Heat Flow in the Cretaceous of Northwestern Kansas and Implications for Regional Hydrology

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Abstract

Temperature logs are interpreted to investigate the thermal structure of the units overlying the Kansas portion of the Cretaceous Dakota aquifer. The aim of this study is to determine if additional heat input by fluids exists and thus clarify whether the overall conductive heat flow from the basement through the sequence might be overprinted by heat advection. Although interval thermal gradients are determined for different lithologic (stratigraphic) units, the shale thermal gradients are preferred for heat-flow estimation. Shale thermal conductivity as measured in Mesozoic shales in Nebraska and South Dakota is extrapolated to the area because of the similar lithology. A few thermal-conductivity values are determined in sandstone samples of the Dakota Formation and also used in heat-flow estimation. In general, the noncalcareous, marine Cretaceous shales (Pierre, Carlile, Graneros, and Kiowa) show different thermal gradients. Gradients in the Pierre (average value 58.5°C/km) and Carlile (55.5°C/km) are slightly higher than the average gradient in the Graneros Shale (45.1°C/km) and Kiowa Formation (46.5°C/km). The higher thermal gradients are limited to the extreme northwestern corner of the study area where the Pierre and Carlile are present. The heat-flow density of 69–74 mW/m^2 observed there is slightly higher than the average of 60 mW/m^2 typical for central and eastern Kansas. The higher heat flow observed is in the range of data reported and mapped for northeastern Colorado and the Nebraska Panhandle on the western flank of the Chadron Arch, an area with geothermal overprint by warm fluids. Regional differences in heat flow in western Kansas seemingly are caused by the different composition, porosity, and permeability of the aquifer and the nearness to recharge areas.

Heat-flow studies in the Great Plains, and Nebraska in particular (Gosnold et al., 1981; Gosnold and Eversoll, 1981, 1982; Gosnold and Fischer, 1986; Gosnold, 1990), have shown that high heat flow and high geothermal gradients occur over extensive areas of ground-water discharge on the eastern margin of the Denver-Julesberg Basin, a major north-south Laramide feature, developed along the eastern front of the Rocky Mountains. The basin lies on the west-central margin of the Great Plains, occupying all of eastern Colorado and extending into Wyoming, Nebraska, and Kansas (Jorgensen et al., 1993). The basal Cretaceous sandstones, which include the Dakota sandstones, make up an important aquifer system within the basin.

In the past, published data on heat-flow density for the Kansas part of the Great Plains was limited: 63 mW/m^2 for central Kansas (Sass et al., 1971); 59 mW/m^2 for southwestern Kansas (Birch, 1947); 52–62 mW/m^2 for eastern Kansas (Blackwell and Steele, 1989). None of these data was from the area of the Dakota aquifer.

Thermal logs, which were measured for a preliminary study of the subsurface temperature conditions in Kansas and used to estimate geothermal resources (Steeples and Stavnes, 1982), show that the mean geothermal gradients range between 25°C/km and 55°C/km. The highest values were observed in the northwestern and north-central portion of the state. Because these mean temperature gradients give only a rough estimate of the thermal conditions of an area, there is room for speculation about the cause of such variations. In sedimentary basins where vertically different lithologic (stratigraphic) units occur, the lithology and petrophysical properties might also vary laterally to a great extent, affecting the subsurface temperature conditions. The separation of the different effects on the geothermal field requires the detailed interpretation of thermal logs. The study of geothermal gradients for lithologically homogenous units (formations) allows, then, in conjunction with the estimation of formation thermal conductivity, an estimate of heat-flow density along the profile, as shown by Blackwell and Steele (1989) for relatively deep boreholes in central and southeastern Kansas.

The temperature profiles in northwestern Kansas are from relatively shallow boreholes that penetrate rocks of Tertiary and Cretaceous age. Because of the structure of the Dakota aquifer, it is likely that the subsurface temperatures and geothermal gradients are affected by heat transfered by fluid flow out of the Denver-Julesberg Basin in the west. Therefore, the focus of this study is to investigate the thermal structure of the units overlying the Kansas portion of the Dakota aquifer to determine if additional heat input by fluids exists and to clarify whether the overall conductive heat flow from the basement through the sequence might be overprinted by heat advection.

Geologic Setting

The structure of the High Plains in western Kansas is relatively simple and straightforward. Cretaceous units overlie older Mesozoic and Paleozoic units and, in turn, are unconformably overlain by a veneer of Tertiary (Ogallala Formation) and Quaternary unconsolidated deposits (fig. 1). The sedimentary units, which rest on a crystalline basement of Precambrian age, range in thickness from about 1,100 m (3,500 ft) in the east to about 2,000 m (6,500 ft) in the west.

The Cretaceous covers much of the western one-half of Kansas. The configuration of these units reflects a different structural development from that expressed in the older Paleozoic units. The Hugoton Embayment, the major structural feature in western Kansas, was no longer active, and the structural pattern changed in the Mesozoic with the rise of the Rocky Mountains to the west. Major structural features recognizable in the Cretaceous are the Sierra Grande Uplift in southeastern Colorado, extending northeastward into western Kansas, and the Cambridge Arch, a southeast-northwest-trending positive feature extending southeastward into Kansas from Nebraska. In general, the Cretaceous units, other than where affected by these major structural features, dip uniformly and gently to the north and northwest as part of the southeastern flank of the Denver-Julesberg Basin, located farther west in Colorado. The units unconformably overlie either the Jurassic Morrison Formation, where present, or Permian red beds. The units are beveled from outcrop to about 915 m (3,000 ft) in thickness in the northwestern corner of Kansas in Cheyenne County.

In this study the thermal state of the stratigraphic units from the Kiowa Formation to Pierre Shale was investigated. The dominant lithology of the geologic units is given in figure 2 (see Merriam, 1957, 1963; Zeller, 1968 for details).

The Kiowa Formation (Kk) is predominantly a medium- to dark-gray, micaceous, silty, carbonaceous, soft to hard marine shale. Also included are minor amounts of limestone ("Champion shell bed"), quartzose sandstone, and bentonite. The shale becomes more silty and sandy east and southeast toward the old shoreline.

