# The Effect of River Valleys and the Upper Cretaceous Aquitard on Regional Flow in the Dakota Aquifer in the Central Great Plains of Kansas and Southeastern Colorado

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## Abstract

In his reports on the regional hydrogeology of the central Great Plains, in particular southeastern Colorado and southwestern and central Kansas, Darton considered the Dakota aquifer to be a classic example of an artesian system. Computer simulations of the flow system in this study, however, suggest that the Dakota is not a regional artesian aquifer in the classic sense. Sensitivity analysis of a steady-state vertical profile flow model demonstrates that the flow system in the upper Dakota in western Kansas is heavily influenced by the Upper Cretaceous aquitard, the Arkansas River in southeastern Colorado, and rivers in central Kansas, such as the Saline, that have eroded through the aquitard and into the Dakota to the west of the main outcrop area of the aquifer. The model shows that local flow systems and the vertical hydraulic conductivity of the Upper Cretaceous aquitard heavily influence the water budget and the flow patterns. The aquitard restricts recharge from the overlying water table to underlying aquifers in western Kansas because of its considerable thickness and low vertical hydraulic conductivity. The Arkansas River intercepts ground-water flow moving toward western Kansas from recharge areas south of the river and further isolates the upper Dakota from sources of freshwater recharge. In central Kansas, the Saline River has reduced the distance between confined portions of the aquifer and its discharge area. In essence, this has improved the hydraulic connection between the confined aquifer and its discharge area, thus helping to generate subhydrostatic conditions in the upper Dakota upgradient of the river.

The Dakota aquifer and its equivalents extend over much of the Great Plains of North America and are known as dependable sources of freshwater in much of this region. In the Arkansas River valley of southeastern Colorado and adjacent parts of southwestern Kansas, the Dakota was probably one of the first sources of water used by the early settlers and railroads because of its flowing well conditions and its shallow depth. Figure 1 shows the extent of the study area in Kansas and Colorado that is the subject of this paper.

Darton (1905, 1906) described the results of the earliest regional reconnaissance investigations into the hydrogeology of the Dakota aquifer in southeastern Colorado and adjacent parts of western and central Kansas. Darton reported numerous flowing wells in the Arkansas River valley and its tributaries in southeastern Colorado and in parts of central Kansas. He demonstrated that water enters the Dakota where it crops out at the surface south of the river, flows northeastward, and eventually discharges in central Kansas, where again the aquifer crops out at the surface (fig. 1). Using Chamberlin's (1885) concept of artesian aquifers, Darton believed that ground-water flow was controlled mostly by the head difference between recharge and discharge areas. Darton reasoned that the flowing wells in the Arkansas River valley and elsewhere in central Kansas could be accounted for only by the elevated recharge areas south of the river and the maintenance of artesian pressure in the Dakota from recharge to discharge areas by overlying Upper Cretaceous shales, referred to here as the Upper Cretaceous aquitard. He described the Dakota as a classic example of an artesian aquifer.

More recent, regional hydrogeologic investigations have not supported Darton's conceptualization of flow in the Dakota. In a review of the literature, Helgeson et al. (1982) questioned the applicability of Chamberlin's (1885) concept to the Dakota on the basis of aquifer geometry and lateral hydraulic continuity. In the Denver basin and adjacent areas of eastern Colorado and western and central Kansas, Belitz (1985), Belitz and Bredehoeft (1988), and Helgeson et al. (1994) reported that some heads in the Dakota and the deeper aquifers are more than 2,500 ft (760 m) lower than the elevation of the overlying water table. An aquifer is usually considered to be in good hydraulic communication with the overlying water table if there are only small head differences between them. Thus the Dakota, with its subhydrostatic heads—heads that are significantly lower than those of the overlying water table—is essentially isolated from the overlying water table. In the Denver basin the Dakota is overlain by the Upper Cretaceous aquitard, which is as much as 10,000 ft (3,050 m) thick. From regional flow models, Belitz (1985), Belitz and Bredehoeft (1988), and Helgeson et al. (1994) concluded that thick aquitards consisting of Upper Cretaceous shale and chalk severely restrict recharge to deeper parts of the flow system. Because of this, it appears that the head in the Dakota and the deeper aquifers is more responsive to the head of discharge areas to the east than to the head of the overlying water table.

In southeastern Colorado and in most of western Kansas, the Dakota is within 1,000 ft (300 m) of land surface and the head difference between the water table and the Dakota is less than 500 ft (150 m) (Helgeson et al., 1994). Heads in the deeper aquifers below the Pennsylvanian are more than 2,000–3,000 ft (600–900 m) lower than heads in the Dakota aquifer (Belitz and Bredehoeft, 1988; Jorgensen et al., 1993). This suggests (1) that thick aquitards below the Dakota continue to restrict recharge to these deeper aquifers eastward of the Denver basin and (2) that the hydrogeology of the Dakota aquifer changes significantly between the Denver basin and the study area to the east because of the thinning and removal of the Upper Cretaceous aquitard and exposure of the Dakota aquifer to the south and east.

## Revised Conceptualization of the Flow System

The Dakota is an important component of the regional flow system because it is hydraulically connected to all the major overlying and underlying aquifer systems, including the overlying water table where the Dakota crops out at the surface in southeastern Colorado and central Kansas. In this research, it is hypothesized that steady-state ground-water flow in the upper part of the regional system, including the Dakota aquifer, is influenced primarily by (1) the Upper Cretaceous aquitard, (2) the Arkansas River, and (3) the drainages in central Kansas that have cut down through the aquitard to the west of the main outcrop belt. The Upper Cretaceous aquitard continues to restrict recharge to the Dakota and underlying aquifers in western Kansas because of its thickness and low permeability. In southeastern Colorado the Arkansas River valley is located downgradient of the primary Dakota aquifer recharge area and removes some of the underflow that otherwise would move into western Kansas and thus contributes to subhydrostatic heads in the Dakota and other aquifers below. In central Kansas the Saline and Smoky Hill rivers have eroded through the Upper Cretaceous aquitard west of the main outcrop belt of the Dakota aquifer. This has effectively moved the discharge area closer to the confined Dakota aquifer in western Kansas and thus contributes to the observed subhydrostatic conditions mentioned above (Belitz, 1985; Belitz and Bredehoeft, 1988).

As mentioned above, east of the Denver basin the Upper Cretaceous aquitard thins and its control on flow systems in the Dakota and underlying shallow aquifers diminishes. Within the last 10 million years differential uplift and intense local dissection of the High Plains surface by erosion have created considerable local and regional topographic relief (Gable and Hatton, 1983;

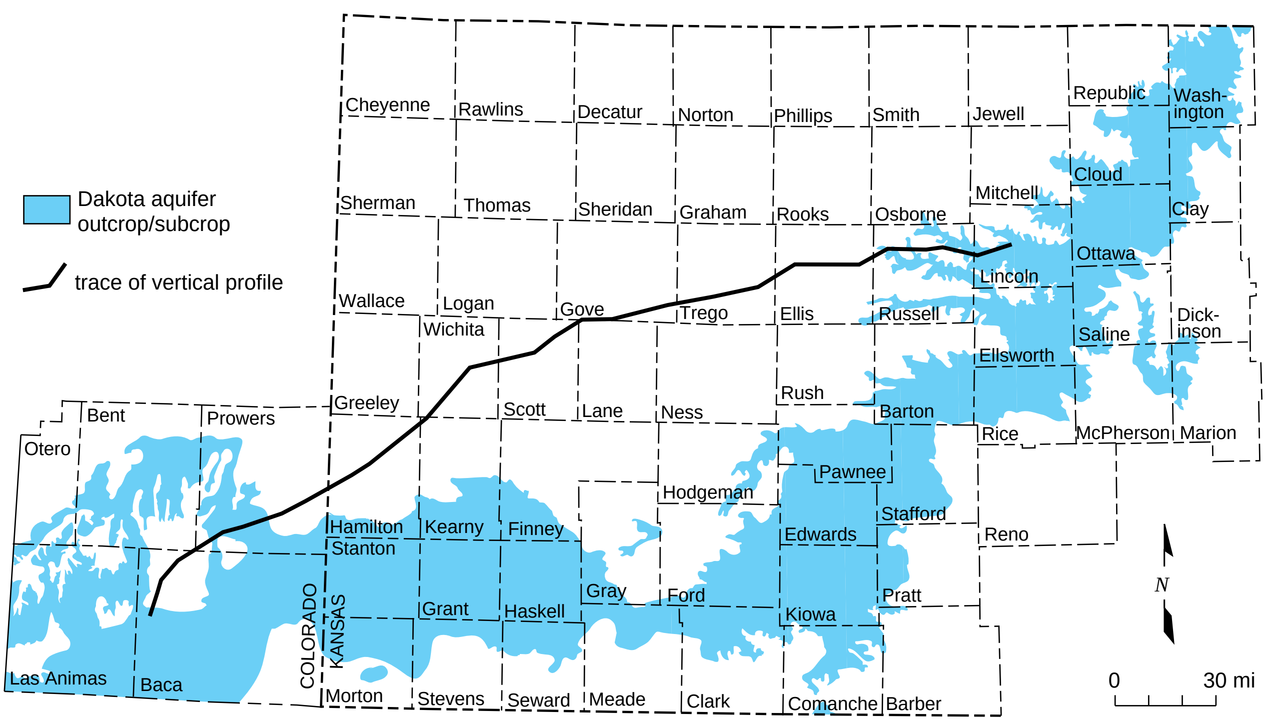


FIGURE 1. Extent of study area in Kansas and Colorado. The shaded area shows the area of outcrop, primarily in central Kansas, and subcrop beneath Pleistocene and Tertiary deposits in southwestern Kansas and southeastern Colorado.

Trimble, 1980; Osterkamp et al., 1987). Many of the rivers that cross the central Great Plains, such as the Arkansas, the Saline, and the Smoky Hill, have cut down through the aquitard and into the Dakota aquifer. High local relief favors the subdivision of the regional flow system into smaller subsystems, especially near river valleys (Toth, 1962, 1963; Freeze and Witherspoon, 1967). Consequently, the head difference between the Dakota aquifer and the overlying water table is reduced in many areas because of the aquifer's proximity to the near-surface hydrologic environment. Helgeson et al. (1995) and Leonard et al. (1983) recognized a separate, less stagnant flow component in the Dakota aquifer in the study area that is not present in the Denver basin and emphasized the importance of cross-formational flow.

Regional investigations into the hydrogeology of the Dakota aquifer since Darton's time have shown that the head differences between regional recharge and discharge areas and the hydrologic properties of the aquifer are not the only significant controls on ground-water flow in the Dakota. Darton (1906) did not recognize that the Upper Cretaceous aquitard could induce subhydrostatic heads in the underlying Dakota by restricting recharge. Neither Darton's work nor the later modeling studies reported by Belitz (1985), Belitz and Bredehoeft (1988), and Helgeson et al. (1994) address the influence of the Arkansas River and the Smoky Hill and Saline rivers on the flow system in the Dakota aquifer in western Kansas.

