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THERMAL STATE OF THE ARKOMA BASIN AND THE ANADARKO BASIN, OKLAHOMA

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THERMAL STATE OF THE ARKOMA BASIN AND
THE ANADARKO BASIN, OKLAHOMA

A DISSERTATION APPROVED FOR THE
SCHOOL OF GEOLOGY AND GEOPHYSICS

BY

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Foreword

One of the most fundamental physical processes that affects virtually all geologic phenomena in sedimentary basins is the flow of heat from the Earth's interiors, its variation through space and time, and the resulting temperature history of sediments within a basin. In turn, temperature of a sedimentary basin affects tectonic activity and evolution of a sedimentary basin (McKenzie, 1978), geochemical processes such as hydrocarbon maturation and migration (Tissot et al., 1987; Mackenzie and Quigley, 1988), diagenesis and ore deposits (e.g., Mississippi Valley-type lead-zinc deposits, Garven et al., 1993; Deming and Nunn, 1991; Bethke, 1986), paleo-climate change (e.g., Deming and Borel, 1995; Harris and Chapman, 1995), and sedimentary basin scale ground water flow (e.g., Deming, 1993).

The Arkoma Basin and the Anadarko Basin, Oklahoma, are a prolific producer of both oil and natural gas. In addition to being important oil and gas provinces, both basins have important geologic phenomena. Various researchers have invoked the movement of hot fluids out of the Arkoma Basin during the Ouachita Orogeny in Late Pennsylvanian-Early Permian time (~290 Ma) to explain the genesis of Mississippi Valley-type (MVT) lead-zinc deposits, paleothermal anomalies, and regional diagenesis in the North American midcontinent. Because subsurface temperature is sensitive to regional groundwater movement, the study of the thermal state of the Arkoma basin can provide insight into the nature of regional groundwater flow systems to explain the genesis of MVT ore deposits.

The Anadarko Basin is known to contain abnormally high fluid pressures, although the Anadarko Basin is an old and tectonically senescent basin (Hunt, 1990; Al-Shaieb et al., 1994). If oil and gas generation is a main cause of overpressures in the Anadarko Basin, the rate at which fluid pressuring occurred in the Anadarko Basin may be temperature-dependent.

Understanding the thermal state of the Arkoma Basin and the Anadarko Basin is therefore crucial to understanding the timing and extent of hydrocarbon generation, past fluid migrations to explain the genesis of Mississippi Valley-type ore deposits, and the origin of overpressures in the Anadarko Basin.

This dissertation consists of four separate chapters concerning different topics in terrestrial heat flow and overpressuring in sedimentary basins. Chapter one, "Heat flow and heat production in the Arkoma Basin and Oklahoma Platform, southeastern Oklahoma", was published in *Journal of Geophysical Research* vol. 101 in 1996. Chapter two, "Evaluation of thermal conductivity temperature corrections applied in terrestrial heat flow studies", was published in *Journal of Geophysical Research* vol. 103 in 1998. Chapter three, "Heat flow and thermal history of the Anadarko Basin and the western Oklahoma Platform" is now in press in *Tectonophysics*. Finally, chapter four, "Overpressures in the Anadarko Basin, southwestern Oklahoma: static or dynamic ?" will be submitted to *American Petroleum Geologist Bulletin.*

In chapter one, heat flow and heat production in the Arkoma basin and Oklahoma Platform are discussed. Estimated heat flow (±20%) in the deep part of the Arkoma Basin near the Ouachita Front is 35-40 mW/m^2 and increases systematically northward to 60-65 mW/m^2 on the Oklahoma Platform. Average heat production, estimated from gamma ray logs, is 2.3±0.2 μW/m^3 for basement rocks underlying the Arkoma Basin and 2.8±0.1 μW/m^3 for basement rocks in the Oklahoma Platform area. Numerical models show that heat refraction from the less conductive sedimentary rocks (~1.6 W/m-K) of the Arkoma Basin to the more conductive crystalline rocks (~3.0 W/m-K at 25°C) of the Oklahoma Platform and the Ouachita Mountains accounts for about 5-10 mW/m^2 of the observed 20-30 mW/m^2 decrease in heat flow from north to south. Changes in crustal heat production related to compositional changes and crustal thinning account for another 5-15 mW/m^2 of the observed heat flow change. If the remaining 0-20 mW/m^2 difference in heat flow is attributed to heat transport by topographically driven groundwater flow, the average basin-scale permeability of the Arkoma Basin and the Oklahoma Platform can be no greater than 10^{-15} m^2. Results of this study are not generally supportive of theories which invoke topographically driven regional groundwater flow from the Arkoma Basin in Late Pennsylvanian-Early Permian time (~290 Ma) to explain the genesis of Mississippi Valley-type lead-zinc deposits, paleothermal anomalies, and regional diagenesis in the North American midcontinent.

In chapter 2, different types of thermal conductivity temperature corrections that are commonly applied in terrestrial heat flow studies are evaluated using 117 sets of temperature-dependent thermal conductivity measurements on minerals and rocks. When accuracy was bench-marked against the entire data set, corrections proposed by *Sekiguchi* [1984], *Somerton* [1992], *Sass et al.* [1992], *Funnell et al.* [1996], and *Chapman et al.* [1984] were found to have mean absolute relative errors of 6.9%, 7.7%, 8.5%, 14.2%, and 17.6%, respectively. The correction proposed by *Somerton* [1992] was found to have the lowest mean absolute relative error for minerals (11.3%) and sedimentary rocks (6.5%), while the correction proposed by *Sekiguchi* [1984] was found to have the lowest mean absolute relative error for igneous (5.4%) and metamorphic rocks (4.9%). The correction schemes were also found to differ in their systematic errors. Corrections proposed by *Sekiguchi* [1984], *Somerton* [1992], *Sass et al.* [1992], *Funnell et al.* [1996], and *Chapman et al.* [1984] were found to have mean relative errors of -3.3%, +2.6%, -5.8%, -11.0%, and +15.5%, respectively. Most of the temperature-dependent thermal conductivity data used to evaluate correction schemes were measurements on air-saturated rocks; however, the invariance of the relative rankings with respect to rock porosity suggests the rankings may be valid with respect to *in situ* conditions.

Chapter three addresses heat flow and thermal history of the Anadarko Basin and the western Oklahoma Platform. The average geothermal gradient in the Anadarko Basin and the Oklahoma Platform estimated from 856 corrected bottom-hole temperatures (BHTs) is 21 °C/km. Analysis of previously published thermal maturation data indicates that the Anadarko Basin has undergone from 1 to 3 km of erosion starting about 40 to 50 Ma. The average thermal gradient at the time of maximum burial was in the range of 22 to 25 °C/km. There is no evidence that the thermal state of the Anadarko Basin has changed significantly since the late Paleozoic (250 Ma).

Estimated heat flow (±20%) ranges from 30 mW/m^2 to 42 mW/m^2, with a mean of 36 mW/m^2. We found no evidence for heat flow to increase significantly from the Anadarko Basin in the south to the Oklahoma Platform to the north. These results contradict a previous study which found heat flow increases from south to north, with heat flow in the Oklahoma Platform as much as 50% greater than heat flow in the Anadarko Basin. At the present time, we do not know if these differences are related to spatial variations or reflect methodological errors. If our results are averaged with those reported in previous work, mean heat flows in the Anadarko Basin and Oklahoma Platform are 39 mW/m^2 and 51 mW/m^2, respectively.