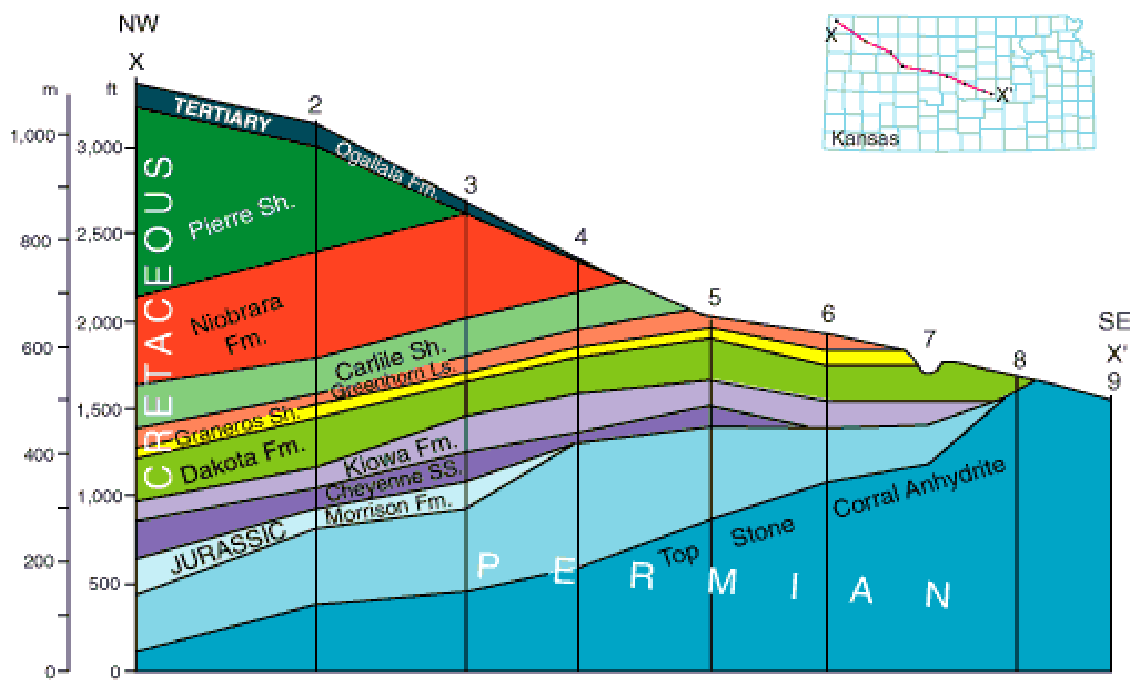

FIGURE 1. Northwest-southeast cross section of Cretaceous units in western Kansas. Cross section shows general relations of units to each other (generally conformable) and to the unconformable underlying Jurassic Morrison Formation and Permian redbeds and the unconformable overlying Tertiary Ogallala Formation. Datum = sea level. Index map shows location of cross section. Boreholes are numbered as follows: (1) Guy F. Atkinson No. 1 Beaumeister, SE SE NE sec. 31, T. 2 S., R. 39 W., Cheyenne County; (2) Miami No. 1 Stewart, NW NW NW sec. 7, T. 7 S., R. 33 W., Thomas County; (3) Prime No. 1 Fallow, SE SE NE sec. 19, T. 10 S., R. 27 W., Sheridan County; (4) Phillips No. 1 Folkers, SW SW SW sec. 31, T. 13 S., R. 23 W., Trego County; (5) Mackall No. 1 Herrman, NE SE NE sec. 2, T. 16 S., R. 18 W., Rush County; (6) Prairie No. 1 Rude, SW SW SW sec. 11, T. 16 S., R. 13 W., Barton County; (7) Empire No. 1 Kroesch, NW SE sec. 33, T. 17 S., R. 9 W., Ellsworth County; (8) Ladow et al. No. 1 King, NW NE sec. 20, T. 19 S., R. 5 W., McPherson County; (9) Dickey No. 1 Reusser, SE SE SE sec. 26, T. 21 S., R. 2 W., McPherson County.

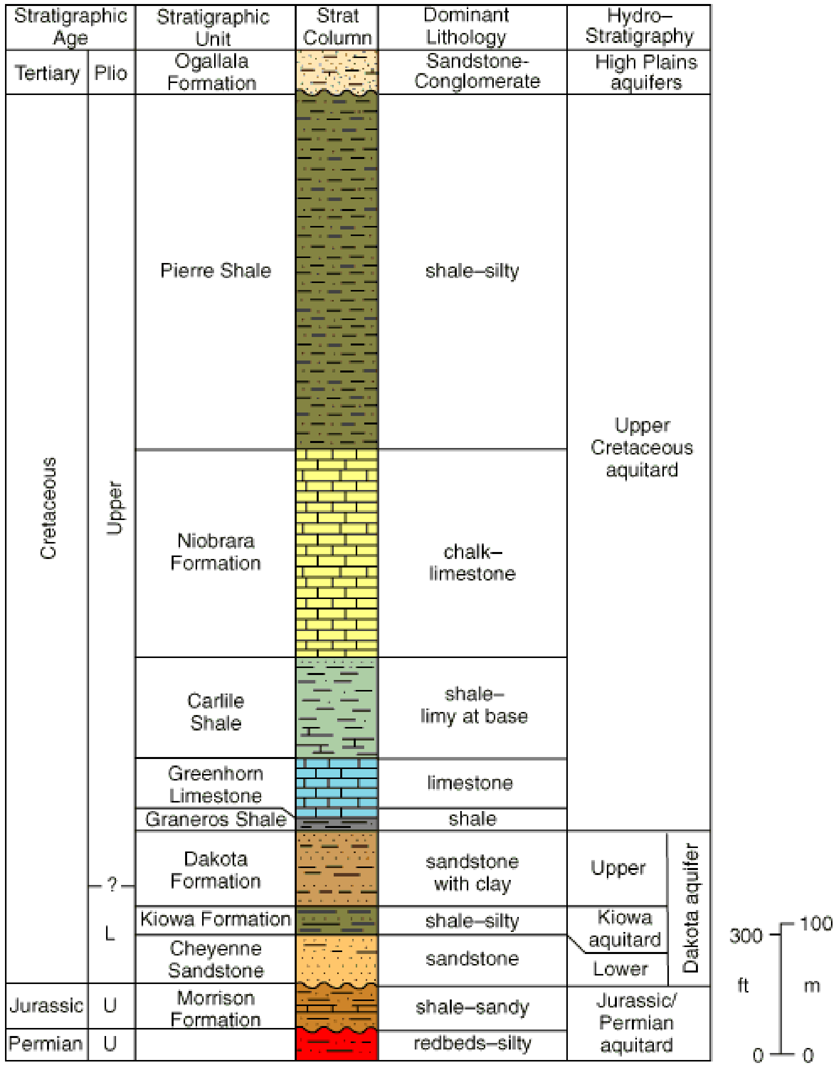

FIGURE 2. Generalized stratigraphic column of Cretaceous units in Kansas showing age, stratigraphic unit, dominant lithology, and hydrostratigraphy (adapted from Moore et al., 1952; Merriam, 1963).

The Dakota Formation (Kd) contains lenticular sandstones composed of light-gray or light-brown, cross bedded, fine- to medium-grained, subrounded quartz fragments. The unit also contains considerable noncalcareous, micaceous, clayey or silty shale with interbedded siltstone.