## Approach

Two of the primary uses of computer simulation in hydrogeology are to evaluate conceptualizations of ground-water flow system dynamics and to make inferences on system dynamics based on these conceptualizations. Anderson and Woessner (1992) refer to these uses collectively as the interpretive application of computer simulation. In this study simulation is used only in this interpretative sense to assess the relative importance of the Upper Cretaceous aquitard and the Saline and Arkansas rivers to the flow system in the Dakota aquifer. A two-dimensional model of ground-water flow in the vertical plane (a vertical profile model) for the upper part of the regional flow system is the basis for testing the conceptual model. The vertical profile extends from the Baca–Las Animas county line in southeastern Colorado to western Lincoln County in central Kansas and is parallel to the flow directions in all the major aquifers in the upper part of the regional flow system (fig. 2).

Because most of the available data on hydrologic properties and heads come from the upper part of the Dakota and other shallow aquifers and from much deeper hydrocarbon reservoirs in Permian and Pennsylvanian rocks, the nature of the flow system below the Dakota and above these hydrocarbon reservoirs can only be inferred. Hence there is less information by

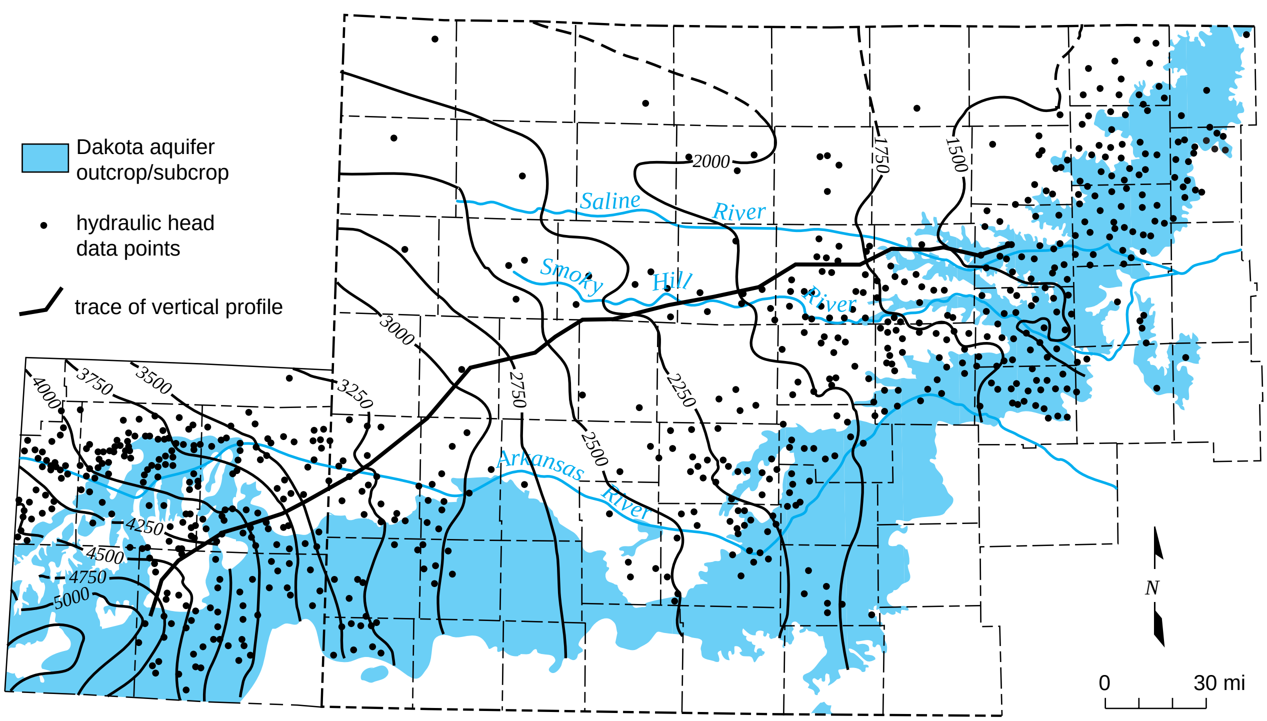


FIGURE 2. Elevation in feet above mean sea level of the predevelopment potentiometric surface of the Dakota aquifer in southeastern Colorado and western Kansas. The shaded area shows the area of outcrop, primarily in central Kansas and subcrop beneath Pleistocene and Tertiary deposits in southwestern Kansas and southeastern Colorado. Hydraulic head data are from the U.S. Geological Survey's Central Midwest Regional Aquifer System Analysis Program.

which to calibrate the model. This is acceptable because a fully calibrated model is not required when the purpose of the simulation is interpretive (Anderson and Woessner, 1992).

Accordingly, the objectives here are (1) to characterize the hydrogeology of the upper part of the regional flow system in southeastern Colorado and western and central Kansas, (2) to describe the construction of a vertical profile model of the upper part of the flow system, (3) to discuss the flow patterns in the partially calibrated steady-state model and its associated water budget, and (4) to present the results of sensitivity analyses that show the effect of the hydrostratigraphy and the river valleys on the flow system.

# Regional Setting

## Physiography

Southeastern Colorado and southwestern and central Kansas are located in the Raton Section, Colorado Piedmont, High Plains, and Plains Border sections of the Great Plains physiographic province (Fenneman, 1946). The land surface slopes to the east and decreases in elevation from approximately 5,000 ft (1,500 m) in southeastern Colorado to 1,400 ft (430 m) in central Kansas (fig. 3). The regional land-surface slope ranges from 26.7 ft/mi (5.06 m/km) in southeastern Colorado to 10.6 ft/mi (2.01 m/km) in western Kansas to 6.7 ft/mi (1.3 m/km) in central Kansas in the vertical profile. The vertical profile traverses the Arkansas, Smoky Hill, and Saline River drainage basins. The valleys cut by these river systems into unconsolidated Cenozoic deposits and Cretaceous bedrock significantly increase the topographic relief. In the Arkansas River valley of southeastern Colorado and southwestern Kansas and in the Saline River valley of central Kansas, the local relief commonly exceeds 200 ft/mi (37.9 m/km).

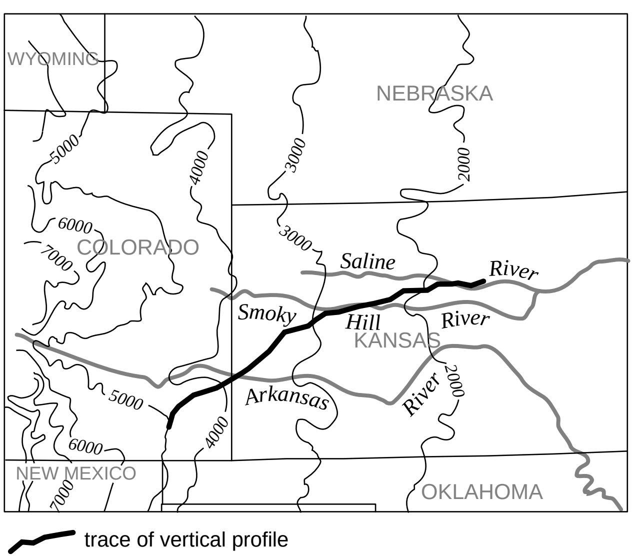


Figure 3. Generalized land-surface topography and the major streams traversed by the vertical profile in eastern Colorado and western and central Kansas and adjacent areas.

## Climate

The climate of the region is warm, continental semiarid in all except the eastern portions of the study area in central Kansas, where the climate is subhumid continental (Dugan and Peckanpaugh, 1985). The mean annual temperature is approximately 54˚F (12.2˚C) across the study area. Mean annual rainfall for the period 1951–1980 ranged from 15 inches (38 cm) in southeastern Colorado to 28.5 inches (72.4 cm) in central Kansas. Approximately 75% of the precipitation falls mainly during the warm season months of the year. Because of low relative humidity, high average wind velocities, and abundant sunshine, potential evaporation exceeds average annual precipitation over most of the region. Dugan and Peckanpaugh (1985) calculated that the potential mean annual recharge to ground water from precipitation ranges from less than 0.1 inch (0.2 cm) in southeastern Colorado to 1–2 inches (5 cm) in central Kansas.

# Regional Hydrostratigraphy

Regional stratigraphy and hydrostratigraphy are summarized in Table 1. The methodology used to define regional hydrostratigraphic units is discussed in detail by Macfarlane et al. (1992) and Macfarlane (1993). The hydrostratigraphy consists of six major aquifers and three aquitards. The most important of these to this research are the upper Dakota aquifer and the overlying Upper Cretaceous aquitard. Previous investigations have established the preeminence of the Upper Cretaceous aquitard as a major factor that exerts control on the flow system in the central Great Plains (Helgeson et al., 1994; Belitz, 1985; Belitz and Bredehoeft, 1988; Leonard et al., 1983; Helgeson et al., 1982). The present investigation is focused on the influence of the aquitard on the underlying flow system. The upper Dakota aquifer is considered important because it is hydraulically continuous across the vertical profile and is more transmissive than the other shallow aquifers below the Upper Cretaceous aquitard. This suggests that the upper Dakota aquifer acts as a drain beneath the aquitard and transmits most of the water moving through the upper part of the flow system from southeastern Colorado to central Kansas.

## Upper Cretaceous Aquitard

The Upper Cretaceous aquitard consists of a thick sequence of rhythmically bedded chalky shale, massive limestone and chalky limestone, dark-gray noncalcareous to calcareous shale and siltstone, and thin seams of bentonite (Hattin, 1962, 1965, 1975, 1982; Hattin and Siemers, 1987). Included in the aquitard are strata from the Niobrara Chalk, the Carlile Shale, the Greenhorn Limestone, and the Graneros Shale (Table 1).

### Upper Dakota Aquifer

The upper Dakota aquifer consists of mudstones and lenticular very fine to coarse-grained and conglomeratic sandstones belonging to the Dakota Formation in Kansas and its stratigraphic equivalent in southeastern Colorado, the Dakota Sandstone (McLaughlin, 1954; Franks, 1966, 1975; Macfarlane et al., 1990; Macfarlane et al., 1991). Sandstone composes 30 to 40% of the aquifer framework regionally (Keene and Bayne, 1977), but locally the percentage of sandstone can range from less than 20% to more than 80% (Macfarlane et al., 1992). The thickness of the upper Dakota aquifer ranges from approximately 350 ft (107 m) in parts of west-central Kansas to approximately 200 ft (61 m) in Baca County, Colorado.

Sediments belonging to the Dakota Formation and the Dakota Sandstone were deposited in fluvial, coastal plain, deltaic, and shallow marine environments in association with the developing Western Interior seaway (Weimer, 1984). Fluvial channel sandstones were deposited in incised valleys and in coastal plain settings in stacked fining-upward sequences up to 100 ft (30 m) in thickness (Hamilton, 1989; Macfarlane et al., 1991). Finer-grained deltaic and shallow marine sandstones are present in the upper part of the Dakota Formation and are generally much less than 100 ft (30 m) in thickness in central Kansas. However, deltaic deposits make up most of the thickness of the Dakota Formation in western Kansas and southeastern Colorado.