In chapter four, overpressures in the Anadarko Basin, southwestern Oklahoma are discussed. In the Anadarko Basin, fluid pressures above hydrostatic start at a depth of about 2.5 km. On a normalized scale where hydrostatic pressure = 0 and lithostatic pressure = 1, estimates of fluid pressures in the Anadarko Basin were found to range from 0.4 to 0.8 over a depth range of 4 to 7 km.

Using scale analyses and a simple numerical model, we evaluated two end-member hypotheses (compaction disequilibrium and hydrocarbon generation) as possible causes of overpressuring in the Anadarko basin. If compaction disequilibrium is the primary cause of present-day overpressures in the Anadarko Basin, the Anadarko Basin is required to have an average permeability of 10^{-23} m^2 or lower. If geopressures in the Anadarko Basin result from hydrocarbon generation, average basin permeabilities can be as high as 10^{-20} m^2. Scaling these average permeabilities down to thicknesses of 100 m or less implies permeabilities lower than 10^{-25} m^2 are required in the most optimistic scenario. The lowest permeabilities ever measured on sedimentary rocks are in the neighborhood of 10^{-22} to 10^{-23} m^2. Thus is not possible to reconcile the existence of overpressures in the Anadarko Basin with classic hydrodynamic theories which maintain that aquicludes do not exist. Either some unknown process is generating excess fluid pressures, or the Anadarko Basin contains pressure seals.

Geopressure models which invoke compaction disequilibrium and are commonly applied to Cenozoic basins undergoing rapid sedimentation (e.g., Gulf Coast Basin of southeastern U.S.) do not appear to apply to the Anadarko Basin. The Anadarko Basin belongs to a group of cratonic basins which are tectonically quiescent and are characterized by the association of abnormal pressures with natural gas.

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CHAPTER 1

Heat flow and heat production in the Arkoma Basin and Oklahoma Platform, southeastern Oklahoma

1. Abstract

Subsurface temperature and thermal gradients along a north-south cross section through the Arkoma Basin and the Oklahoma Platform in southeastern Oklahoma were estimated from 345 bottom hole temperatures from 199 oil and gas wells. The average geothermal gradient in the southern part of the basin near the Ouachita Front is 20 °C/km, exceeds 30 °C/km in the middle part of the basin, and is 24 °C/km on the Oklahoma Platform to the north. Drill cuttings obtained from 11 oil and gas wells were used for 843 thermal conductivity measurements. Thermal conductivity data, corrected to *in situ* conditions, were used to estimate heat flow. Estimated heat flow (±20%) in the deep part of the Arkoma Basin near the Ouachita Front is 35-40 mW/m^2 and increases systematically northward to 60-65 mW/m^2 on the Oklahoma Platform. Average heat production, estimated from gamma ray logs, is 2.3±0.2 μW/m^3 for basement rocks underlying the Arkoma Basin and 2.8±0.1 μW/m^3 for basement rocks in the Oklahoma Platform area. Numerical models show that heat refraction from the less conductive sedimentary rocks (~1.6 W/m-K) of the Arkoma Basin to the more conductive crystalline rocks (~3.0 W/m-K at 25°C) of the Oklahoma Platform and the Ouachita Mountains accounts for about 5-10 mW/m^2 of the observed 20-30 mW/m^2 decrease in heat flow from north to south. Changes in crustal heat production related to compositional changes and crustal thinning account for another 5-15 mW/m^2 of the observed heat flow change. If the remaining 0-20 mW/m^2 difference in heat flow is attributed to heat transport by topographically driven groundwater flow, the average basin-scale permeability of the Arkoma Basin and the Oklahoma Platform can be no greater than 10^{-15} m^2. Results of this study are

not generally supportive of theories which invoke topographically driven regional groundwater flow from the Arkoma Basin in Late Pennsylvanian-Early Permian time (~290 Ma) to explain the genesis of Mississippi Valley-type lead-zinc deposits, paleothermal anomalies, and regional diagenesis in the North American midcontinent.

2. Introduction

The Arkoma basin (Figure 1) is one of the most active natural gas exploration areas in North America [*Houseknecht et al.*, 1992]. However, the timing of gas generation is poorly understood because the thermal history of the Arkoma Basin is not well known.

Many studies have been done of coal rank and vitrinite reflectance [*Houseknecht and Matthews*, 1985; *Cardott et al.*, 1986; *Underwood et al.*, 1988; *Houseknecht and McGilvery*, 1990; *Houseknecht et al.*, 1992; *Houseknecht and Spötl*, 1993; *Spötl et al.*, 1993], but these data have not yet been integrated into a comprehensive thermal history model. Interpretation of the thermal history of the Arkoma Basin and Ouachita Mountains varies with study areas and stratigraphic units. Vitrinite reflectance of the Hartshorne coal bed in the Arkoma Basin increases from 0.6% in the west to 2.0% in the east [*Houseknecht et al.*, 1992; *Houseknecht and Spötl*, 1993], and vitrinite reflectance of Carboniferous rocks in the Ouachita Mountains increases from 1.0% in the west to 4.0% in the east [*Houseknecht and Matthews*, 1985]. *Houseknecht and Matthews* [1985] suggested that relatively high vitrinite reflectance in the eastern part of Ouachita Mountains could be attributed to Mesozoic intrusive activity near the eastern end of the Ouachita Mountains, while lower thermal maturity