The Graneros Shale (Kg) is a medium-gray to black, noncalcareous or slightly calcareous, silty, marine shale. It contains thin streaks of bentonite.

The Greenhorn Limestone (Kgh) consists mostly of limestone and chalky shale. The limestone is gray to light brown, chalky or crystalline, and very fossiliferous. The shale is gray to brownish, calcareous, and fossiliferous.

The lower part of the Carlile Shale (Kc) is a very fossiliferous, chalky shale containing stringers of limestone and thin seams of bentonite. The middle part is a gray to blue-gray, fossiliferous, clayey, noncalcareous shale. The upper part, the Codell Member, is a thin, silty sandstone.

The lower part of the Niobrara Formation (Kn) (Fort Hays Limestone Member) is a massive, chalk or chalky limestone separated by thin beds of chalky shale. The upper part (Smoky Hill Chalk) is a chalky shale with massive chalk beds and many bentonite beds.

The Pierre Shale (Kp) is predominantly shale, which is light to dark gray, soft, slightly calcareous, micaceous, fissile, and fossiliferous. Many bentonite beds are present as well as many concretionary zones. The lower part (Sharon Springs Member) is a dark-gray to black, micaceous, clayey shale.

Hydrology

The Dakota aquifer underlies most of the western one-half of Kansas and nearly all of eastern Colorado and most of Nebraska. It can be divided into a lower aquifer, comprising stratigraphically the Lower Cretaceous Cheyenne Sandstone and the Longford Member of the

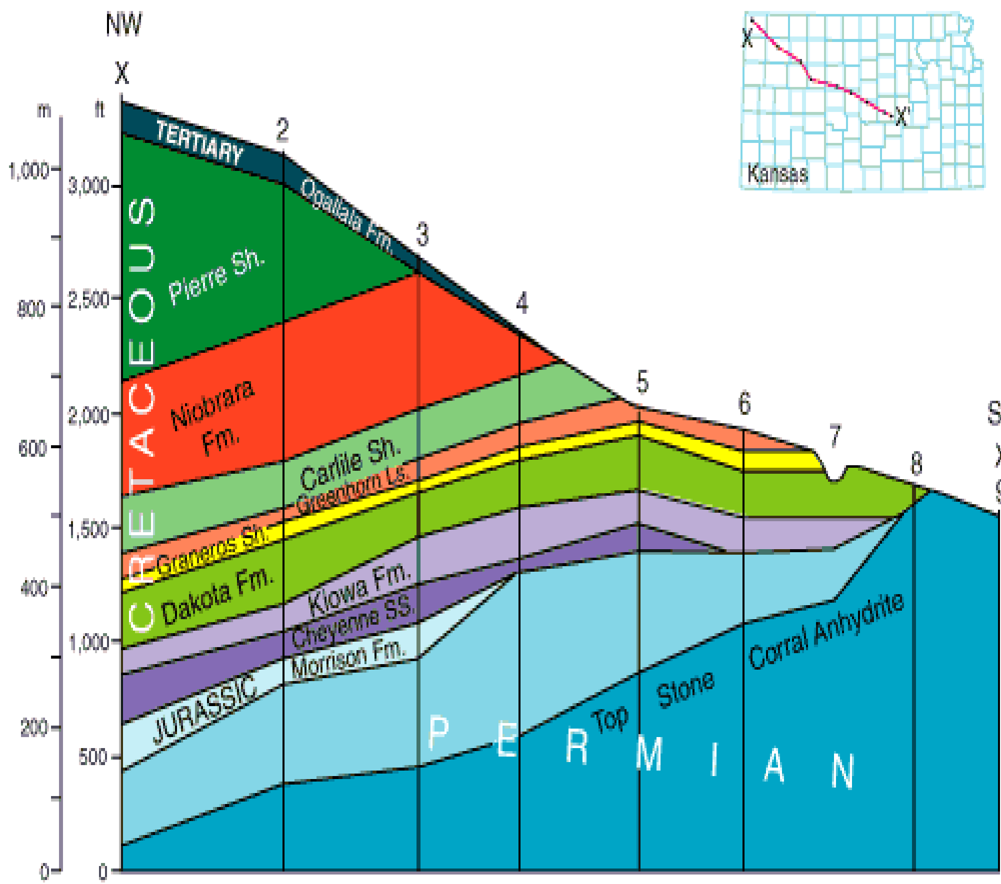

FIGURE 3. Map of structure on top of Dakota Formation in Kansas, Nebraska, eastern Colorado, and southwestern Wyoming showing general direction of fluid movement in Dakota aquifer (structure from Merriam, 1958; fluid movement from Helgesen et al., 1984; Pruit, 1978; Whittemore et al., 1993; McGovern and Wolf, 1993). Major structural features noted; outcrop of Dakota shown; CI = 500 ft (152 m).

Kiowa Formation, and an upper aquifer comprising the Upper Cretaceous Dakota Formation. These upper and lower aquifer units are separated by the Kiowa aquitard, composed of marine black shale (fig. 2). Upper Cretaceous rocks (aquitard of mostly shale) confine the aquifer system. Because of the limited depths of the boreholes that are available for the study, the geothermal investigations mostly are focused on the interval of the upper Dakota aquifer (Dakota Formation) and the overlying Upper Cretaceous aquitard.

Data on fluid movement in the Dakota aquifer was compiled from studies by Belitz and Bredehoft (1988), Helgesen et al. (1984), Macfarlane (1995), Pruit (1978), and Whittemore et al. (1993). Figure 3 shows the direction of fluid movement in the Dakota. The regional, steady-state ground-water flow in the Denver Basin and adjacent Midcontinent is characterized by subnormal fluid pressure and subnormal fluid potential according to the simulated and observed potentiometric surfaces shown by Belitz and Bredehoeft (1988). The largest differences between land surface and potentiometric surface occur in the Nebraska Panhandle and the extreme northwestern part of Kansas. With an input of ground-water flow rates of 1 m/year in heat-flow models as calculated for the Dakota in the Panhandle (Gosnold et al., 1981), there is a good agreement with the heat-flow data ranging between 80 and 100 mW/m^2.

In the Kansas part of the Dakota aquifer, the main pattern of ground-water flow is from the topographically high recharge areas in southeastern Colorado (Sierra Grande region) and southwestern Kansas to discharge areas in the river valleys of central and north-central Kansas (Macfarlane, 1995). Additional recharge is from the overlying Ogallala Formation and alluvial valley aquifers in southwestern Kansas, and from the underlying Permian aquifer where these aquifers are connected hydraulically in the central part of the state (Macfarlane et al., 1992). The average linear velocities in the confined portion of the aquifer are determined to be about 1 to 75 ft/year (0.3–23.4 m/year) (P. A. Macfarlane, personal communication, 1994). Because of the lack of thermal data for the eastern flank of the Denver Basin, it is unknown how far the heat-flow anomaly outlined for the basin area (Blackwell and Steele, 1992) extends into eastern Colorado and western Kansas and to what extent eastward flow of fluids could affect the geothermal conditions there.