## Steady-State Regional Ground-Water Flow in the Major Aquifers

The major aquifer systems in the shallow subsurface of southeastern Colorado and western Kansas are the High Plains and alluvial valley aquifers, the Dakota aquifer, the Morrison-Dockum aquifer, and the Permian sandstone aquifer (Table 1). A deep aquifer in Lower Paleozoic carbonate rocks is not included in this discussion because it is present only in the shallow subsurface of southeastern Colorado. For this discussion, only the flow system in the upper Dakota aquifer is discussed in detail because only in southeastern Colorado and extreme southwestern Kansas are the hydraulic head data adequate to fully portray the flow system in the lower Dakota aquifer and the Morrison-Dockum aquifers. However, the flow patterns in the lower Dakota are believed to be similar to those in the upper Dakota in most of Kansas. In the Permian sandstone aquifer, data are insufficient to delineate the predevelopment head distribution (Macfarlane, 1993).

The Dakota aquifer is the most geographically extensive of all the aquifer systems in the shallow (upper 2,000 ft; 610 m) subsurface of western Kansas and southeastern Colorado. In parts of southwestern Kansas and southeastern Colorado, the Dakota is hydraulically connected to the High Plains, alluvial valley, and Morrison-Dockum aquifers (Robson and Banta, 1987; Kume and Spinazola,

TABLE 1. Stratigraphy and hydrostratigraphy of the shallow (upper 2,000 ft) subsurface in the vertical profile from southeastern Colorado to western and central Kansas.

| Era | System | Rock stratigraphic units | Hydrostatigraphic units |
|---|---|---|---|
| Cenozoic | Quaternary | Unconsolidited sediments | High Plains and alluvial valley aquifers |
| | Tertiary | Ogallala Formation | |
| Mesozoic | Cretaceous | Colorado Group | Upper Cretaceous aquitard |
| | | Dakota Sandstone/ Dakota Formation | Upper Dakota aquifer |
| | | Purgatoire Formation/ Kiowa Formation | Kiowa shale aquitard |
| | | | Lower Dakota aquifer |
| | Jurassic/Triassic | Morrison Formation Dockum Group | Morrison-Dockum aquifer |
| Paleozoic | Permian/ Pennsylvanian | Permian undifferentiated | Permian-Pennsylvanian aquitard |
| | | Lyons Sandstone/ Cedar Hills Sandstone | Permian sandstone aquifer |
| | | Permian/Pennsylvanian undifferentiated | Permian-Pennsylvanian aquitard |

1985). In central Kansas it is hydraulically connected to the Permian sandstone aquifer and alluvial valley aquifers (Macfarlane et al., 1988). The Dakota aquifer is confined by the Upper Cretaceous aquitard in most of western Kansas and southeastern Colorado and is the near-surface aquifer in western Baca and eastern Las Animas counties in Colorado.

Two distinct ground-water flow corridors can be distinguished on the predevelopment potentiometric surface map of the Dakota (fig. 2). The northern flow corridor begins in eastern Las Animas County and extends northeastward across the Arkansas River into west-central and northwestern Kansas and turns eastward into central Kansas. The southern flow corridor begins in eastern Las Animas County and extends eastward into southwestern Kansas and then turns northeastward toward central Kansas. The primary recharge area for the Dakota aquifer in Kansas is southeastern Las Animas and western Baca counties in Colorado on the Sierra Grande uplift. In this area the Dakota aquifer is at the surface and is recharged directly by infiltrating precipitation (fig. 1). The primary ground-water discharge area appears to be central Kansas where the major drainages, such as the Smoky Hill and Saline rivers, cross the outcrop of the Dakota aquifer. In this area salt springs, seeps, and marshes are a common occurrence (Macfarlane et al., 1990).

The vertical profile is parallel to one of the flow paths in the northern flow corridor. The slope of the potentiometric surface in the vertical profile ranges from 24.8 ft/mi (4.70 m/km) in southeastern Colorado to 10 ft/mi (1.9 m/km) in southwestern Kansas to 6.8 ft/mi (1.3 m/km) in central Kansas and reflects the eastward decrease in regional topographic slope.

Figure 4 shows the overall high degree of correlation ($r = 0.993$, where $r$ is the correlation coefficient) between hydraulic head in the Dakota aquifer and land-surface elevation, which suggests that regional topography is a primary control on regional ground-water flow. However, near the 3,250-ft (991-m) land-surface elevation, in areas of western Kansas where the Upper Cretaceous aquitard is thickest, some of the data points significantly depart from

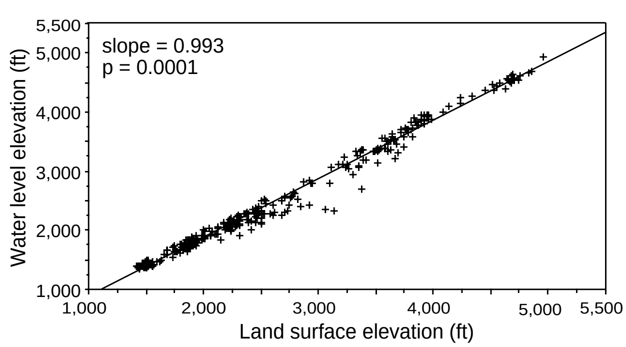


FIGURE 4. Water-level elevation vs. land-surface elevation in the confined Dakota aquifer, southeastern Colorado and western and central Kansas.

the best-fitting line. This suggests that the effect of topography on regional flow is diminished by the effect of the Upper Cretaceous aquitard on hydraulic heads in the Dakota aquifer.

The lateral flow component in the Dakota aquifer and the tendency for downward flow from the surface to the Dakota are shown by nearly all the fluid pressure vs. depth profiles (fig. 5) and the well depth vs. fluid pressure plots (figs. 6–8). Most of the fluid pressure vs. depth profiles come from sites located just upgradient from the Dakota aquifer discharge area in central Kansas. In all but two of these (fig. 5, profiles 1 and 7), the slope of the profile in the Dakota aquifer interval approximates the hydrostatic rate of increase of fluid pressure but is shifted downward below the hydrostatic line. This indicates a tendency for lateral flow within the Dakota and downward flow from units above. The fluid pressure versus depth profile from NW NW NW sec. 6, T. 14 S., R. 13 W., indicates downward flow across both the Upper Cretaceous aquitard and the upper Dakota to lower zones (fig. 5, profile 1). The other profile (fig. 5, profile 7) is from the Haberer salt marsh in northwestern Russell County, Kansas, and is located where the Dakota aquifer discharges to the overlying alluvial aquifer in the Saline River valley. In the upper 130 ft of the Dakota aquifer and the lower part of the alluvial aquifer, the slope of the fluid pressure versus depth profile is higher than the slope of the hydrostatic line. This indicates a significant tendency for upward flow from the Dakota to the overlying alluvial aquifer.

The fitted slopes of the fluid pressure vs. depth regression lines are less than the slope of the hydrostatic line both for wells located in the confined Dakota aquifer (fig. 6) and in the Dakota outcrop area (fig. 7). Figure 6 indicates that the direction of flow is generally downward across the aquitard and affirms that heads are subhydrostatic in the confined Dakota. Figure 7 indicates that over most of its outcrop area, the Dakota is readily recharged by precipitation. The t-test shows that the slope of the best-fitting lines (0.29 psi/ft for both confined and outcrop areas) is significantly less than the slope of the hydrostatic line (0.433 psi/ft). The t-test determines whether the slope difference between the best fit regression line and the hydrostatic line is a real phenomenon. The high correlation between fluid pressure and depth in the confined areas ($r = 0.90$) suggests that flow in the Dakota is uniform and primarily lateral (Fogg and Prouty, 1986).

Fluid pressures plot both above and below the hydrostatic line where the Dakota is overlain by the High Plains and alluvial valley aquifers (fig. 8). A t-test shows that the slope of the best-fitting line (0.38 psi/ft) is significantly less than the slope of the hydrostatic line by a small margin. The relatively high slope of the best-fit regression line as compared with the lines in figures 6 and 7 indicates that the Dakota and the overlying High Plains and alluvial valley aquifers act as a single aquifer system. The upward

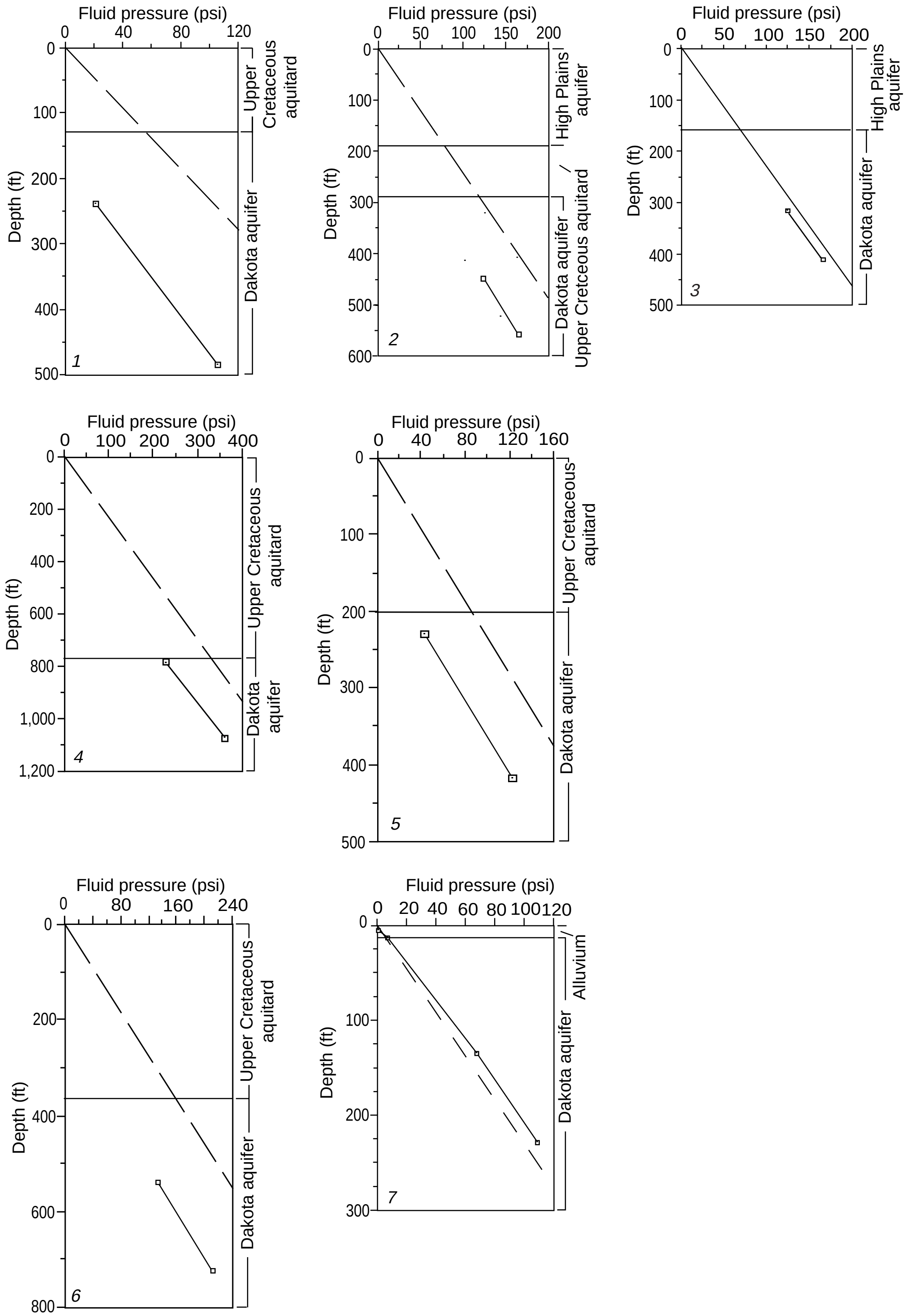


FIGURE 5. Fluid pressure vs. depth profiles from field measurements collected at seven sites: (1) NW NW NW sec. 6, T. 14 S., R. 13 W.; (2) SE NE NW sec. 29, T. 28 S., R. 26 W.; (3) NE SE SW NW sec. 30, T. 28 S., R. 22 W.; (4) NW SE SW sec. 2, T. 8 S., R. 23 W.; (5) NW SW NW sec. 14, T. 12 S., R. 16 W.; (6) SW SW SW sec. 31, T. 12 S., R. 17 W.; and (7) NW sec. 14, T. 12 S., R. 15 W. The dashed line represents the hydrostatic line, 2.309 ft/psi.