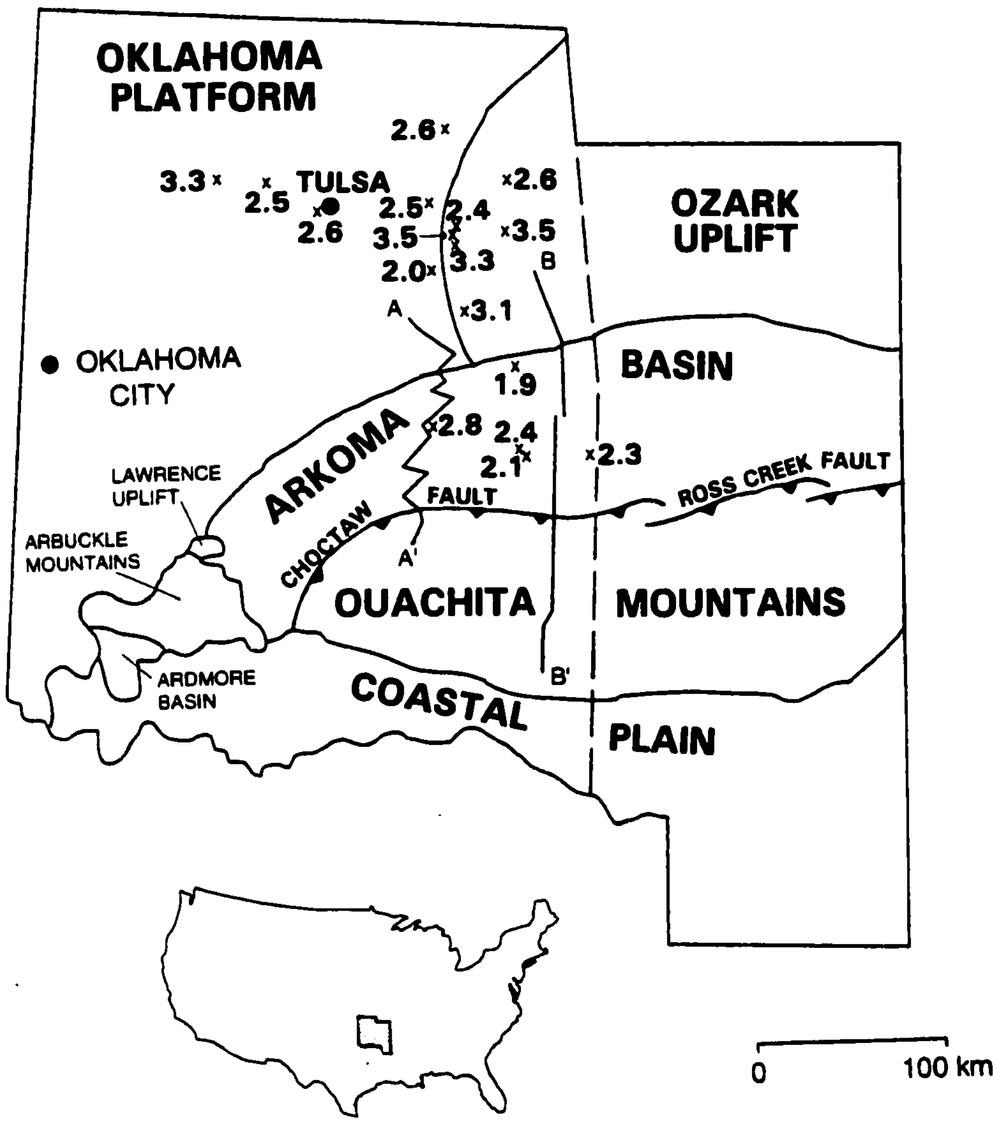


Figure 1. Arkoma Basin of Oklahoma and Arkansas and surrounding geological provinces [after *Johnson et al.*, 1988]. Locations of heat production estimates (crosses) from gamma ray logs, and values in $\mu W/m^3$ are shown.

in the western part of Ouachita Mountains was simply related to sedimentary and tectonic burial. At the present time it is unclear if the observed pattern of increasing thermal maturity to the east simply reflects differential removal of overburden or is due to higher geothermal gradients in the east at some time in the past.

Various researchers have invoked the movement of hot fluids out of the Arkoma Basin during the Ouachita Orogeny in Late Pennsylvanian-Early Permian time (~290 Ma) to explain the genesis of Mississippi Valley-type lead-zinc deposits, paleothermal anomalies, and regional diagenesis in the North American midcontinent [*Oliver*, 1986; *Leach and Rowan*, 1986; *Bethke and Marshak*, 1990; *Leach et al.*, 1991; *Sverjensky and Garven*, 1992; *Cathles*, 1993; *Walton et al.*, 1995]. Paleothermal indicators (e.g., fluid inclusions) and reconstructions of the known geologic section indicate anomalously high temperatures at relatively shallow burial depths throughout much of southeast Kansas and southwest Missouri, 200-300 km from the hypothetical fluid source in the Arkoma Basin. For example, *Walton et al.* [1995] reported that sandstones and limestones in southeastern Kansas underwent diagenesis at temperatures as high as 100-150°C at depths of 1400-1800 m or less. Similarly, reviews of fluid inclusion data from the Northern Arkansas, Tri-State, Viburnum Trend, and Central Missouri ore districts indicate temperatures in the range 80-150°C at depths that are believed to generally have been about 500 m but not greater than 1500 m [*Sverjensky*, 1986; *Shelton et al.*, 1992; *Leach and Rowan*, 1993]. These apparently anomalous thermal conditions are commonly attributed to the invasion of hot, migrating brines in an integrated model that relies upon orogenically generated regional fluid flow from the Arkoma Basin to simultaneously explain ore deposits, diagenesis, and thermal anomalies.

Mathematical models of fluid migration have largely relied upon topographically driven flow [*Bethke and Marshak*, 1990; *Deming and Nunn*, 1991; *Ge and Garven*, 1992; *Garven et al.*, 1993] or episodic, compaction-driven flow [*Sharp*, 1978; *Cathles*, 1993]. Both mechanisms are able to satisfy the thermal constraints imposed by paleothermal indicators but only under certain conditions. For example, topographically driven flow requires high heat flows and relatively high permeabilities. The episodic release mechanism requires a physical mechanism to hold in brines and release them in pulses, along with spatial focusing of flow to conserve the limited fluid available.

The present-day thermal state of the Arkoma Basin is relevant to achieving an improved understanding of both hydrocarbon maturation and past fluid migrations. Nearly half (~40%) of the surface heat flow on continents comes from the decay of radioactive elements in the upper crust [*Pollack and Chapman*, 1977] and is relatively constant over hundreds of millions of years due to the long half-lives of U, Th, and K (order 10^9 years). The present-day thermal state may also reveal geologic processes such as groundwater flow that conceivably may operate over hundreds of millions of years. Thus a study of present-day heat flow along with measurements of thermophysical properties forms a logical starting point for a synthesis of the thermal history of the Arkoma Basin.

A comprehensive survey of the thermal state of the Arkoma Basin would entail the collection and analysis of thousands of bottom hole temperatures (BHTs) from throughout the basin, as well as thousands of thermal conductivity measurements and analysis of a multitude of geophysical well logs containing information relevant to the thermophysical properties of the Arkoma Basin rocks (e.g., porosity, radioactivity). Such a study is beyond the scope of our present investigation. Instead, we have chosen to do a detailed thermal analysis along one geological cross section perpendicular to the strike of the basin structure (Figures 1 and 2). Although our analysis thus arguably fails to reveal some lateral variations that may be present, it does provide 11 new heat flow and 17 new heat production values for an important area from which such data were previously sparse to nonexistent and also is informative of the extent to which processes such as heat refraction and groundwater flow may be important determinants in both the past and present thermal state of the Arkoma Basin.

3. Geologic Setting

The Arkoma Basin (Figure 1) is a peripheral foreland basin (Figures 3 and 4) associated with the Ouachita Fold Belt of southern Arkansas and southeastern Oklahoma. It is bounded to the north and northwest by the Ozark Uplift and the central Oklahoma Platform, respectively. To the south, the Choctaw Fault and the Ross Creek Fault demarcate the Arkoma Basin from the Ouachita Mountains (Figure 1) [*Johnson et al.*, 1988]. The following discussion is a brief synopsis of the geologic history of the Arkoma Basin; more detailed summaries are given by *Houseknecht et al.* [1983], *Zachry and Sutherland* [1984], *Houseknecht* [1986], *Johnson et al.* [1988], *Sutherland* [1988], and *Arbenz* [1989].

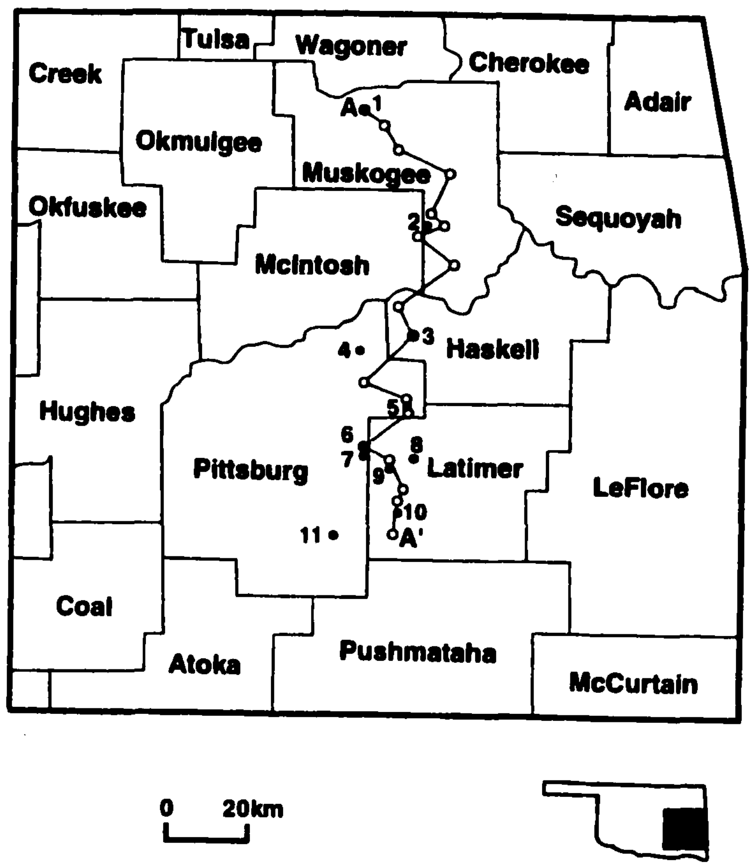


Figure 2. Location map showing cross section A-A' and 18 wells (open circles) used in construction of cross section [after *Johnson*, 1986], and 11 wells (solid circles) from which samples for thermal conductivity measurements were obtained.