TABLE 1. Boreholes discussed in this paper.

Borehole	Location	Longitude	Latitude	Temperature data from Stavnes	Temperature data from Blackwell	Temperature data from this paper	Completion year	Logging year
1CN2/Gibs	SW SE 24, 4S, 37W	-101.43	39.68	to 179 m	to 1,445 m		1981	summer 1981, 10/82
Finegan	NE SW 13, 6S, 38W	-101.51	39.53		to 1,400 m		1979	7/82
1SH4	SW SE NW 22, 6S, 38W	-101.55	39.51	to 178 m			1981	summer 1981
1TH3	SE NW 30, 8S, 35W	-101.27	39.33	to 183 m			1981	summer 1981
1GH5	N2 36, 7S, 25W	-101.07	39.40	to 302 m			1981	summer 1981
1NT5	NE SE SE 27, 5S, 21W	-99.67	39.40	to 174 m			1980	summer 1980
1RO1/Rooks	NW SE SW 27, 9S, 20W	-99.54	39.24	to 366 m	to 1,045 m		1980	summer 1980, 11/80
1EL1	SW NE SW 17, 13S, 18W	-99.35	38.92	to 118 m			1980	summer 1980
Rush/1RH1	NW NW NE 6, 17S, 17W	-99.25	38.61	to 163 m	to 162 m		1980	summer 1980, 11/80
1HG1	NW NE SE 27, 23S, 24W	-99.95	38.03	to 333 m			1980	summer 1981
1FO1	S2 N2 11, 28S, 21W	-99.59	37.62	to 182 m			1980	summer 1981
Hodgeman	NW NW SE 29, 23S, 23W	-99.88	39.03			to 122 m	1991	3/94
Stanton	SE SW SE 21, 29S, 43W	-102.02	37.51			to 75 m	1991	3/94
Finney	SE SW SW 9, 24S, 33W	-100.95	37.97			to 146 m	1991	3/94

Methodology of heat-flow determination

Temperature profiles from several boreholes and from different logging studies (table 1) are the basis for the determination of interval temperature gradients and heat flow. Most boreholes were measured in summers of 1980 and 1981 (see Stavnes, 1982; Steeples and Stavnes, 1982) using a thermistor probe with an inherent accuracy of ±0.2°C. Temperatures were obtained at regular depth intervals. The logged intervals range between 1.5 m (4.9 ft) in shallow holes less than 125 m (410 ft), 3.0 m (9.8 ft) in medium and deep holes up to 375 m (1,230 ft), and 4.6 m (15 ft) in air-filled holes. Temperature logs for the boreholes Gibs (identical with well description 1CN2 used by Stavnes), Finegan, Rooks (1RO1 by Stavnes), and Rush (1RH2 by Stavnes) were provided by D. D. Blackwell (personal communication, 1994). Temperatures for these boreholes also were measured with a thermistor probe to the nearest 0.001°C at intervals of 2 m (6.5 ft) and 5 m (16.4 ft). Additional logs were measured by the authors in February 1994 using the Distributed Optical Fibre Temperature Sensing technique (DTS), which allows a continuous temperature recording along the profile after the installation of the tool (fiber optic cable) in the borehole (for details see Hurtig et al., 1994; Förster et al., 1996). This technique has a resolution of ±0.1°C, a precision of about 0.3°C, and a sample interval of 1.0 m (3.3 ft).

The wells logged by Stavnes had been completed by drilling at least three weeks prior to the logging date to allow temperatures to return to equilibrium. The thermal logs obtained in the boreholes 1CN2/Gibs, Finegan, 1RO1/Rooks by D. D. Blackwell (personal communication, 1994) also probably represent thermal equilibrium because of the lapsed time from drilling; the same is true for the logs we obtained in the boreholes Hodgeman, Finney, and Stanton. The Stanton borehole, however, is very shallow, so the temperature profile may reflect seasonal changes.

The temperature profiles selected for this study show a general increase of temperature with depth. Although some of the wells are constructed as monitoring sites for the Dakota aquifer, hydrological disturbances in the studied boreholes are not substantial nor significant, especially given the long time between the last pumping test and the logging date, which was on the order of years. The problem of intrahole fluid flow is minor because all the boreholes considered in this study had small diameters, 2 or 5 inches (4.1 or 12.7 cm), and were cased.

Lithologic descriptions and stratigraphic details used in interpreting the temperature profiles were taken mostly from sample logs and wire-line logs. Many of the descriptions were either general or nondescript, so that it was not always possible to identify exactly the stratigraphic contacts from descriptions. For example, in places it was difficult to recognize the Graneros-Dakota contact or the Dakota-Kiowa contact, especially in sample logs, because the lithology of both units is similar. Also the lower part of the Carlile is similar to the upper part of the Greenhorn. Therefore, interval temperature gradients were computed from depth intervals with fairly homogeneous lithology, with the assumption that these units also are characterized by a fairly homogeneous thermal conductivity. In some situations, the stratigraphic unit (group) had to be broken into two or more intervals according to lithologic changes. We also have adjusted some of the stratigraphic contacts to better fit the gradient data, which in many instances may be more reliable than the stratigraphic descriptions.

Thermal conductivity measurements were made in the laboratory of the Bundesanstalt für Geowissenschaften und Rohstoffe in Hannover (Germany) on cores from two boreholes, the Guy F. Atkinson No. 1 Beaumeister well in Cheyenne County and the Stanton County well (table 2). The measurements were carried out on both dry and saturated samples using a divided bar apparatus. The thermal-conductivity values determined have an accuracy of 3%.

In the remaining boreholes with thermal logging data (table 1), however, thermal-conductivity data were lacking. Therefore, the interpretation of interval thermal gradients from the sequence overlying the Dakota Formation was limited to such lithology for which thermal conductivity measured elsewhere could be extrapolated into the study area. We assumed that, because of the relatively small variability in the thermal conductivity of shale in eastern Kansas and in the northern Great Plains (Gosnold et al., 1981; Sass and Galanis, 1983; Blackwell and Steele, 1989; Gosnold, 1990), shale (preferably dark, noncalcareous shale) could be used to evaluate thermal gradients and estimate heat-flow density. As those studies showed, although the Paleozoic shales are more indurated than the younger and softer Mesozoic shales, the thermal conductivity does not differ to a great extent.