shift of the best-ftting line probably results because the Dakota is in good hydraulic connection with the overlying water table. In figure 5, the pressure vs. depth profile 3 is shifted downward only slightly with respect to the hydrostatic line, which also suggests good hydraulic connection between the Dakota and the overlying water table.

# Simulation of the Steady-State Flow System

Computer simulation of the flow system was a three-stage process, which included model design, derivation of hydraulic properties, and sensitivity analysis of the resultant partially calibrated model. Model design involved discretization of the vertical profile to produce a model grid and setting the boundary and initial conditions and the initial hydraulic parameter estimates for each of the hydrostratigraphic units. In the second stage the steady-state model was used to estimate the vertical hydraulic conductivity of the aquitards and the transmissivity of the aquifers using known hydraulic head and flow rate information. Finally, sensitivity analysis was applied to the partially calibrated model developed in the second stage to determine the major influences on the flow system.

## Governing Equation

The governing equation that describes the flow of ground water in a vertical profile parallel to the flow direction is (Anderson and Woessner, 1992)

$$\frac{\partial}{\partial x}\left(K_X \frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial z}\left(K_Z \frac{\partial h}{\partial z}\right)+R=0, \qquad (1)$$

where $R$ is a source/sink term and $K_X$ and $K_Z$ are the $x$ and $z$ components of hydraulic conductivity. Equation 1 describes ground-water flow through a heterogeneous and anisotropic porous medium where the principal axes of hydraulic conductivity are aligned with the orthogonal $x$ and $z$ coordinate system axes. Sources of recharge to and discharge from the model are not indicated explicitly because they are handled separately as part of the boundary conditions, which are discussed below.

MODFLOW (McDonald and Harbaugh, 1988) was used to solve equation 1 along with its attendant boundary and initial conditions in the vertical profile. MODFLOW is a block-centered, finite-difference code that can be used to simulate ground-water flow in two or three dimensions. The model has a modular structure and consists of a main program and a series of subroutines referred to as modules. These subroutines are grouped into packages that deal with specific features of the hydrologic system to be simulated or with a numerical technique to solve the finite-difference formulation of the flow equation. MODFLOW was selected for this application because it can be readily adapted to a vertical profile model (Anderson and Woessner, 1992).

## Model Grid

The vertical profile model grid consists of 8 layers, 1 row, and 73 columns of cells (fig. 9). The row has a unit length of 1 ft (0.3 m) perpendicular to the plane of the page and the column length of each cell varies from 5,709 ft (1,740 m) to 28,545 ft (8,700.5 m) along the profile (the $x$-axis in fig. 9). This spacing in the $x$-direction is variable to more accurately simulate the upper water table boundary in regions of significant topographic relief. The total length of the vertical profile from southeastern Colorado to central Kansas is 326.8 mi (536.5 km). Each of the

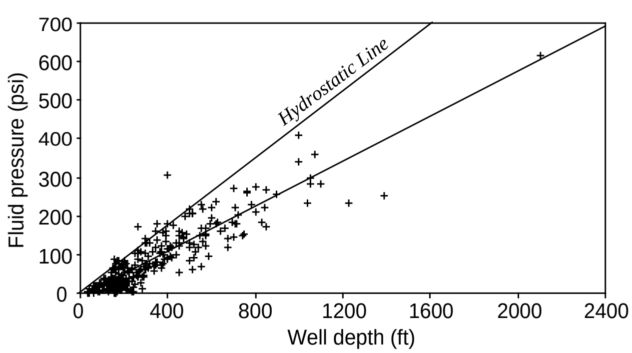


FIGURE 6. Well depth vs. fluid pressure for data collected from areas where the Dakota is a confined aquifer. Fluid pressure is calculated from water-level and well-construction information. The slope of the best-fit line is 0.29. The correlation coefficient, *r*, is 0.90 and is statistically significant ($p$ = 0.0001). The best-fit line slope is significantly less than the slope of the hydrostatic line ($p < 0.025$).

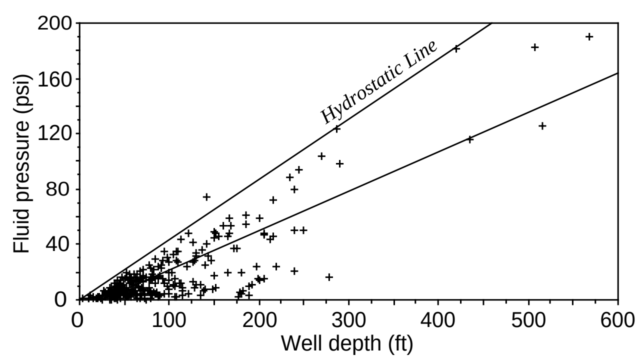


FIGURE 7. Well depth vs. fluid pressure for data collected from areas where the Dakota is the near-surface aquifer. Fluid pressure is calculated from water-level and well-construction information. The correlation coefficient, *r*, is 0.85 and is statistically significant ($p$ = 0.0001). The slope of the best-fit line is 0.29 and is significantly less than the slope of the hydrostatic line ($p < 0.025$).

TABLE 2. Input hydraulic conductivity data for the hydrostratigraphic units in the vertical profile.

| Hydrostratigraphic unit | Horizontal hydraulic conductivity (ft/day) | Vertical hydraulic conductivity (ft/day) |
|---|---|---|
| High Plains aquifer | 80 | 8.0 |
| Alluvial valley aquifers | 250 | 25 |
| Upper Cretaceous aquitard | $9.0 \times 10^{-7}$ | $9.0 \times 10^{-8}$ |
| Upper Dakota aquifer | 4 – 10 | $3.1 \times 10^{-3}$ |
| Kiowa shale aquitard | $1.3 \times 10^{-5}$ | $1.3 \times 10^{-6}$ |
| Lower Dakota aquifer | 2.3 –2.0 | $3.1 \times 10^{-3}$ |
| Morrison-Dockum aquifer | 0.15 – 0.5 | 0.015 – 0.05 |
| Permian-Pennsylvanian aquitard | $2.7 \times 10^{-3} – 2.7 \times 10^{-5}$ | $2.7 \times 10^{-4} – 2.7 \times 10^{-6}$ |
| Permian sandstone aquifer | 1.6 | 0.16 |

model layers represents a hydrostratigraphic unit. Layer 1 is the High Plains and alluvial valley aquifers and is treated as the upper, unconfined aquifer by MODFLOW . Layers 2 and 3 are the Upper Cretaceous aquitard and the upper Dakota aquifer, respectively. These layers are treated by MODFLOW as fully convertible layers between confined and unconfined conditions. Layers 4 through 8, the Kiowa shale aquitard, the lower Dakota aquifer, the Morrison-Dockum aquifer, the Permian-Pennsylvanian aquitard, and the Permian sandstone aquifer, are all treated as confined layers by MODFLOW . The terms *confined* and *unconfined* do not necessarily denote aquifer units in the model and, with the exception of layer 1, indicate only whether a particular layer is uppermost at some point in the model. The values for horizontal and vertical hydraulic conductivity, shown in Table 2, were based on data collected from the literature and from unpublished sources (Macfarlane, 1993).

All the major geologic and geomorphic features traversed by the vertical profile along its length were simulated, including the Sierra Grande and Las Animas uplifts, Two Buttes dome, the Arkansas River, the Saline River drainage, and Big Creek. The Smoky Hill River valley was not simulated because of its low relief. The Bear Creek fault was not simulated in the vertical profile because its influence on the flow system is uncertain and is probably only local.

The model grid was subdivided into three major sections on the basis of relative local and regional topographic relief in vertical profile view (fig. 9). The southeastern Colorado upland extends from the southwestern end of the model to the Arkansas River in column 17. In this part of the model, regional topographic slope is steep and local topographic relief is only moderate. The western Kansas plains section of the model extends from column 18 to the north bank of Big Creek in column 50. In this section, regional topographic slope is moderate and local topographic relief is low. The central Kansas dissected plains section extends from column 51 to the northeast end of the model in column 73. Here regional topographic slope is low and local topographic relief is high.

The model grid and the input parameters were designed to reflect changes in the hydrostratigraphy in the vertical profile caused by the pinching out of model layers. Such pinch-out is seen in layers 4 (the Kiowa Shale aquitard) and 6 (the Morrison-Dockum aquifer) in figure 9 and is taken into account by continuing the layer across the model as a phantom with a transmissivity and a layer thickness of zero. Vertical hydraulic continuity is maintained by assigning the same vertical conductance to the cells in the phantom layer that is assigned to cells in the overlying real layer. The vertical conductance of each cell in the real layer above was calculated by assuming that both real layers are in physical contact.

## Boundary Conditions

The boundary conditions define the hydraulic conditions on the perimeter of the model and are necessary to produce a unique solution to the flow equation (Anderson and Woessner, 1992). The upper boundary of the model

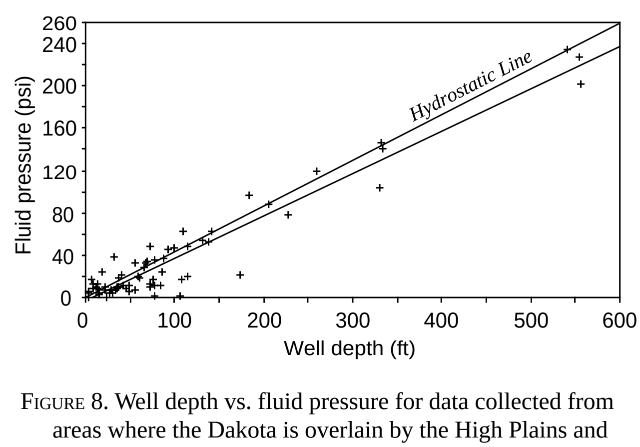


FIGURE 8. Well depth vs. fluid pressure for data collected from areas where the Dakota is overlain by the High Plains and alluvial valley aquifers. Fluid pressure is calculated from water-level and well-construction information. The slope of the best-fit line is 0.38 psi/ft. The correlation coefficient, *r,* is 0.90 and is statistically significant ($p = 0.0001$). The best-fit line slope is significantly less than the slope of the hydrostatic line ($p < 0.025$).