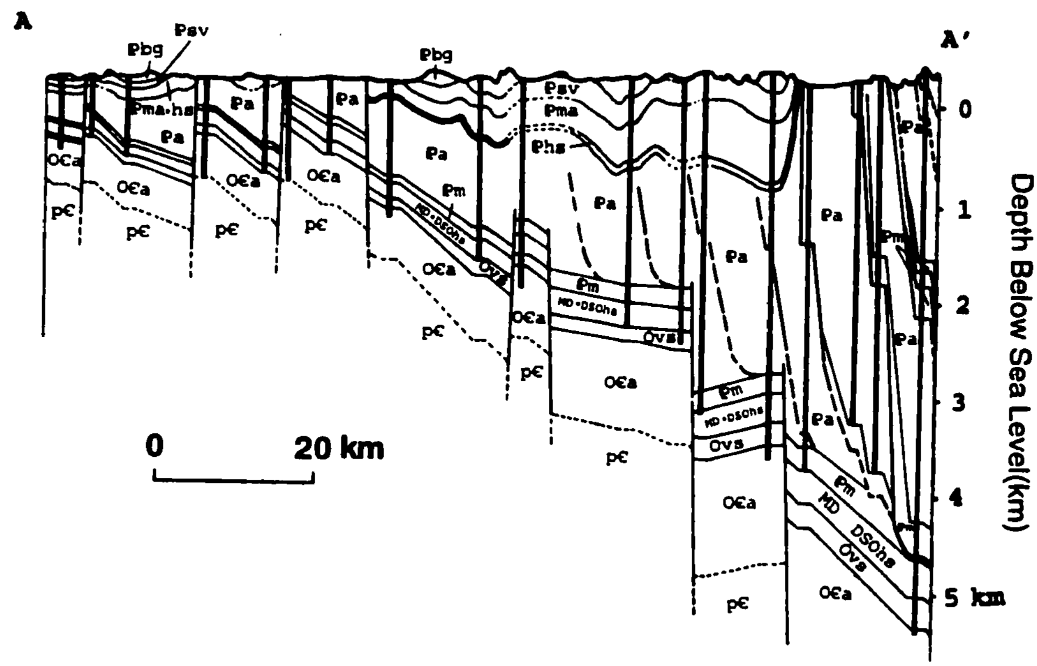


Figure 3. Cross section A-A' through the Arkoma Basin and the Oklahoma Platform in southeastern Oklahoma [after *Johnson*, 1986]. Unit symbols explained in Figure 4.

	SERIES		FORMATION	
PENNSYLVANIAN	Desmoinesian	Krebs Gp.	Boggy Fm.	Pbg
			Savanna Fm.	Psv
			McAlester Fm.	Pma
			Hartshorne: Upper / Lower	Phs
	Atokan		Atokan Fm.	Pa
	Morrowan		Wapanucka Ls. Limestone Gap Sh. Union valley Ls. Cromwell ss.	Pm
MISSISSIPPIAN	Chesterian		Goddard Fm.	MD
	Meramecian		Delaware Creek Sh. (="Caney")	
	Osagean			
	Kinderhookian		Welden Ls.	
DEVONIAN	Upper		Woodford Sh.	
	Lower	Hunton Gp.	Frisco Ls. Bois d'Arc Ls. Haragan Ls.	DSOhs
SILURIAN	Niagaran		Henryhouse Fm.	
	Alexandrian		Chimneyhill Subgroup	
ORDOVICIAN	Upper		Sylvan Sh.	
		Viola Gp.	Welling Fm. Viola Springs Fm.	Ovs
	Middle	Simpson Gp.	Bromide Fm. Tulip Creek Fm. McLish Fm. Oil Creek Fm. Joins Fm.	
	Lower	Arbuckle Gp.	West Spring Creek Fm. Kindblade Fm. Cool Creek Fm. Mckenzie Hill Fm. Butterly Dol.	O€a
CAMBRIAN	Upper		Singnal Mountain Ls. Royer Dol. Fort Sill Ls.	
		Timbered Hills Gp.	Honey Creek Ls. Reagan Ss.	
PRECAMBRIAN			Granite and Rhyolite	p€

Figure 4. Generalized stratigraphic column for the Arkoma Basin [after *Cardott et al.*, 1986].

The most widely accepted tectonic model for the genesis of the Arkoma Basin and Ouachita Orogenic Belt invokes extensional tectonics and rifting followed by a collision between the southern margin of North America and the continental plate Llanoria with south dipping subduction [*Briggs and Roeder*, 1975; *Graham et al.*, 1975; *Wickham et al.*, 1976; *Lillie et al.*, 1983; *Houseknecht*, 1986]. The following discussion is after *Houseknecht* [1986].

During the latest Precambrian or earliest Paleozoic (~750-500 Ma), rifting resulted in the opening of an ocean basin and the southern margin of North America evolved into an Atlantic-type margin that persisted through the middle Paleozoic (~400 Ma). During Devonian or early Mississippian time (~400-350 Ma), collision between the North American and Llanorian plates resulted in the closing of the ocean basin and southward subduction of the North American continent beneath Llanoria. By early Atokan time (~320 Ma), the remnant ocean basin had been consumed by subduction and the northward advancing subduction complex was being obducted onto the rifted continental margin of North America. By late Atokan time (~310 Ma), foreland-style thrusting became predominant as the subduction complex pushed northward. Resultant uplift along the frontal thrust belt of the Ouachitas completed the formation of a peripheral foreland basin.

Since formation of the Arkoma as a foreland basin in late Atokan time (~310 Ma), erosion has been the primary tectonic activity; the total amount of erosion can be estimated from the thermal maturity of strata presently exposed at the surface. Such estimates are fraught with uncertainties and involve assumptions concerning past geothermal gradients and surface temperatures. Nevertheless, approximate estimates of total erosion can be made by relating thermal maturity in the form of vitrinite reflectance to maximum paleotemperature [e.g., *Barker and Pawlewicz,* 1994] and by simply using the present-day geothermal gradient and surface temperature.

The vitrinite reflectance of near-surface coal beds in the Arkoma Basin ranges from 0.6 to 2.0% [*Houseknecht and Spötl,* 1993; *Spötl et al.*, 1993]. The vitrinite reflectance of near-surface shale of the Late Devonian Chattanooga Formation in the southwestern part of Ozarks is 0.5-0.8% [*Carr,* 1987]. Surface shale in the Frontal Ouachita Mountains has a vitrinite reflectance of 1.0-2.0% [*Underwood et al.*, 1988; *Houseknecht and Spötl,* 1993]. These maturity levels, given the above assumptions, suggest that total erosion has been in the range of 3 to 7 km in the Arkoma Basin, 2 to 4 km in the southwestern part of Ozarks, and 5 to 7 km in the Frontal Ouachita Mountains. Near profile A-A' (Figure 1), *Cardott et al.* [1986] found that the surface and near-surface vitrinite reflectance of the Boggy, McAlester, and Hartshorne Formations ranged 0.7-1.4%, implying a total erosion of 3.5-6 km. Although considerable uncertainties are inherent in all of these estimates, it is nonetheless clear that the Arkoma Basin and nearby Ouachita Mountains have undergone several kilometers of uplift and erosion.