For example, for Paleozoic shales showing porosities between 6% and 12% from four boreholes in central and eastern Kansas, the in situ thermal conductivity inferred from the heat-flow density of the underlying carbonates was in the range of 1.2 ± 0.1 W/mK (Blackwell et al., 1981; Blackwell and Steele, 1989). This value is consistent with data from seven wells penetrating the Paleozoic shales in Nebraska (Gosnold and Eversoll, 1981). For the Upper Cretaceous Pierre Shale, a thermal conductivity between 1.19 ± 0.05 W/mK (vertical component) and 1.38 ± 0.04 W/mK (horizontal component) was measured by Sass and Galanis (1983) on a preserved core sample from a well near Hayes, South Dakota, using the needle-probe technique. These results suggest the shales have only a modest compaction effect on thermal conductivity because dense lower Paleozoic shale have a thermal conductivity similar to Upper Cretaceous shale (see Blackwell and Steele, 1988). Gosnold's later study (1990) showed shale conductivity in the same range as obtained previously. He obtained facies-specific conductivity values for the Pierre Shale in Nebraska from differences in temperature gradients according to lithology. For the eastern facies (marine shales) of the Pierre Shale (Tourtelot, 1962), which extends from Nebraska into northwestern Kansas, a general increase in thermal conductivity from about 0.9–1.0 W/mK in the lower members (i.e., the Sharon Spring Member) to the higher members of the section was reported. Gosnold also observed that those low conductiv-

TABLE 2. Thermal-conductivity data from two Dakota cores measured at different temperatures. Borehole 1 = Guy F. Atkinson No. 1 Beaumeister well, SE SE NE sec. 31, T. 2 S., R. 39 W. in Cheyenne County; borehole 2 = Stanton Co. well, SE SW SE sec. 21, T. 29 S., R. 43 W.

Borehole	Depth	Stratigraphy	Lithology	Temperature (˚C)	Thermal conductivity (W/mK) unsaturated sample	Temperature (˚C)	Thermal conductivity (W/mK) saturated sample
1	622.3 m	Dakota Formation	sandstone,	19.7	1.60		
		(D sequence)	fine grained	58.9	1.55		
1	626.7 m	Dakota Formation	siltstone,	20.1	1.36		
		(D sequence)	fine grained	59.4	1.18		
1	702.1 m	Dakota Formation	sandstone,	18.8	2.48	23.5	3.24
		(J sequence)	medium grained	58.1	2.31	52.6	3.06
2	12.2 m	Dakota Formation	sandstone,	9.8	2.27	27.0	3.12
		(D sequence)	fine grained	47.6	2.03	46.8	3.10
2	30.8 m	Dakota Formation	sandstone,	10.4	2.69	26.8	3.96
		(J sequence)	fine grained	47.6	2.47	46.6	3.97
2	36.6 m	Dakota Formation	sandstone,	10.2	2.55	27.8	3.64
		(J sequence)	medium to coarse grained	47.9	2.31	47.0	3.51
2	42.4 m	Kiowa Formation	interlaminated shale and	10.7	1.34	27.8	2.72
			sandstone, fine grained	58.9	1.26	47.5	2.70

ity values in the lower members of the Pierre Shale are about equal to the apparently equivalent thermal conductivity of the Carlile and Graneros Shales. The effective conductivity of the Pierre Shale can be accounted for by about 1.2 W/mK (eastern facies), whereas values of 1.1 W/mK for the Pierre Shale were reported to be typical in the Williston Basin and in southern South Dakota (Gosnold, 1990). Because the Pierre Shale is homogeneous and the shales in our area generally have a lithology similar to those with thermal-conductivity estimates outside the area, it is assumed that a thermal-conductivity value of 1.1–1.2 W/mK is reliable for the shales investigated here.

Heat-flow Data

Geothermal gradients were computed for each logging interval and plotted versus depth (figs. 4 and 5). In general, temperature as a function of depth shows a relationship to lithology changes. However, because of the relatively large logging interval, the thin-bedded nature of the sequence (especially in the Carlile Shale and Greenhorn Limestone), and the small thickness (Graneros Shale), it is difficult to generalize the results in terms of formation gradients. Table 3 lists the interval temperature gradients summarized for lithotypes within the stratigraphic groups, the thermal conductivity assigned to each lithotype, and the heat-flow density obtained.

In general, the noncalcareous, marine shales (Pierre, Carlile, Graneros, and Kiowa), for which a thermal conductivity of 1.2 W/mK was assigned, had different gradients. The gradients range from 53.2 to 65.8°C/km (average 58.5°C/km) in the Pierre Shale, from 51.0 to 60.5°C/km (average 55.5°C/km) in the Carlile Shale, and from 45.0 to 49.7°C/km in the Graneros Shale and Kiowa Formation. Regionally, there is a decrease of temperature gradients to the east. It also is evident that the temperature gradients in the noncalcareous, marine shales are higher than in those shales with a higher silt content—for example, in the Graneros Shale (range of 36.7 to 45.2°C/km, average 40.8°C/km) and in the Kiowa Formation (range of 34.5 to 43.0°C/km, average 39.9°C/km). In comparison,

