gulf Coast aquifer system in Texas.

The match between simulated and measured land-surface subsidence from predevelopment to near present day in the Harris-Galveston-Fort Bend County area, where compaction of subsurface material and thus subsidence has been monitored continuously since the 1970s, is close. As much as 10 ft of subsidence has occurred in southeastern Harris County near the northern end of Galveston Bay. A larger geographic area encompassing the maximum land-surface-subsidence area and much of central to southeastern Harris County has subsided at least 6 ft.

Away from the Harris-Galveston-Fort Bend County area, subsidence of as much as 3 ft was simulated in the Evadale-Beaumont withdrawal area in southwestern Jasper County. No subsidence was simulated in the coastal irrigation area centered in southern Wharton County. No recent (near 2000) subsidence measurements are available for either area, although small amounts of subsidence (less than 2 ft) have been documented historically in both areas.

Several factors limit, or detract from, the ability of the GAM to reliably predict aquifer responses to future conditions. For example, associated with each of the input datasets is a level of uncertainty and a degree of bias, neither of which is quantitatively known. The uncertainty arises from the fact that point measurements or estimates of the input data represent regions around the points. The bias originates from the facts that some properties are better known than others and individual properties are better known in some areas than others. The result is that the optimum (but non-unique) distributions of input data arrived at through calibration, or history matching, are distributions of effective properties, not actual properties. In all likelihood, the property distributions reflect the order of magnitude of the real-system properties, but not the true distributions of the real-system properties. What can be said about the distributions of aquifer-system properties after calibration is that, collectively, they are one set of probably many sets of input data that allows the model to reasonably reproduce selected historical heads, subsidence, and flows. This implies that the reliability of the model for predictive simulation is uncertain.

The GAM is a regional-scale model, and as such it is intended for regional-scale rather than local-scale analyses. Discretization of the GAM area into 1-mi^2 grid blocks in which aquifer properties and conditions are assumed to be averages over the area of each grid block precludes site-specific analyses. Discretization detracts in another way as well: Because flow that enters and exits the real system within the area encompassed by a single grid block cannot be simulated except by superposition of sources or sinks, the model does not simulate total recharge (and thus total [real-system] ground-water flow). What this implies is that any simulated components of flow not explicitly specified (for example, natural recharge and discharge) will be less than their real-system counterparts.

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