4. Temperature Data

We estimated subsurface temperature along a north-south cross section (A-A') through the Arkoma Basin and the Oklahoma Platform (Figures 1, 2, and 3) from 345 bottom hole temperatures (BHTs) from 199 oil and gas wells [*Lee et al.*, 1994; *Lee and Deming*, 1994]. The temperature data were collected from well log headers on file at the log library of the Oklahoma Geological Survey in Norman, Oklahoma. All BHTs used in this study were measured within 10 km of cross section A-A'. The procedures employed to analyze BHT data have been discussed previously by *Deming* [1989], *Deming et al.* [1990], *Deming* [1994b], and *Lee et al.* [1994]. Log headers were first scrutinized for inconsistent data reflecting nonphysical conditions. From 345 BHTs, we obtained 56 sets of multiple temperatures, measured at the same depths in the same wells but at different times. These 56 sets of BHTs were suitable for a Horner plot correction (Figure 5). The Horner correction is

$$BHT = T_\infty + A \log_e\left(\frac{t_c + t}{t}\right) \qquad (1)$$

where T_∞ is the estimated equilibrium or true formation temperature, *BHT* is the bottom hole temperature recorded on the well log header, t_c is the duration of the drilling disturbance or circulation time, and t is the shut-in time, or time elapsed between the end of circulation and the time at which the *BHT* was measured [*Bullard*, 1947; *Lachenbruch and Brewer*, 1959, p. 78]. Because circulation time is not given on U.S. log headers, we used a standard circulation time (t_c) of 5 hours as has been done in similar studies [e.g., *Chapman et al.*, 1984; *Deming and Chapman*, 1988a, b]. Studies of the temperature buildup in a well show that circulation time is a relatively insensitive parameter when calculating equilibrium temperature from Horner plots [*Luheshi*, 1983]. The remaining 221 BHTs not correctable by means of a Horner plot were corrected by an average depth-time correction scheme [*Deming*, 1989] in which the average slope of the Horner line (A) is found as a function of depth from the subset of data amenable to Horner plot corrections,

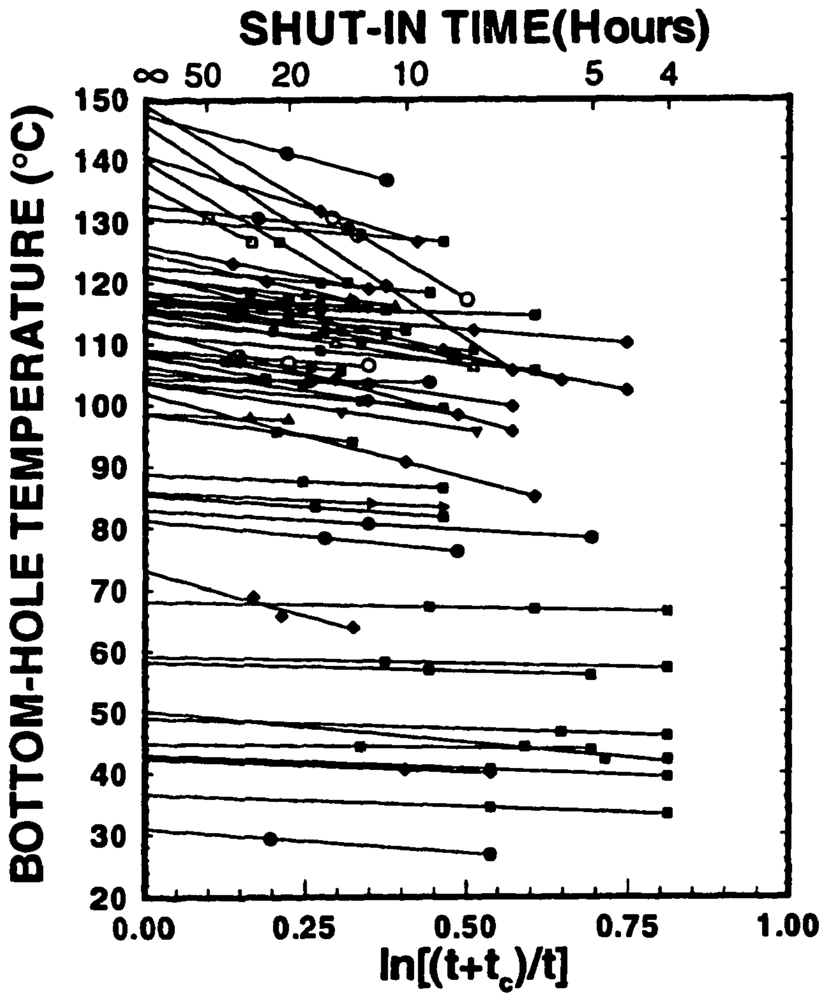


Figure 5. Horner plots for BHT data from 56 wells on or near cross section A-A'. Parameter t is shut-in time. A standard circulation time t_c of 5 hours was used for all Horner plots.

$$A = az + bz^2 \tag{2}$$

where A is the Horner slope (equation (1)) and z is depth. From a least squares fit to 56 A-z pairs we found a = -9.03, and b = 1.095, for A in degrees Celsius and z in kilometers.

We estimated the mean annual ground surface temperature along cross section A-A' to be 20°C; the intercept of a least squares linear regression on 277 corrected BHTs (Figure 6). The mean annual average surface air temperature in the vicinity of cross section A-A' was 16°C [*NOAA*, 1990]. As mean annual ground temperatures are usually a few degrees Celsius warmer than mean annual surface air temperatures [e.g., *Blackwell et al.*, 1980; *Powell et al.*, 1988], the meteorological data indicate that our estimate of the mean annual ground temperature is likely in error by no more than 1-2°C.

The average geothermal gradient within the Arkoma Basin and the Oklahoma Platform determined by a least squares regression on 277 corrected BHTs is 23°C/km (Figure 6). At a depth of ~4 km, the approximate spread in the data is 25°C. This is appreciably greater than the probable average error level in the corrected BHTs (see discussion in *Deming* [1994b]). It follows that much of the divergence of the corrected BHTs from the least squares line probably represents lateral variation in temperature throughout the Arkoma Basin and the Oklahoma Platform, not random error.