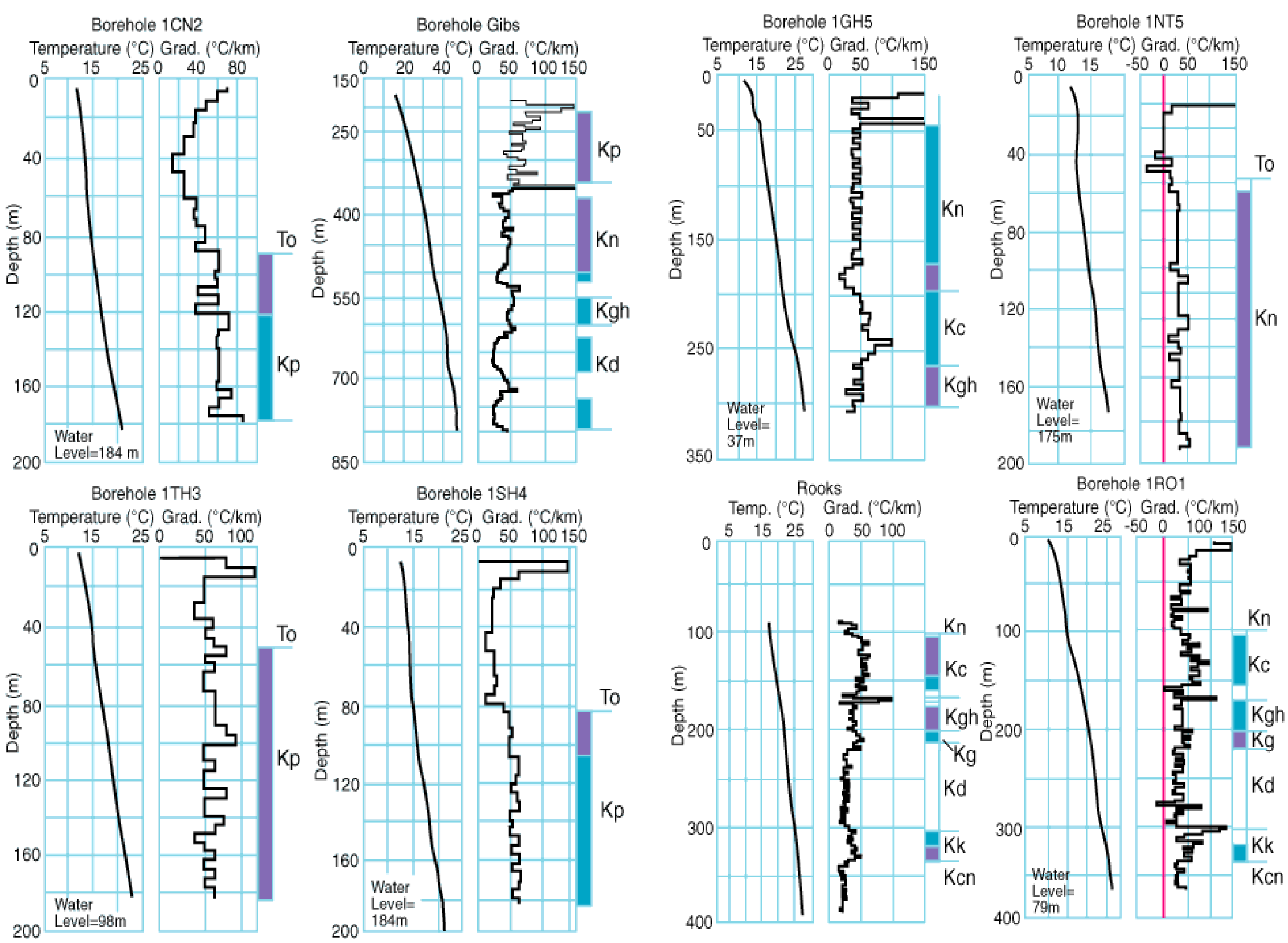

FIGURE 4. Temperature-depth and gradient-depth curves. Depth intervals used for estimation of interval gradients (striped boxes) and stratigraphy on group level also are shown. *(Continued on next page.)*

the gradients in the calcareous or chalky shales of the Niobrara Formation have an average value of 40.6°C/km, nearly equal to the average gradient of 41.0°C/km in the alternating beds of limestone, shale, and calcareous shale of the Greenhorn Limestone.

Based on the interval gradients, estimates of heat-flow density in the marine, noncalcareous shales in the Pierre Shale range between 63.8 and 79.0 mW/m² and in the Carlile Shale between 61.2 and 72.6 mW/m². Unfortunately, no interval gradients from shale units below the Pierre were available to substantiate the heat-flow value. In the Finegan borehole (fig. 5), however, a thermal gradient was obtained in a sandstone of the Dakota Formation. This interval gradient is related to a thermal conductivity of 3.1 W/mK, as measured on a sample from the Guy F. Atkinson No. 1 Beaumeister well (table 2), where the heat-flow density is 72.8 mW/m². The average value for the Finegan borehole site is 69.0 ± 1.8 mW/m². The highest heat-flow values in the Pierre Shale come from the borehole 1CN2/Gibs (average 73.6 ± 2.4 mW/m²). Two different logs were run at this site (table 3): log 1CN2 was measured under air conditions in the borehole to a depth of 179 m, whereas for log Gibs, because of a faster logging rate, the first useable results are below the water table at 190 m. The gradients for the Pierre Shale from the 1CN2/Gibs borehole differ slightly, yielding heat-flow values ranging from 79.0 mW/m² to 75.2 mW/m² (table 3). As with the Finegan borehole, in the 1CN2/Gibs borehole a thermal gradient was obtained in a sandstone of the Dakota Formation (fig. 4). If a thermal conductivity of 3.1 W/mK, as measured on a sample from the Guy F. Atkinson No. 1 Beaumeister well (table 2), is related to the thermal gradient, then the heat-flow density is 71.9–73.8 mW/m², which is in the range indicated by the shales.

Only two boreholes, 1RO1/Rooks and Finney, penetrate different shaly units below the Pierre Shale down to the Kiowa Formation. The two logs measured at the 1RO1/Rooks site on different logging dates (table 1) are interpreted starting below the water table. In the Carlile Shale, in this borehole the heat-flow values measured are 61.7 mW/m² (Rooks) and 71.6 mW/m² (1RO1). Below the Carlile, in the Graneros Shale, heat-flow values are lower (54.2 to 58.8 mW/m²), whereas the average heat-flow density at the 1RO1/Rooks borehole site is 61.0 ± 2.6 mW/m². In the Finney borehole, where heat-flow values are obtained in shales (Graneros Shale and Kiowa Formation) and in a sandstone of the Dakota Formation, the average heat-flow density is 53.3 ± 2.6 mW/m².

At borehole sites 1NT5, 1TH3, 1GH5, 1EL1, Rush, 1HG1, 1FO1, and Hodgeman, interval thermal gradients for shales are estimated only in one stratigraphic unit.