(fig. 9) represents the water table and is considered a specified-head boundary. At this boundary temporal fluctuations in the head of the water table are small relative to the total head difference across the model (3,560 ft; 1,085 m) and the maximum vertical extent of the model [up to 1,700 ft (518 m)]. The specified-head boundary condition was applied instead of a flux boundary to minimize the number of parameters that needed adjusting during calibration. The specified-head condition allows a flux of water (recharge or discharge) to cross the water table during model execution to maintain the constant head in each cell.

Specified-head boundary conditions were also applied at the southwestern and northeastern boundaries of the model. On these boundaries there are time-invariant, vertical hydraulic head gradients that are not significant relative to the scale of the regional model. The southwestern boundary corresponds to the 5,000 ft (1,524 m) equipotential, which is assumed to be vertical in the profile (fig. 9). Due to the limited amount of hydraulic head information available, placement of this boundary was guided by the results of modeling experiments discussed by Macfarlane (1993). The northeastern boundary corresponds to an assumed, vertical head difference of 0.5 ft (0.15 m) between the upper and lower Dakota aquifers at the northeastern end of the model beneath a tributary stream of the Saline River drainage. It is assumed that the model terminates beneath the discharge area of a local flow system involving the tributary stream. The small head difference allows for the discharge of ground water from the local flow system into the stream.

A vertical no-flow boundary was used along the bottom of the model to simulate the horizontal flow line that approximates the boundary separating the shallow, intermediate-flow system from the deeper, regional-flow system (fig. 9). This flow line was drawn on the basis of modeling results described by Macfarlane (1993).

## Calibration

Calibration of ground-water flow models usually consists of adjusting the input parameters until a satisfactory match is achieved between the observed and the simulated hydraulic heads, fluxes, or other calibration targets (Wang and Anderson, 1982). In this research a fully calibrated model of the flow system was not possible because of the lack of head data for many of the layers below the upper Dakota aquifer. Model calibration was

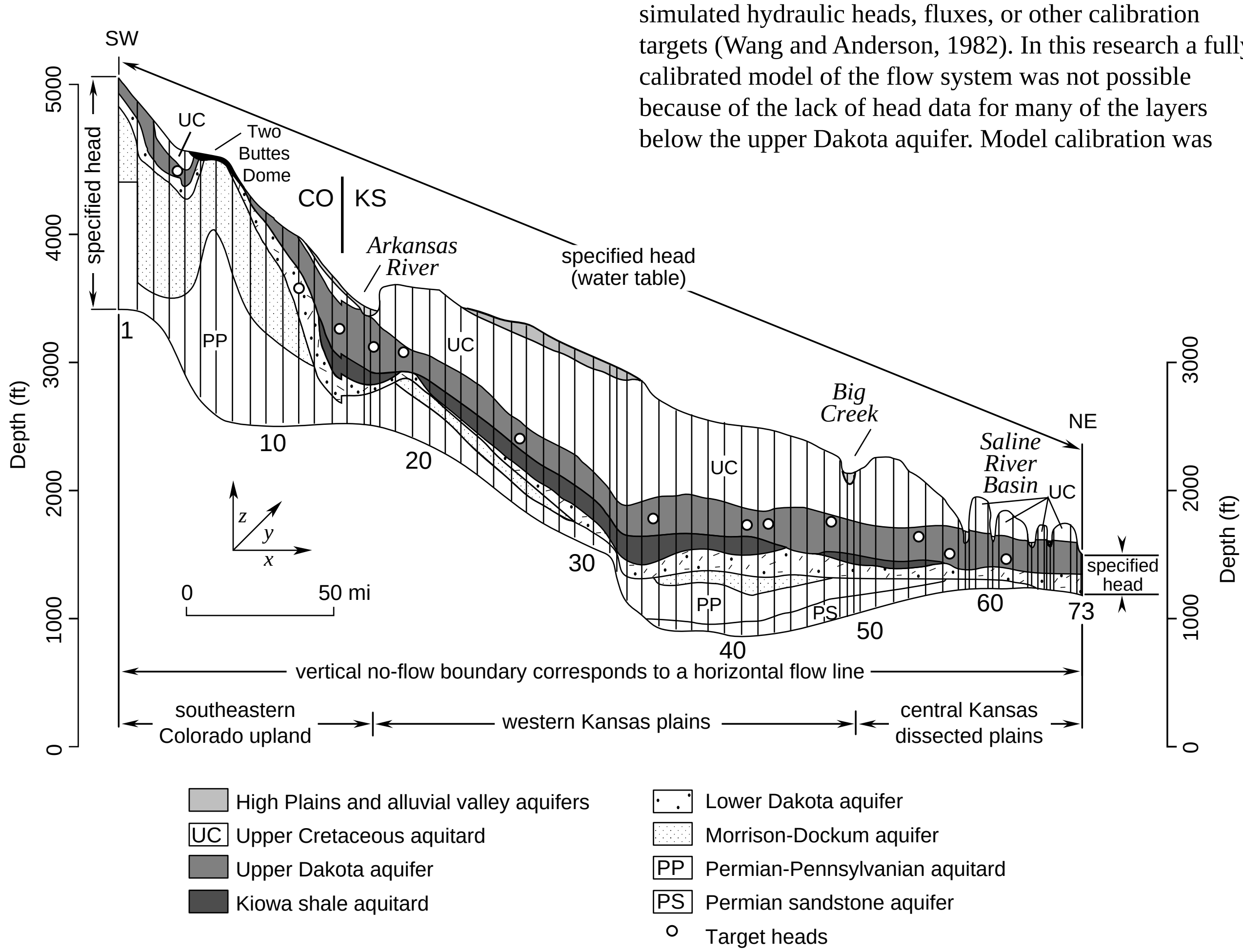


FIGURE 9. The vertical profile model grid consists of 8 layers, 1 row, and 73 columns of cells. The length of each cell in the column direction (along the *x* axis) is variable and ranges from 5,709 ft to 28,545 ft. The length of each cell in the row direction (along the *y* axis) is 1 ft. The location of the vertical profile is shown in figures 1 to 3. The vertical exaggeration is 217.6×.

carried out manually by trial-and-error adjustment of the hydraulic conductivity input data to match hydraulic head measurements and flow rates in the model. Because most of the head data were primarily from the High Plains, alluvial valley, and upper Dakota aquifers, little adjustment was made in the hydraulic parameters of layers below the upper Dakota aquifer. All the adjustments made in the values of these parameters were guided by the sensitivity analyses. Fourteen target head values were available in the upper and lower Dakota aquifers to check the progress of calibration (fig. 9). Calibration of the vertical profile model was also guided by the results of pumping tests of nearby wells in central Kansas and in southeastern Colorado and measurements of baseflow in the Saline River in central Kansas from seepage runs.

The results of each round of calibration were evaluated by computing the root mean square (RMS) error (Anderson and Woessner, 1992):

$$\text{RMS error} = [(1/n)\sum (h_m - h_s)^2]^{0.5}, \qquad (2)$$

where $h_m$ and $h_s$ are the measured and simulated heads, respectively. The RMS error was chosen because it is thought to be the best measure of uncertainty if the errors are normally distributed (Anderson and Woessner, 1992). It was also used to evaluate the sensitivity of the model to systematic changes in layer hydraulic conductivity and boundary conditions.

Another criterion used to evaluate the calibrated model was the distribution of differences between the simulated and target (measured) head values ($h_m - h_s$) across the model (Anderson and Woessner, 1992). The differences were examined for trend by producing a plot of these differences with distance across the model and computing a best-fitting regression line by least squares through the data points. No trend is indicated in the errors if the slope of the best-fitting line is judged to be not significantly different from zero. The absence of a trend in these differences across the model suggests that further adjustment of the model parameters and boundary conditions may not be needed to bring the model into calibration.

The steady-state model was considered to be partially calibrated when the RMS error was less than 50 ft (15 m), which is 1.4% of the total head decline (3,560 ft; 1,085 m) across the model. This value of the RMS error is also within the error of many of the calibration target heads. The RMS error of the partially calibrated model is 46 ft (14 m). The slope of the best-fitting line was 0.046 ft/mi (0.009 m/km) and was found to be not significantly different from zero (fig. 10).

## Ground-Water Flow in the Steady-State Model

The head distribution in the partially calibrated model of the steady-state flow system is indicated by the pattern of the equipotentials shown in figure 11. Figure 12 shows the distribution of recharge and discharge across the upper model boundary. The cell-by-cell flow rates within the aquifer units were computed by MODFLOW for the constant-head cells. The positive and negative flow rates represent the net recharge and discharge, respectively, through each of the cells along the upper boundary. Not considered in this calculation is the flow of water between adjacent constant-head cells. Thus the model only calculates flow vertically into or out of the model. For the other cells in the model, ZONEBUDGET (Harbaugh, 1990) was used to calculate cell-by-cell water budgets.

### Southeastern Colorado Uplands

In the southeastern Colorado upland section of the model, the equipotentials are spaced closely together and are vertical in orientation (fig. 11). The steep head gradient from the southwestern end of the model to the Arkansas River is controlled by the nature of the boundary conditions: the high regional slope of the water table and the specified head at the upgradient end of the model. Figure 12 shows an alternation of recharge and discharge across the upper model boundary in this section. The alternation of recharge and discharge suggests laterally adjacent local flow systems, especially in the vicinity of Two Buttes dome. The U.S. Geological Survey 1:100,000 scale, 30´ × 60´ topographic maps of the Springfield and Two Buttes quadrangles show an abundance of springs in the vicinity of Two Buttes. The equipotentials near the upper model boundary poorly define these local flow systems because of the coarseness of the model grid and the boundary conditions. However, the moderate local relief suggests that the local flow systems are probably shallow in vertical extent (Toth, 1963). Darton (1906) considered most of the valley in the Two Buttes Creek drainage

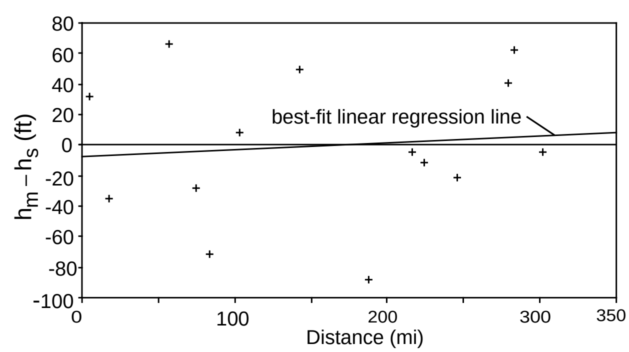


FIGURE 10. Distribution of the difference between target ($h_m$) and predicted heads ($h_s$) in the partially calibrated steady-state model. The distance is measured from the southwestern end of the vertical profile model. The slope of the best-fit linear regression line is 0.046 ft/mi (0.009 m/km) and is not significantly different from zero. The estimated error for the target heads is ± 50 ft (15.2 m).

between Two Buttes dome and the Arkansas River as an area where flowing-well conditions in the Dakota could be expected. This is consistent with the model results and the interpretation of laterally adjacent local flow systems. Figure 12 shows a small discharge of water from the flow system in columns 9 and 10, near where Two Buttes Creek intersects the vertical profile. The model also shows that nearby uplands are recharge areas.