The fundamental problem in estimation of the subsurface temperature field is to partition scatter in the data (Figure 6) between random error in the corrected BHTs and lateral variation in the thermal state; this is the classic trade-off between

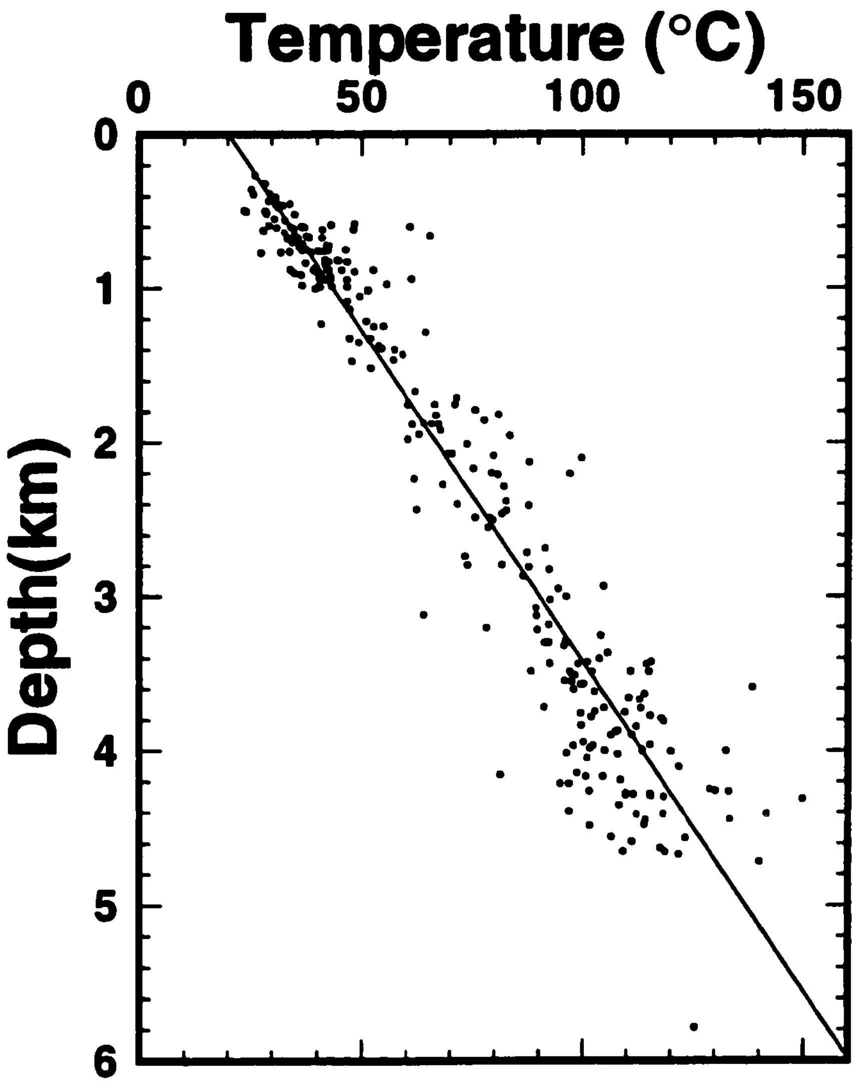


Figure 6. Corrected bottom hole temperatures from the Arkoma Basin and the Oklahoma Platform. A least squares fit yields an average geothermal gradient of 23°C/km and an intercept of 20°C at the surface.

resolution and variance, well-known in geophysical inversion theory [e.g., *Menke*, 1984]. It is impossible to determine precisely the optimum trade-off, because the exact nature and magnitude of error in the data are unknown. Nevertheless, significant improvements in resolution can be obtained beyond those offered by defining a single geothermal gradient for the entire Arkoma Basin and the Oklahoma Basin (Figure 6).

We estimated subsurface temperature in the Arkoma Basin and the Oklahoma Platform (Figure 7) using the methodology of *Deming et al.* [1990]. Surface temperature is assumed to be known, and the average geothermal gradient is calculated at different locations from a weighted least squares regression. The weighting procedure is chosen such that the closest N temperatures receive substantial weighting, while more distant data receive lower weightings [see *Deming et al.*, 1990]. The trade-off between resolution and variance is determined by two factors: an a priori estimate of the number of data N necessary to reduce error by averaging, and the spatial density of the data.

Figure 7 shows estimated subsurface temperatures along cross section A-A' for N=1, N=10, and $N=\infty$. For $N=\infty$ (Figure 7a), the estimated temperatures are the same as those determined by the average-gradient model (Figure 6), and there is no lateral variation across the basin. Model variance is low, but so is resolution. For N=1 (Figure 7b), the estimated temperature field shows high-frequency variations in temperature across the basin that may be nonphysical. In this case, resolution is high, but so is variance. For N=10 (Figure 7c), estimated temperatures and geothermal gradients across the basin show a large lateral variation, but the isotherms are still

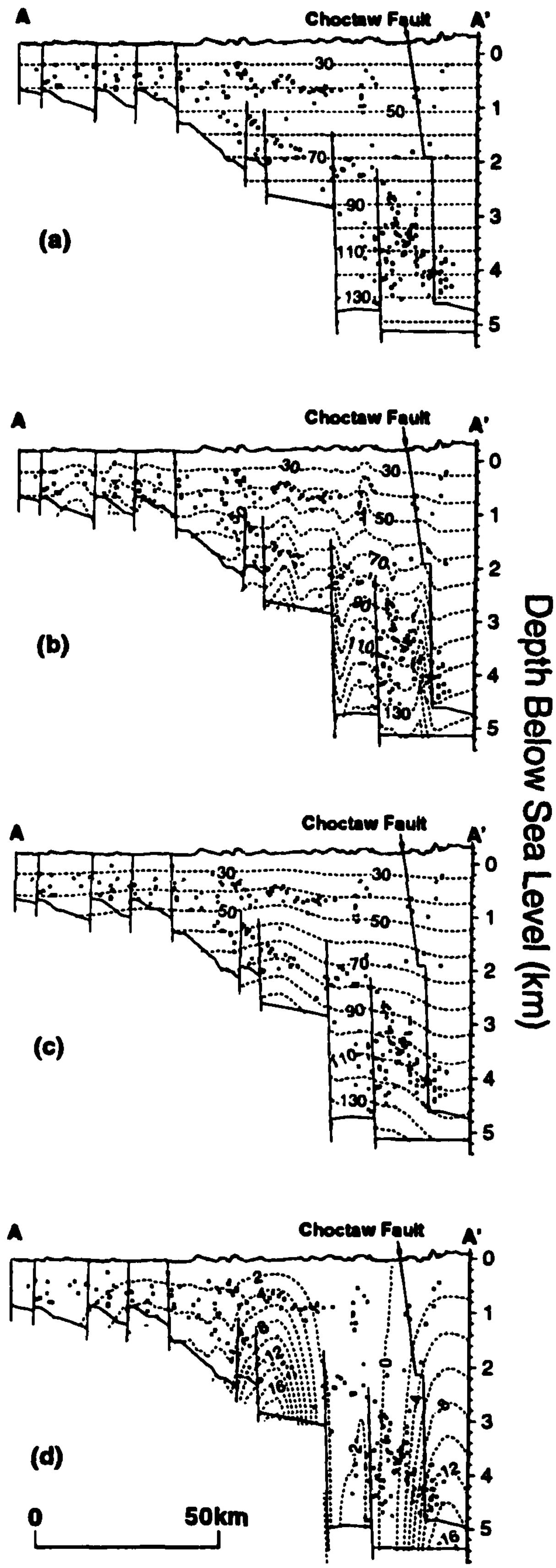

A
A'
Choctaw Fault
(a)
(b)
(c)
(d)
0
50km
Depth Below Sea Level (km)

Figure 7. Estimated temperature (degrees Celsius) along cross section A-A' through the Arkoma Basin and the Oklahoma Platform. The Choctaw Fault is the boundary between the Arkoma Basin and the Ouachita Mountains. Location of BHT measurements (crosses) are projected onto cross section A-A'. Solid lines along the bottom of each cross section show top of Precambrian granite and rhyolite, as shown in Figure 3. Dashed lines show estimated temperature in degrees Celsius. Three plots illustrate trade-off between resolution and variance in estimation of temperature. (a) All 277 corrected BHTs are given equal weighting; resolution is poor, but variance is low. (b) The closest corrected BHT is weighted heavily; resolution is high, but variance is also relatively high. (c) The closest 10 corrected BHTs are weighted heavily; resolution is better than for Figure 7a, but variance is lower than for Figure 7b. (d) Temperature difference between Figures 8c and 8a. Central part of the Arkoma Basin is up to 17°C warmer than average; southern end of the Arkoma Basin is up to 16°C colder then average.

smooth, indicating that an acceptable trade-off between resolution and variance has been achieved.