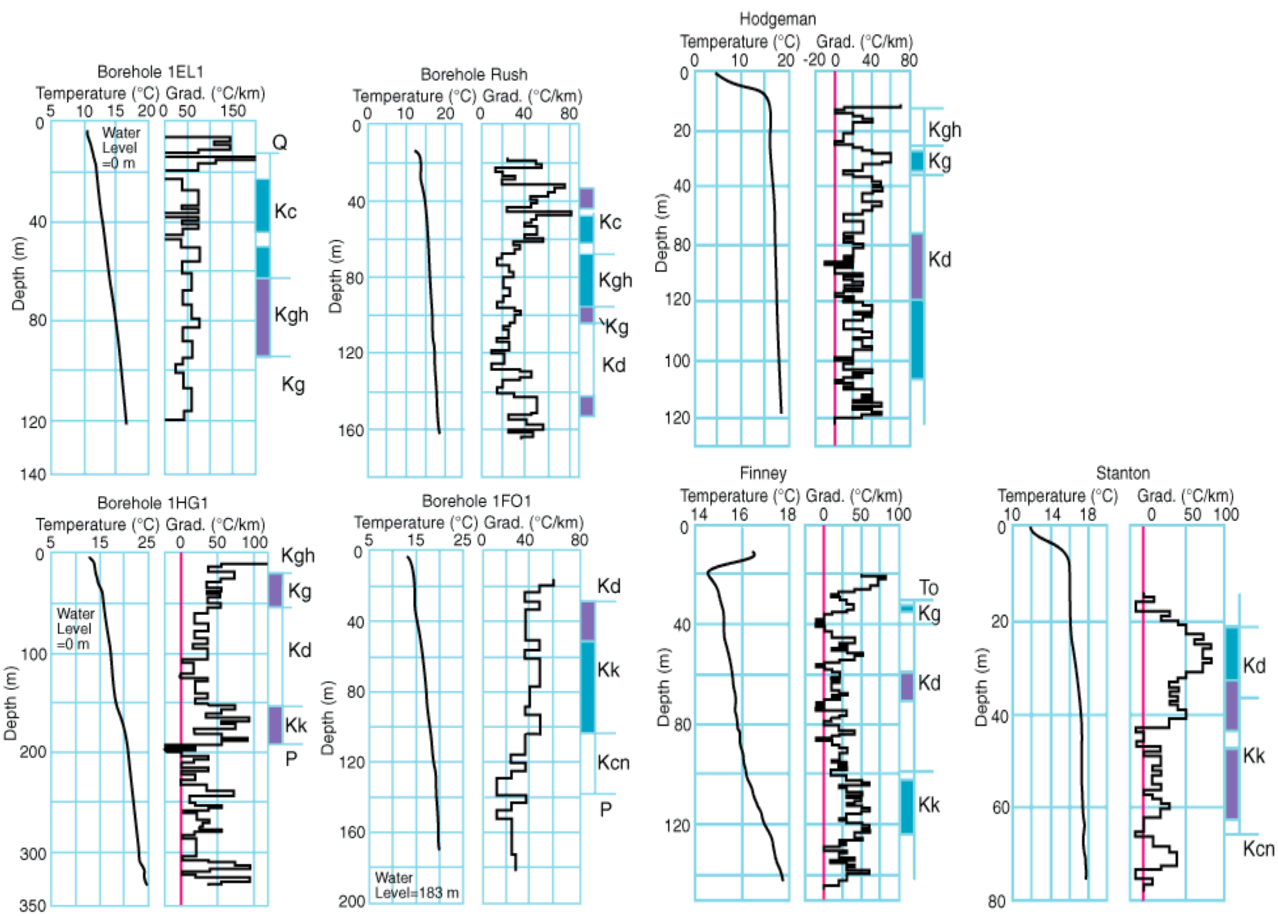

FIGURE 4. *(Continued from previous page.)*

Consequently, heat-flow values from these boreholes are less reliable. For the boreholes 1GH5, 1EL1, Rush, and 1FO1, however, the unreliability in heat-flow estimation can be minimized because the temperature gradients in lithologic units other than shales are within the expected range of values typical for those rocks. Because of the limited depth of the Stanton borehole, interval gradients determined from the temperature log are listed only and not used for further interpetation.

Summary and Conclusions

1. The heat-flow data obtained in this study (table 3, fig. 6) show for the most part that the Dakota aquifer is under fairly normal geothermal conditions similar to other parts of the stable Midcontinent Platform. The heat-flow density in Kansas, as shown on the Geothermal Map of North America (Blackwell and Steele, 1992), ranges mostly from 50 to 60 mW/m^2. The values observed in the outcrop area are in the order of 50 to 65 mW/m^2. The values are scattered regionally and no general trend is apparent, athough the heat-flow density in the southern part of the aquifer may be slightly lower than in the central and eastern part. To what extent this could be controlled by the water recharge to the system remains to be investigated.

2. Slightly higher heat-flow density values occur in the extreme northwestern corner of the study area, with the highest value of 73.6 ± 2.4 mW/m^2 occurring in the 1CN2/Gibs borehole and a relatively high value in the 1GH5 borehole (72.6 ± 2.1 mW/m^2). The heat-flow density data reported and mapped for western Colorado and the Nebraska Panhandle region on the western flank of the Chadron Arch also are in this range. Regional differences and inhomogeneities in the heat-flow data of northwestern

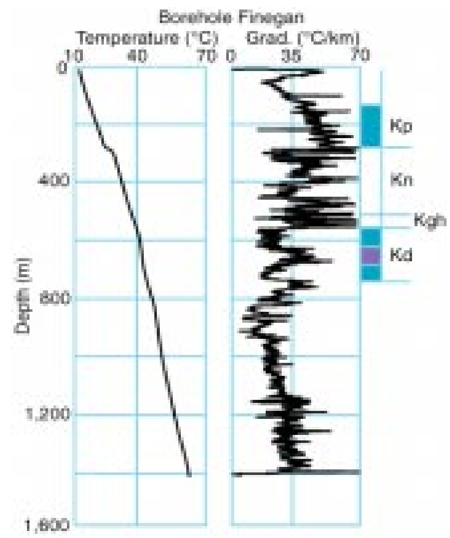

FIGURE 5. Temperature-depth and gradient-depth curves for Finegan borehole.