In column 11, upgradient of the Arkansas River, more than 4 $ft^3$/day (0.1 $m^3$/day) enters the model across the water-table boundary or approximately 0.6 inch/yr (1.5 cm/yr) of net recharge (fig. 12). Toward the river the rate of recharge decreases rapidly until column 14, at which point water is discharged from the model at steadily increasing rates until the Arkansas River valley is reached in column 16. This pattern of recharge and discharge suggests that the vertical profile traverses a local flow system involving the south side of the Arkansas River valley and the adjacent upland. In this part of the model, recharge entering the upper Dakota aquifer must cross the Upper Cretaceous aquitard, which has a thickness of less than 100 ft (30 m) and a vertical hydraulic conductivity of approximately $10^{-5}$ ft/day ($3 \times 10^{-6}$ m/day). In the Arkansas River valley the simulated steady-state heads in the upper Dakota aquifer are much higher than the elevation of the water table. Haworth (1913) mentioned flowing wells in the river valley near Coolidge, Kansas, and reported that the static water level in the first wells was approximately 20 ft (6 m) above land surface.

Approximately 10 $ft^3$/day (0.3 $m^3$/day) enters the flow system in the southeastern Colorado upland southwest of the Arkansas River and 6.3 $ft^3$/day (0.2 $m^3$/day) is discharged to springs and streams locally (fig. 12). The remainder moves on toward the Arkansas River with the subsurface inflow that enters at the southwestern end of the model (1.6 $ft^3$/day; 0.05 $m^3$/day). The model results indicate that all of the flow in the upper Dakota aquifer in column 16 is discharged to the Arkansas River valley at a rate of 3.7 $ft^3$/day (0.1 $m^3$/day) per foot of river channel or approximately 0.2 $ft^3$/sec/mi (0.004 $m^3$/sec/km). This value of baseflow is consistent with anecdotal accounts by travelers through the region in the 1840s and 1850s, which mention the low rates of stream discharge in this stretch of the Arkansas River. The model results show that only a small amount is discharged from the upper Dakota aquifer directly beneath the river in comparison with the

Figure 11. Steady-state head distribution in the partially calibrated vertical profile model. Ground-water flow is from regions of higher to regions of lower hydraulic head.

discharge on the south side of the valley. Most of the flow to the river comes from recharge that passes through a considerable thickness of the Upper Cretaceous aquitard from the north side of the valley before it enters the river or the upper Dakota aquifer.

### Western Kansas plains

In the western Kansas plains section of the model, ground water is transmitted laterally through the Dakota aquifer beyond the Arkansas River to the central Kansas dissected plains section of the model (fig. 11). The High Plains aquifer is readily recharged by infiltrating precipitation. However, the nearly horizontal orientation of the equipotentials in the underlying Upper Cretaceous aquitard indicates that the flow system beneath the aquitard is hydraulically isolated from the High Plains aquifer. Net recharge rates are negligible, on the order of $10^{-3}$ $ft^3$/day through each of the cells along the upper model boundary, and the total recharge for this section of the model is relatively small, approximately 0.15 $ft^3$/day (0.004 $m^3$/day) or approximately $7.5 \times 10^{-4}$ in/yr (0.002 cm/yr) (fig. 12). Ground-water flow through the aquitard to each of the model cells in the upper Dakota aquifer amounts to 0.5% or less of the total cell-by-cell volumetric flow rate [approximately 1.6 $ft^3$/day (0.05 $m^3$/day)].

### Central Kansas Dissected Plains

Farther east in the central Kansas dissected plains section of the model, local topographic relief is pronounced where the model intersects the Saline River and its tributaries and the regional slope of the land surface is relatively low. The high local topographic relief and the low regional slope favor local flow-system development rather than a continuation of the intermediate-scale flow system into this section of the model from the western Kansas plains (Toth, 1963). The total net recharge to this section of the model is 4.7 $ft^3$/day (0.3 $m^3$/day) or approximately 0.06 in/yr (0.15 cm/yr) and is much lower than the amount in the southeastern Colorado upland. Recharge through the aquitard to the upper Dakota constitutes up to 100% of the highly variable cell-by-cell flow rate in the upper Dakota aquifer, but on the average is less than 20%.

Beneath the Saline River, the model results show that heads in the upper Dakota should be approximately 14 ft (4.3 m) higher than the elevation of the water table at this location. Field measurements indicate that the head difference is approximately 8 ft (2.4 m) near here. Low-flow measurements in the Saline River indicate a discharge of approximately 1.3 $ft^3$/day/mi (0.02 $m^3$/day/km) of river channel from the regional flow system (J. B. Gillespie, personal communication, 1993). This rate of discharge is reasonably close to the approximately 1.9 $ft^3$/day/mi (0.05 $m^3$/day/km) predicted by the partially calibrated steady-state model when the vertical hydraulic conductivity of the Upper Cretaceous aquitard is approximately $10^{-6}$ ft/day in the vicinity of the river.

## Steady-state Water Budget

The water budget is summarized in figure 13 and shows recharge to and discharge from the steady-state flow system for each of the three model sections. The total inflow through the vertical profile is approximately 16.6 $ft^3$/day (0.47 $m^3$/day) through the 1 ft (0.3 m) wide cross section. In the southeastern Colorado upland and in the central Kansas dissected plains, most of the water that enters the model through the upper boundary is discharged locally and little is contributed to regional flow (fig. 13). Recharge entering the flow system across the water table in the southeastern Colorado upland accounts for 61% of the total water budget, but 60.5% is discharged locally to springs and streams. In the central Kansas dissected plains 28.4% of the total water budget enters the flow system through the Upper Cretaceous aquitard as local recharge, but 38.9% of the total is discharged locally. In contrast, only 0.9% of the total water budget recharges the flow system in the western Kansas plains section of the model. There is no local discharge to surface waters in this part of the flow system.

Recharge entering through the Upper Cretaceous aquitard in all three model sections accounts for approximately 30% of the total water budget to the model. This is consistent with Belitz's (1985) report that the probable amount of recharge through this confining layer in the Denver basin and adjacent areas to the east is in the range of 15–32% of the total inflow to the flow system. More than 97% of the total net recharge through the aquitard enters the system in the central Kansas dissected plains and in the southeastern Colorado upland section of the model where the aquitard is thinner and more permeable.

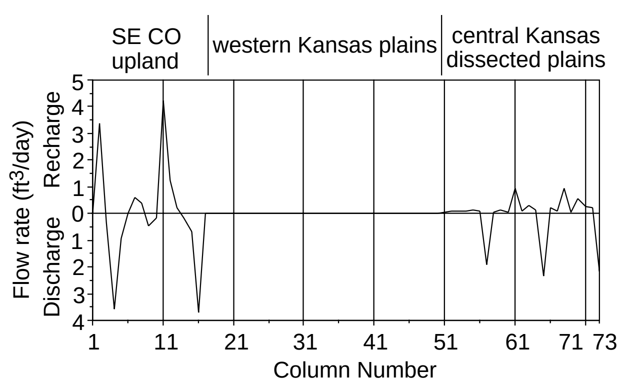


FIGURE 12. Distribution of recharge and discharge across the upper boundary of the partially calibrated vertical profile model calculated per foot of cross section. In the western Kansas plains part of the model only a very small amount of recharge, on the order of $10^{-2}$ to $10^{-3}$ $ft^3$/day per model cell is added.

The steady-state water budget shows that local flow systems and the vertical hydraulic conductivity of the Upper Cretaceous aquitard heavily influence the water budget in all the model sections. Local flow systems are present in the southeastern Colorado upland and in the central Kansas dissected plains because the local relief and surface drainage systems are sufficiently developed. However, almost twice as much water cycles through the southeastern Colorado upland as through the central Kansas dissected plains, where the flow of water is restricted by the Upper Cretaceous aquitard's greater extent and lower vertical hydraulic conductivity. By comparison, the flow through the western Kansas plains section of the model is sluggish. The low local relief and moderate regional slope of the land surface do not favor local flow system development, and the aquitard has much greater thickness and lower hydraulic conductivity in this part of the model than elsewhere. Thus cross-formational flow characterizes the upper part of the flow system in the southeastern Colorado upland and in the central Kansas dissected plains, whereas lateral flow characterizes the system in the western Kansas plains and in the deeper subsurface in the southeastern Colorado upland.

This view of ground-water flow in the upper part of the regional system is generally supported by the major ion ground-water geochemistry along the vertical profile. Bicarbonate-type waters with low total dissolved solids (TDS) concentrations (less than 500 mg/L) are the most common in the Dakota aquifer in the southeastern Colorado upland and reflect recharge from infiltrated precipitation or the overlying High Plains aquifer (Robson and Banta, 1987). The TDS concentrations increase slightly along the flow path and beneath the Arkansas River valley, and sulfate becomes a dominant constituent in the water. The deeper part of the flow system is believed to contain a mixed cation-sulfate, bicarbonate-type water with moderate TDS concentrations (500 mg/L to 1,000 mg/L). The increasing sulfate concentration in the upper Dakota aquifer probably results from water moving downward across the Upper Cretaceous aquitard near the valley (Macfarlane et al., 1992).

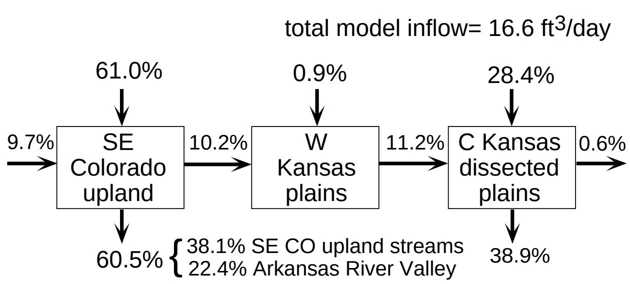


FIGURE 13. Components of the total water budget for the steady-state flow system in the vertical profile by model section per foot of cross section. Vertical arrows pointing into or out of each model section represent recharge or discharge into the section. Horizontal arrows represent underflow into or out of the model as a whole or moving between model sections.