Estimated temperature (Figure 7c) along cross section A-A' varies from ~20°C at the ground surface to >140°C at the Precambrian boundary in the southern part of basin. The average geothermal gradient is 20°C/km in the southern part of the basin, exceeds 30°C/km in the middle part of the basin, and is 24°C/km in the northern part of the basin. At depths of 2-3 km, the lateral variation in basin temperature is ~20-25°C (Figure 7d), more than 4 times the probable level of random error in the corrected BHTs. We thus believe that the observed variation in thermal gradient through the basin is not an artifact of noisy data but rather reflects actual variation in the subsurface thermal state.

5. Temperature Error Analysis

Comparisons of uncorrected BHTs with equilibrium temperatures show that raw BHTs are systematically lower than true rock temperatures by about 10-15°C [*Kehle*, 1972; *Ben Dhia*, 1988; *Deming*, 1989]. The error level in corrected BHTs is more difficult to estimate. Certainly, a significant component of random error remains after correction due to uncertainty in correction parameters such as circulation time. Our correction procedures are intended to reduce the systematic error associated with the circulation of drilling mud from 10-15°C to perhaps 2-3°C. The random error that remains in corrected BHTs is then reduced through averaging. Although a definitive study has not been done, the average random error in corrected BHTs is probably of the order of ±5°C, depending upon local conditions and the type of correction procedure used [*Deming*, 1989, 1994b]. Averaging, as we have done, 10 corrected BHTs for each estimated thermal gradient, probably reduces the error level in estimated temperature (Figure 7c) below ±5°C.

Given an estimate of the subsurface temperature field (Figure 7c), error in estimated thermal gradients depends upon depth. Meteorological data (see above) indicate that estimated ground surface temperature is probably inaccurate by no more than ±1-2°C. If it is assumed that the error in surface temperature is uncorrelated with the error in estimated subsurface temperature (±5°C), then, using standard propagation of error techniques for uncorrelated random variables [*Barry*, 1978, p. 73], the total error in the temperature change from surface to total depth is ±5°C. In the midsection of profile A-A', where the average depth of temperature measurement is about 3 km and the average thermal gradient is 30°C/km, the estimated uncertainty in the geothermal gradient is ±6%. In the deeper sections of the basin, where the maximum depth is about 4 km and the average geothermal gradient is 20°C/km, the estimated error level is ±6%. On the Oklahoma Platform to the north, where the sedimentary cover thins, an average depth of 1.5 km and average geothermal gradient of 24°C/km imply an estimated error of ±14%.

6. Thermal Conductivity

6.1 Methodology

To estimate heat flow, we made 843 thermal conductivity measurements (Table 1) on drill cuttings from 11 wells (locations shown in Figure 2) using the cell technique of *Sass et al.* [1971]. Sampled depth intervals are shown in Figure 8b. Measurements

Table 1. Thermal Conductivity and Porosity Data, Southeastern Oklahoma

Geologic Unit	Lithology	N[a]	Matrix Conductivity[b] W/m°K	Standard Deviation[c] W/m°K	Standard Error[d] W/m°K	Porosity[e] %
Savanna Fm.	shale, sandstone	32	2.05	1.04	0.18	5.8
McAlester Fm.	shale, sandstone	63	1.84	0.68	0.09	8.1
Hartshorne Fm.	Sandstone	26	2.28	0.58	0.11	7.1
Atoka Fm.	shale, sandstone	392	1.71	0.44	0.02	8.1
Morrowan Fm.	limestone, shale	130	1.49	0.69	0.06	2.3
Mississippian-Devonian Systems	shale, limestone	52	1.73	0.55	0.08	2.4
Devonian-Silurian-Ordovician Systems	limestone, shale	23	1.97	0.56	0.12	0.0
Ordovician System	limestone, shale, sandstone	55	2.40	1.39	0.19	0.6
Ordovician-Cambrian Systems	limestone, dolomite, sandstone	70	4.11	0.95	0.11	2.2

[a] Number of measurements.
[b] Harmonic mean perpendicular to bedding at 22°C.
[c] From the arithmetic mean.
[d] Of the arithmetic mean.
[e] Estimated from density logs and matrix density measurements.

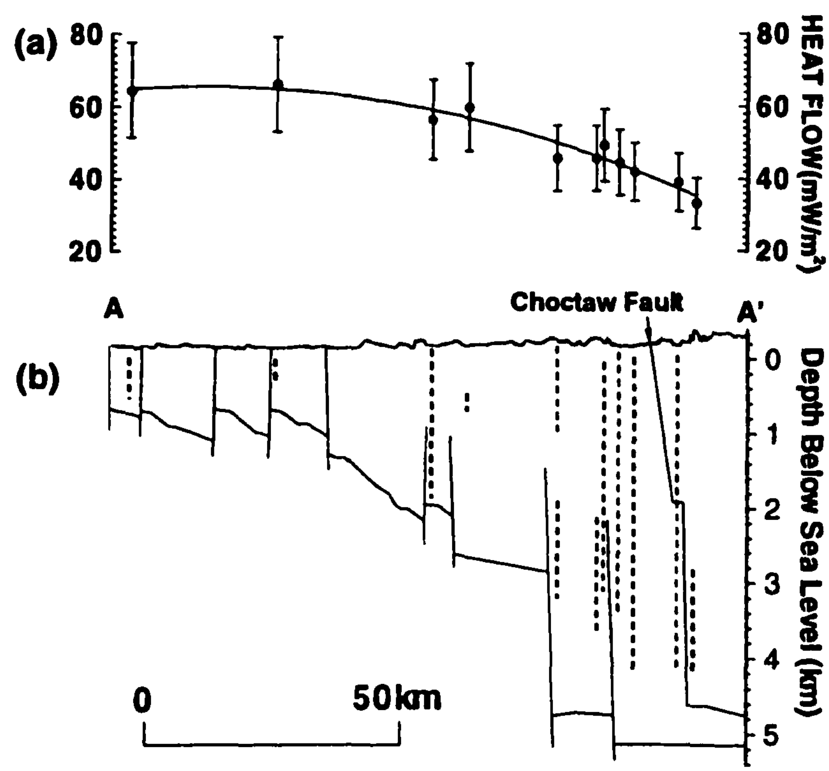


Figure 8. (a) Estimated heat flow. Solid line is polynomial fit to discrete points. (b) Vertical dashed lines show sampling intervals for thermal conductivity measurements. The Choctaw Fault is the boundary between the Arkoma Basin and the Ouachita Mountains.

were made with a divided-bar apparatus [*Birch*, 1950; *Beck*, 1957; *Roy et al.*, 1981; *Sass et al.*, 1984]. The cell technique of *Sass et al.* [1971] allows the determination of the thermal conductivity of a randomly oriented aggregate of rock matrix at room temperature (λ_{ag}). To estimate λ_{pr}, the *in situ* thermal conductivity of a porous rock perpendicular to bedding, corrections must be made for the effects of anisotropy, temperature, and porosity.

We made an anisotropy correction using the method of *Deming* [1994a], calibrated by measurements on Pennsylvanian age sedimentary rocks in northcentral Oklahoma reported by *Deming and Borel* [1995]. Matrix conductivity perpendicular to bedding (λ_z) was calculated as

$$\lambda_z = \exp\left[\frac{\log_e(\lambda_{ag}) - 0.6267}{0.5480}\right] \tag{3}$$

for $1.87 < \lambda_{ag} < 4.0$ W/m-K. For $\lambda_{ag} > 4.0$ W/m-K no correction is needed, for $\lambda_{ag} < 1.87$ W/m-K, λ_z was taken as 1.0 W/m-K.