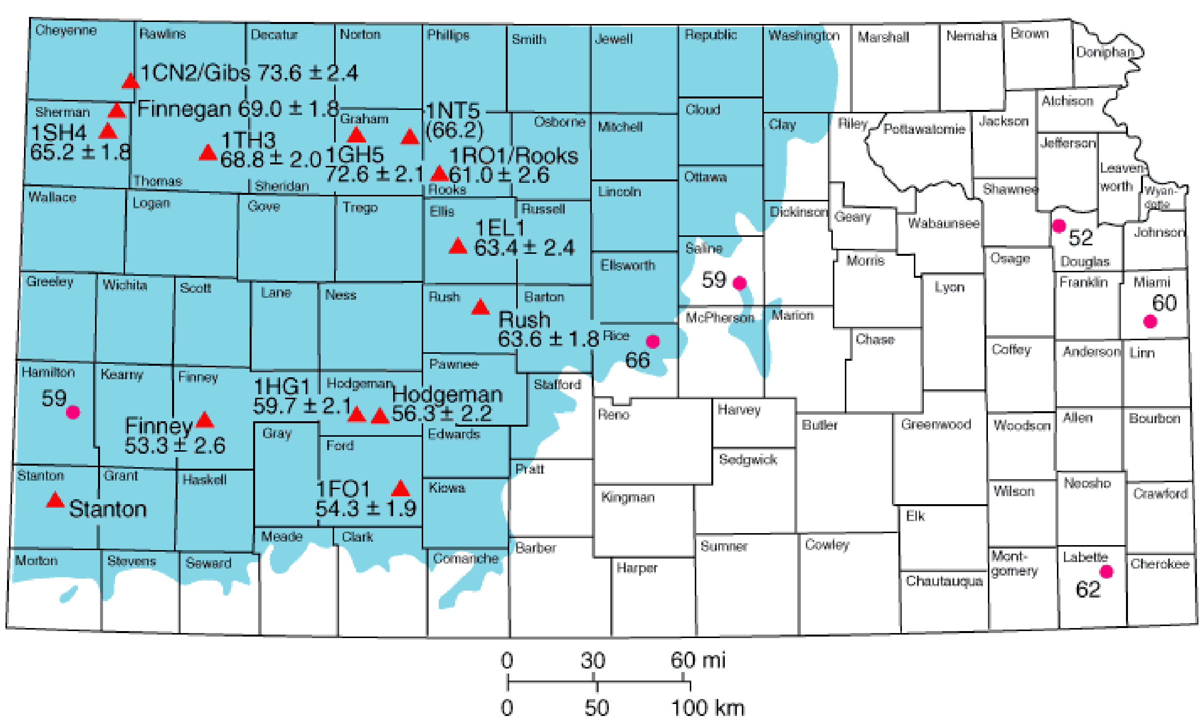

FIGURE 6. Location of boreholes discussed in this paper (shown as triangles). Numbers are estimates of heat-flow density. Standard errors are listed with values. Circles represent borehole sites and heat-flow density values measured by different authors (Blackwell and Steele, 1989). Extent of Dakota aquifer also shown.

TABLE 3. Interval temperature gradient, thermal conductivity, and heat-flow density for stratigraphic units. Thermal-conductivity values in parentheses are inferred from heat flow estimated in noncalcareous, marine shales in other depth intervals of borehole sites.

Unit	Lithology	Temperature gradient (˚C/km)	Thermal conductivity (W/mK)	Heat-flow density (mW/m²)	Borehole
Pierre Shale (Kp)	silty shale	53.6 ± 3.3	(1.4)	(75.2)	1CN2
	shale	62.7 ± 2.6	1.2	75.2 ± 1.8	1CN2
	shale	65.8 ± 4.0	1.2	79.0 ± 2.2	Gibs
	shale	56.7	1.2	68.0	Finegan
	silty shale	53.2 ± 2.4	1.2	63.8 ± 1.7	1SH4
	shale	55.5 ± 2.4	1.2	66.6 ± 1.7	1SH4
	shale	57.3 ± 3.3	1.2	68.8 ± 2.0	1TH3
Niobrara Formation (Kn)	chalky shale	43.0 ± 4.4	(1.75–1.8)	(75.2–79.0)	Gibs
	limestone	29.2 ± 1.3	(2.6–2.7)	(75.2–79.0)	Gibs
	chalky shale	36.8 ± 3.0	(1.8)	(66.2)	1NT5
	chalky shale	42.0 ± 2.4	(1.73)	(72.6)	1GH5
	limestone	26.2 ± 2.8	(2.77)	(72.6)	1GH5
Carlile Shale (Kc)	shale	60.5 ± 3.7	1.2	72.6 ± 2.1	1GH5
	shale	59.7 ± 3.8	1.2	71.6 ± 2.1	1RO1
	shale	51.4 ± 2.1	1.2	61.7 ± 1.6	Rooks
	shale	51.0 ± 4.7	1.2	61.2 ± 2.4	1EL1
	shale	54.7 ± 4.3	1.2	65.6 ± 2.3	1EL1
	shale	53.0 ± 2.8	1.2	63.6 ± 1.8	Rush
	chalky shale	47.1 ± 2.3	(1.35)	(63.6)	Rush
Greenhorn Limestone (Kgh)	limestone/shale	49.7 ± 2.2	(1.45–1.59)	(72–79)	Gibs
	limestone/shale	39.6 ± 3.0	(1.83)	(72.6)	1GH5
	limestone/shale	32.8 ± 2.5			1RO1
	limestone/shale	34.3 ± 2.0			Rooks
	calcareous shale	48.6 ± 3.5	(1.3)	(63.1)	1EL1
	limestone	22.9 ± 2.4			Rush
Graneros Shale (Kg)	silty shale	45.2 ± 2.9	1.2–1.3	54.2–58.8 ± 1.9	1RO1
	silty shale	40.6 ± 2.2	(1.4)	(58.0)	Rooks
	shale	49.7 ± 3.7	1.2	59.7 ± 2.1	1HG1
	shale	45.0 ± 3.6	1.2-1.3	54.0–58.5 ± 2.1	Hodgeman
	silty shale	36.7 ± 2.1	(1.4)	(51.4)	Finney
	silty shale	31.3 ± 1.5			Rush
Dakota Formation (Kd)	sandstone	23.8 ± 1.6	3.1	73.8 ± 2.2	Gibs
	sandstone	23.2 ± 2.2	3.1	71.9 ± 2.6	Gibs
	sandstone	23.5	3.1	72.8	Finegan
	sandy shale	41.6			Finegan
	siltstone/sandstone	21.0	3.1	65.1	Finegan
	shale	65.5 ± 3.3	1.1–1.2	72.0–78.6 ?	Stanton
	sandstone	35.0 ± 2.2	3.6	126 ?	Stanton
	sandstone	16.4 ± 2.5	3.2	52.5 ± 2.8	Finney
	sandstone	25.0 ± 3.4			Hodgeman
	sandy shale	49.0 ± 1.4			Rush
Kiowa Formation (Kk)	shale/sandstone	46.7 ± 2.2			Stanton
	siltstone	15.3 ± 2.8			Stanton
	shale	49.0 ± 3.5	1.2	58.8 ± 2.0	1RO1
	sandstone	29.2 ± 1.9			Rooks
	silty shale	34.5 ± 2.4			Rooks
	shale/siltstone	58.6 ± 4.4 ?	1.2	70.3 ?	1HG1
	silty shale	40.5 ± 2.4	1.3	52.6 ± 1.8	1FO1
	silty shale	43.0 ± 2.4	1.3	55.9 ± 1.8	1FO1
	silty shale	41.6 ± 3.5	1.3	54.0 ± 2.1	Finney

Kansas are seemingly caused by differences in the make up of the aquifer and variations in the porosity and permeability.

3. Temperature effects linked with the regional water flow out of the Denver Basin are not striking on the shallow basin flank in western Kansas. This also could be evidence for a mingling of different water flows in the area, one source coming from the recharge areas in the south and east, with another coming eastward out of the basin. This could be substantiated by thermal,

geochemical, and mineralogical analyses in boreholes to be drilled in the area in future.

Acknowledgments

We thank the GeoForschungsZentrum Potsdam and the Kansas Geological Survey for support of this study. Christoph Grissemann and Günther Koß of the Bundesanstalt für Geowissenschaften und Rohstoffe Hannover (Germany) performed the laboratory measurements of thermal conductivity. We also thank Al Macfarlane, Don Steeples, and Charles Barker, David Blackwell, and John Sass for comments and suggestions on improvement of this paper.

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