In the western Kansas plains, the TDS concentrations of water samples from the Dakota aquifer continue to rise abruptly up to moderate (500 mg/L to 1,000 mg/L) to high (greater than 1,000 mg/L) levels and the water type changes to a sodium bicarbonate type. In contrast, ground water in the High Plains aquifer is a calcium bicarbonate-type water with low TDS concentrations. Eastward, the chloride concentration rises gradually in the upper Dakota aquifer from less than 50 mg/L to approximately 1,000 mg/l, and the water is a high-TDS-concentration, sodium chloride type. Water samples from minor aquifers below the Permian-Pennsylvanian aquitard contain high concentrations of chloride and are believed to be a sodium chloride type that has resulted from the dissolution of halite (Whittemore and Fabryka-Martin, 1992). Ground water from the Permian sandstone aquifer has very high chloride concentrations, above 5,000 mg/L, and more typically 15,000 to 20,000 mg/L. The presence of sodium bicarbonate- and sodium chloride-type ground water in the upper Dakota and other aquifers beneath the Upper Cretaceous aquitard provides additional evidence of the very small amount of freshwater recharge that is available to flush this part of the flow system of remnant formation water (Macfarlane et al., 1990; Whittemore and Fabryka-Martin, 1992).

In the central Kansas dissected plains, ground-water TDS concentrations rise abruptly up to more than 10,000 mg/L in the upper Dakota beneath the Saline River. Ground water beneath the Saline River is a highly saline, sodium chloride type that eventually makes its way into the stream as baseflow. The discharge of saline ground water to the Saline River provides additional evidence of upward flow from the lower Dakota and Permian sandstone aquifers beneath the river in local flow systems. Downgradient of the Saline River, the chloride and TDS concentrations slowly decrease in the upper Dakota aquifer to moderate levels. Near the northeast end of the vertical profile and beyond, the TDS concentrations fall to low levels and bicarbonate type waters again become the most prevalent. The presence of sodium bicarbonate- and sodium chloride-type ground water in the upper Dakota where it is confined by the Upper Cretaceous aquitard provides good evidence that only a small amount of freshwater recharge moves through the aquitard to the upper Dakota in this part of the flow system (Macfarlane et al., 1990).

## Influence of the Upper Cretaceous Aquitard on the Flow System

Before evaluating the model's sensitivity to the Upper Cretaceous aquitard, first it had to be determined whether a uniform or a nonuniform vertical hydraulic conductivity best explained the observed heads in the upper Dakota aquifer. This was done by first calibrating the model assuming a uniform value of vertical hydraulic conductivity in the layer. Adjustments were made in the hydraulic properties of all the layers to produce a minimum RMS error calculated using equation 2. In the next series of model runs, in which the vertical hydraulic conductivity in the aquitard was allowed to vary from cell to cell, the model was then recalibrated using the adjusted hydraulic conductivites in the other layers from the first series of simulations, adjusting only vertical hydraulic conductivity to produce a new minimum RMS error. Model sensitivity to the vertical hydraulic conductivity was evaluated by running a series of simulations in which the vertical hydraulic conductivity was varied systematically through a range of values spanning two to three orders of magnitude. The effect on the model of changing the value of the parameter was determined by calculating the RMS error after each model run.

In the first series of simulations, where values for hydraulic conductivity were uniform, only a maximum value of the vertical hydraulic conductivity of the Upper Cretaceous aquitard could be determined from the sensitivity analysis. The asymmetry of the error curve shown in figure 14 suggests a maximum vertical hydraulic conductivity of $3 \times 10^{-7}$ ft/day ($9 \times 10^{-8}$ m/day). The RMS error increases from 79 ft (24 m) to 164 ft (80 m) as the vertical hydraulic conductivity increases to $3 \times 10^{-5}$ ft/day ($9 \times 10^{-6}$ m/day). Decreases in the vertical hydraulic conductivity below $3.0 \times 10^{-7}$ ft/day ($9 \times 10^{-8}$ m/day) have little effect on the RMS error.

In the second series the vertical hydraulic conductivity of the aquitard was assumed to vary from cell to cell in the model. The resulting RMS error decreased by 42% to 46 ft (14 m). This indicates that a nonuniform vertical hydraulic conductivity more appropriately characterizes the layer than a constant value. In the individual sections of the model, the vertical hydraulic conductivity ranges from $3.0 \times 10^{-4}$ to $2.0 \times 10^{-5}$ ft/day ($9.1 \times 10^{-5}$ to $6 \times 10^{-6}$ m/day) in the southeastern Colorado upland, $3.8 \times 10^{-7}$ to $1.7 \times 10^{-7}$ ft/day ($1.2 \times 10^{-7}$ to $5.2 \times 10^{-8}$ m/day) in the western Kansas plains, and $6.3 \times 10^{-6}$ ft/day to $7.1 \times 10^{-7}$ ft/day ($1.9 \times 10^{-6}$ to $2.2 \times 10^{-7}$ m/day) in the central Kansas dissected plains.

The results from both simulations are consistent with the results of Belitz (1985) and Belitz and Bredehoeft (1988). They found that only a maximum vertical hydraulic conductivity could be determined from model sensitivity if the vertical hydraulic conductivity was treated as a uniform property of the layer. Their maximum value is one order of magnitude less than the maximum value reported from the first simulation series. They also reported improvement in the error of their multilayer model of the Denver basin when the vertical hydraulic conductivity was treated as a depth-dependent variable ranging over three orders of magnitude from 100 ft to 10,000 ft (30–3,000 m) of depth. Figure 15 is a plot of the partially calibrated cell-by-cell vertical hydraulic conductivity versus the depth to the center of each cell in layer 2. The log-log plot and the best-fitting line through the data show that vertical hydraulic conductivity generally decreases with depth. Vertical hydraulic conductivity decreases over three orders of magnitude for a depth range from 17.5 ft to 522.5 ft (5.3 m to 159 m). The $r^2$ value indicates that approximately 61% of the variation in the data is explained by the log-log relationship between the two variables and demonstrates that vertical hydraulic conductivity is a depth-dependent variable in this aquitard.

In figure 16 the RMS error is very sensitive to the increase in vertical hydraulic conductivity above the calibrated cell-by-cell set of values. The sensitivity of heads in the upper Dakota aquifer to increases in this parameter results because of the increasing recharge that enters the model through the aquitard as it becomes more permeable. The heads in the upper Dakota between the Arkansas and the Saline rivers begin to approach the head of the water table. The net recharge increases by more than 100 times in this section of the model when the vertical hydraulic conductivity is increased by a factor of 1000. The RMS error is not as sensitive to decreases in vertical hydraulic conductivity below the partially calibrated cell-by-cell set of values. Decreases in vertical hydraulic conductivity cause the head difference between the water table and the upper Dakota aquifer to increase by only approximately 50 ft (15 m), which is nearly within the error tolerance of the calibration. This indicates that the net recharge is negligible and underflow accounts for most of the flow in the upper Dakota aquifer.

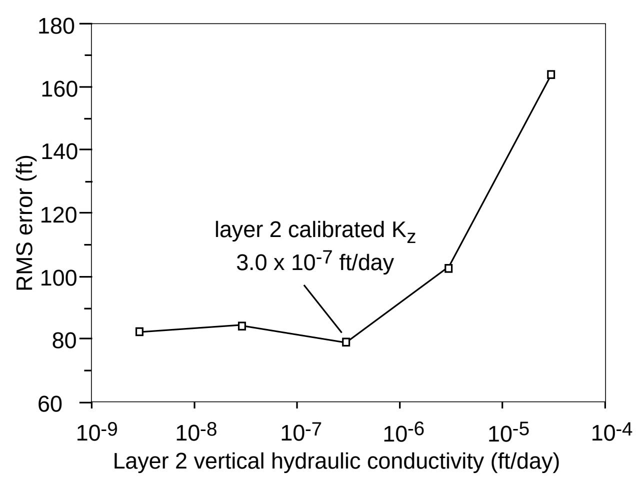


FIGURE 14. Model sensitivity to the uniform vertical hydraulic conductivity of the Upper Cretaceous aquitard, expressed as the RMS error. The calibrated value of vertical hydraulic conductivity for the aquitard in this simulation is $3.0 \times 10^{-7}$ ft/day.

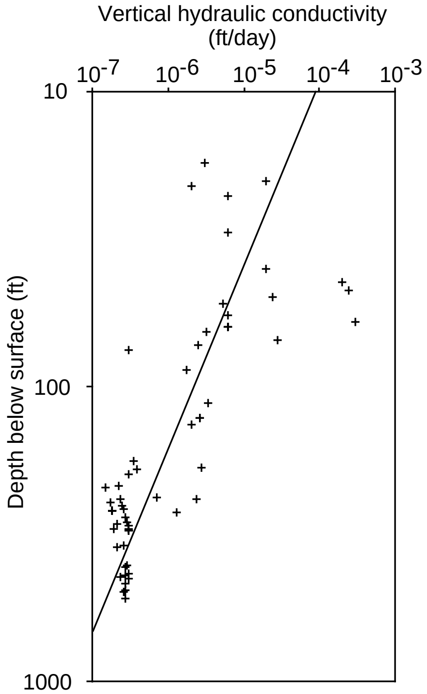


FIGURE 15. Vertical hydraulic conductivity of the Upper Cretaceous aquitard vs. depth below land surface showing the best-fit line by least-squares regression. The value of the correlation coefficient, *r*, is 0.78.

# Role of River Valleys

The effect of the Arkansas River valley on the flow system was evaluated by removing the feature from the simulation. Layer 1, the High Plains and alluvial valley aquifers, was removed and replaced by layer 2 as the uppermost model layer (fig. 9). The upper model boundary was modified in the vicinity of the river by removing the valley and restoring hypothetically a semblance of the pre-erosional topographic profile. The specified heads along the upper boundary were increased in the vicinity of the river accordingly (columns 12–18). Other changes were made to the model input to reflect changes in layering, layer thicknesses, and location of the water table with respect to layer boundaries.

The influence of the Saline River and its tributary streams on the flow system was also investigated using this same procedure. Once the model grid was redesigned, the heads along the water table were changed in columns 53–73 to reflect the modified upper model boundary. The head difference between the upper Dakota potentiometric surface and the water table was calculated for the restored topography simulations and plotted. These head differences were compared with the head differences in the partially calibrated model for each cell to determine the effect of the river valleys on the flow system.

## Arkansas River Valley

Removing the Arkansas River valley from the model grid produced large increases in the head in the upper Dakota near the former position of the river in column 17 and for some distance downgradient into western Kansas (fig. 17). From column 17 to 32, a distance of approximately 67 mi (108 km), heads in the upper Dakota aquifer are approximately equal to or greater than the elevation of the water table in this simulation. In comparison, heads in the upper Dakota in this section of the partially calibrated model are 200–300 ft (61–91 m) lower than the elevation of the water table. Downgradient of column 32 to the Saline River (column 57) the effect of removing the river diminishes with distance. Without the river, only a small amount of water (0.04 $ft^3$/day; 0.001 $m^3$/day) is discharged across the upper model boundary from the Upper Cretaceous aquitard and there is a net decrease in the total inflow to the model of 24%. This reduction in the flow of water through the model reflects the removal of the local flow system near the Arkansas River valley and its associated flow of water in the partially calibrated model.

The extent of head change caused by removing the river valley from the steady-state model demonstrates that it is a significant influence on the flow system in the upper Dakota aquifer in western Kansas. The pattern of recharge and discharge in the model shows that the river valley is a major discharge area that removes underflow from the upper Dakota aquifer just downgradient of the recharge area in southeastern Colorado. Thus it appears that the local flow system in the valley vicinity helps to maintain subhydrostatic conditions by discharging water from the flow system and hydraulically isolating the upper Dakota downgradient.

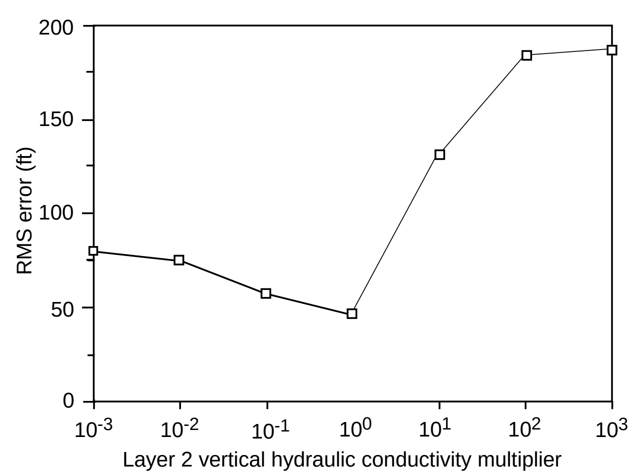


FIGURE 16. Model sensitivity to the nonuniform vertical hydraulic conductivity of the Upper Cretaceous aquitard in the partially calibrated model, expressed as RMS error.

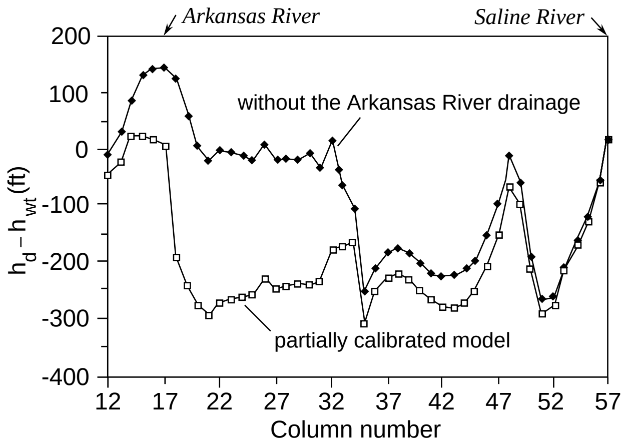


FIGURE 17. The head difference between the upper Dakota aquifer ($h_d$) and the overlying water table ($h_{wt}$) when the Arkansas River is removed from the simulation. The head difference in the partially calibrated model is shown for comparison.

## Saline River Drainage

Removal of the Saline River and its tributary streams from the model produces smaller changes in head in the upper Dakota between columns 17 and 57 than when the Arkansas River is removed from the model (fig. 18). The increase in head in the upper Dakota ranges from 1 ft (0.3 m) beneath the Arkansas River in column 17 to slightly more than 100 ft (30 m) in west-central and central Kansas (columns 42 to 52). The total water budget for the model decreases by 18% because of the removal of the local flow systems associated with the Saline River valley and its tributaries.

The head increase that results when the Saline River and its tributaries are removed from the simulation can be explained using a simple model proposed by Belitz (1985). Belitz related the head in a confined Dakota aquifer to its hydraulic properties and those of the overlying Upper Cretaceous aquitard, the head on the overlying water table and in the discharge area, and the geometry of the flow path from the water table to the discharge area (fig. 19):

$$(K_c/K_a)\ (L/D) = (h_b - h_d)/(h_r - h_b), \qquad (3)$$

where $K_a$ and $K_c$ are the horizontal and vertical hydraulic conductivities of the aquifer and aquitard, respectively; $L$ is the distance from a point in the confined aquifer to the discharge area; $D$ is the aquitard thickness; and $h_b$, $h_d$, and $h_r$ are the head in the confined aquifer at any point, the head in the discharge area, and the head on the overlying water table, respectively.

Removal of the Saline River drainage from the steady-state model moves the discharge point beyond the northeast end of the model and increases the distance $L$. Because the heads on the water table overlying the confined aquifer and in the discharge area remain fixed, the increase in the distance to the discharge area $L$ will cause an increase in the head in the upper Dakota. In essence, the hydraulic connection between the aquifer and the overlying water table improves when the discharge area is moved farther away. In central Kansas the Saline and Smoky Hill rivers have cut valleys through this aquitard to the west of the main discharge area of the confined Dakota aquifer. These valleys have effectively reduced the lateral distance between the discharge area and the confined Dakota aquifer in western Kansas and thus have helped generate subhydrostatic conditions in the upper Dakota aquifer. This effect is suggested by the slight bending of the 1,500-ft (457-m) and 1,750-ft (533-m) potentiometric contours near where the Saline and Smoky Hill rivers have cut through the Upper Cretaceous aquitard in Ellis and Russell counties, Kansas (fig. 2).

# Conclusions

The hypothesis advanced here is that the upper part of the steady-state regional flow system, including the Dakota aquifer, is influenced primarily by the Upper Cretaceous aquitard, the Arkansas River in southeastern Colorado, and the drainages in central Kansas that have cut down through the aquitard to the west of the main outcrop belt. The Upper Cretaceous aquitard is believed to allow significant recharge to the Dakota aquifer from the overlying water table in southeastern Colorado and in central Kansas where it is much thinner and more permeable or absent. The Arkansas River is located just downgradient from the main recharge area in southeastern

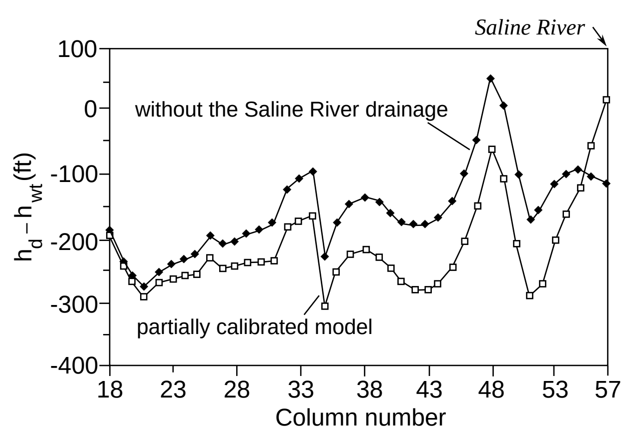


FIGURE 18. The difference in head between the upper Dakota aquifer ($h_d$) and the overlying water table ($h_{wt}$) when the Saline River drainage is removed from the model. The head difference in the partially calibrated model is shown for comparison.

Colorado and is believed to be a major discharge point for the western part of the flow system. The river influences the downgradient flow system by removing underflow that otherwise would continue into the confined Dakota aquifer in western Kansas. Consequently, the flow system beneath the Upper Cretaceous aquitard in western Kansas is isolated from its recharge area south of the river and from the overlying water table. The Saline River has eroded through the Upper Cretaceous aquitard to the west of the main outcrop area of the Dakota Formation in central Kansas and has effectively reduced the distance between the deeper, more confined parts of the Dakota aquifer and the discharge area. This has resulted in a further reduction of head in the Dakota aquifer upgradient from the discharge area. The Smoky Hill River in central Kansas may also influence the upgradient flow system in a similar way.

A steady-state numerical simulation of a portion of the intermediate-scale flow system in vertical profile view was developed to investigate the influence of these factors. The model results reveal significant development of local flow systems in the southeastern Colorado upland and the central Kansas dissected plains model sections. Local flow systems dominate in the central Kansas dissected plains section to the exclusion of the intermediate-scale flow system because of the high local relief associated with deeply incised river valleys. With the exception of the Arkansas River valley, local flow systems in the southeastern Colorado upland are not as well developed because of the high regional topographic slope but only moderate local relief. The steady-state water budget through the 1-ft (0.3 m) wide vertical profile is approximately 16.6 ft$^3$/day (0.5 m$^3$/day). Most of the recharge to the flow system is discharged to surface water locally in the southeastern Colorado upland and the central Kansas dissected plains sections. Ten percent of the inflow that enters the model in the upland moves beyond the Arkansas River and into the western Kansas plains. In this latter part of the model the amount of water moving into the intermediate-scale flow system through the Upper Cretaceous aquitard is 0.9% of the total inflow.

The sensitivity analyses demonstrate that the Arkansas River and the Saline River and its tributary streams in concert with the Upper Cretaceous aquitard heavily influence the flow system in the confined Dakota aquifer of western Kansas. The model is very sensitive to increases in the vertical hydraulic conductivity of the Upper Cretaceous aquitard. The lowest RMS error resulted when the aquitard vertical hydraulic conductivity was treated as a nonuniform parameter. Large increases in the RMS error caused by increasing vertical hydraulic conductivity indicate improvement in the hydraulic connection between the upper Dakota and the water table. Relatively small increases in the RMS error from decreasing the vertical hydraulic conductivity below the partially calibrated set of cell-by-cell values shows that most of the flow in the upper Dakota beneath western Kansas is underflow. Thus, the aquitard is a major regional influence on subhydrostatic conditions in the underlying flow system in the western Kansas plains section of the model. Simulated removal of the Arkansas River valley increased heads in the upper Dakota aquifer by as much as 200–300 ft (61–90 m) for a distance of 67 mi (108 km) downgradient in western Kansas. Simulated removal of the Saline River drainage increased heads in the upper Dakota aquifer by more than 100 ft (30 m) in parts of western and central Kansas.

The modeling results also have implications for the future development of water resources in the upper Dakota in west-central and central Kansas. The primary factors determining the success of long-term use of the upper Dakota in western and central Kansas are the sources and rates of recharge to the aquifer. In west-central Kansas, the major source of recharge is the underflow from upgradient sources. This, coupled with the low aquifer transmissivities, indicates that well-fields should be designed using low-capacity wells and a large well spacing to capture the underflow and to minimize overdrafting. In central Kansas, the major sources of recharge are more local but involve both fresh and saltwater sources. Thus, an additional concern is upconing of high TDS concentration saline ground water from the deeper aquifers during pumping.

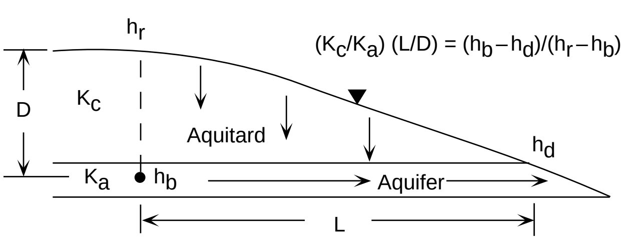


FIGURE 19. The combined effect of hydraulic conductivity, flow path geometry, and head on the water table and in the discharge area on the head in the confined aquifer at steady state; modified from Belitz (1985).

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