In situ temperature was estimated from corrected BHTs (Figure 7c) and used to correct matrix thermal conductivities measured in the laboratory at 22°C using the method of *Sekiguchi* [1984]. The correction applied was

$$\lambda_T = \lambda_m + \left(\frac{T_0 T_m}{T_m - T_0}\right)(\lambda_z - \lambda_m)\left(\frac{1}{T} - \frac{1}{T_m}\right) \tag{4}$$

where T is the estimated *in situ* temperature in Kelvins, λ_T is the estimated matrix conductivity perpendicular to bedding at *in situ* temperature T, λ_m and T_m are the thermal conductivity and absolute temperature (Kelvins) at what *Sekiguchi* [1984, p.

75] refers to as "the assumed point", and λ_z is the matrix thermal conductivity perpendicular to bedding at room temperature T_o. We used values for λ_m and T_m suggested by Sekiguchi of 1.8418 W/m-K and 1473 K, respectively. For sedimentary rocks over a range of temperatures corresponding to *in situ* temperatures in the Arkoma Basin and the Oklahoma Platform (~20-140°C), we found that the *Sekiguchi* [1984] method matched available data [*Birch and Clark*, 1940; *Sugawara and Yoshizawa*, 1961; *Kawada*, 1964, 1966; *Anand et al.*, 1973; *Kappelmeyer and Haenel*, 1974; *Sibbitt et al.*, 1979; *Roy et al.*, 1981; *Cermak and Rybach*, 1982; *Mongelli et al.*, 1982; *Robertson*, 1988; *Seipold*, 1990] better than alternative corrections [*Zoth and Haenel*, 1988; *Sass et al.*, 1992], especially for rocks with conductivities at 25°C lower than 2.0 W/m-K.

Porosity corrections were made using a geometric mean model. *In situ* conductivity (λ_{pr}) was estimated as

$$\lambda_{pr} = \lambda_T^{1-\phi} \lambda_w^{\phi} \tag{5}$$

where λ_T is the estimated matrix conductivity perpendicular to bedding at *in situ* temperature T, λ_w is the estimated conductivity of pore fluid (water) at *in situ* temperature T, and ϕ is the average formation porosity. The conductivity of the saturating pore fluid (λ_W) was assumed to be the same as pure water and was calculated as $\lambda_w(T) = 0.5648 + 1.878\times10^{-3}T - 7.231\times10^{-6}T^2$ for $0 \le T \le 137$°C, and $\lambda_w(T) = 0.6020 + 1.309\times10^{-3}T - 5.140\times10^{-6}T^2$ for $137 \le T \le 300$°C [*Touloukian et al.*, 1970]. Porosity was estimated from density logs calibrated by density measurements on drill cuttings used for thermal conductivity measurements (Table 1). The estimated porosity of most formations in the Arkoma Basin and the Oklahoma Platform was found to be relatively low, likely reflecting former great burial depths and subsequent erosion. In one case, for Lower Devonian-Silurian-Upper Ordovician age units, the calculated porosity was zero. We believe that the actual porosity of these units is probably low, perhaps 1-2%, and the zero number is a reflection of imprecision in porosity estimates related to sampling error. Geologists with working knowledge of Arkoma Basin stratigraphy have confirmed to us that the actual porosity of these units is very low (N. Suneson (Oklahoma Geological Survey), personal communication, 1995; J. M. Forgotson Jr. (School of Geology and Geophysics, The University of Oklahoma), personal communication, 1995).

6.2 Results

Matrix conductivities perpendicular to bedding at 22°C (λ_z) ranged from 1.5 W/m-K for the Morrowan Series to 4.1 W/m-K for Ordovician-Cambrian age units (Table 1). Average *in situ* conductivities (λ_{pr}) (Table 2) at 11 sites where heat flow was estimated, ranged from 1.7 to 2.6 W/m-K. The relatively low thermal conductivites reflected the dominance of shale in the lithologic section.

6.3 Error Analysis

Error in estimated *in situ* thermal conductivities may arise from (1) systematic errors in measuring devices, (2) sampling errors, and (3) errors in corrections for anisotropy, temperature, and porosity.

Table 2. Heat Flow Summary

Well Name	N[a]	Latitude °N	Longitude °W	Section, Township and Range	In-situ Conductivity[b] W/m-K	Geothermal Gradient °C/km	Heat Flow[c] MW/m²
Foltz #1	1	35.93°	95.72°	28-15N-17E	2.59	24.9	65
Rogers #1	2	35.47°	95.33°	31-12N-19E	2.39	27.6	66
Griffin #1	3	35.22°	95.37°	26-9N-18E	2.01	28.0	56
Max Brandt #1	4	35.16°	95.53°	17-8N-17E	2.00	29.9	60
Jack Murdaugh #1	5	35.05°	95.39°	27-7N-18E	1.93	23.6	46
Manschrich #1	6	34.97°	95.51°	28-6N-17E	1.85	24.7	46
Brown #1	7	34.95°	95.51°	33-6N-17E	2.00	24.6	49
Davis #1	8	34.94°	95.92°	2-5N-18E	1.90	23.4	45
Bud Hampton #2	9	34.91°	95.44°	18-5N-18E	1.83	22.9	42
Spears #1	10	34.82°	95.43°	16-4N-18E	1.91	20.4	39
Garrett unit 1-A	11	34.76°	95.59°	2-3N-16E	1.72	19.4	33

[a]Well numbers shown on Figure 2.

[b]Harmonic mean of measurements after corrections for anisotropy, temperature, and porosity.

Estimate error is ±20%.

Measurement error. The University of Oklahoma divided bar is calibrated with disks of fused silica using values given by *Ratcliffe* [1959]; individual sample holders (cells) are calibrated using the known thermal conductivity of water. We estimate that the probable inaccuracy due to systematic errors in our divided bar and cell technique is no greater than ±5%. In addition to systematic errors inherent in the measurement device itself (divided bar), the cell technique relies upon a mixing model which can introduce additional errors [*Sass et al.*, 1971]. However, we have found that the conductivity of an isotropic crushed aggregate of fused silica estimated with the cell technique is within 1-2% of the value used to calibrate the divided bar. We thus believe that the use of a geometric-mean mixing model by itself introduces a negligible error, at least when working with isotropic materials.

Sampling error. Sampling errors may be due to inadequate vertical or lateral sampling. The standard error of the (arithmetic) mean for nine different geologic units (Table 1) ranged from 1% to 9%, with a mean standard error of 5%. We were not able to obtain complete vertical profiles at all of the sites where heat flow estimates were made; about half (48%) of the thermal conductivities used in heat flow estimates (see below) are average values determined from measurements on samples from nearby wells (average distance ~10 km). Sampling errors related to lateral variability are difficult to estimate without measurements. Conversely, if we had sufficient measurements, such sampling errors would be negligible.

Correction errors. Errors due to corrections for anisotropy are difficult to estimate precisely. The correction procedure we used is based on an inverse correlation between thermal conductivity perpendicular to bedding and anisotropy

[*Deming*, 1994a]. Available anisotropy data from the literature show considerable scatter about the average correction proposed by *Deming* [1994a]; however, in any suite of data, some errors are positive, while others are negative. The average correction we applied in this study, calibrated by measurements on Pennsylvanian age sedimentary rocks in north central Oklahoma [*Deming and Borel*, 1995], differs little from the generalized correction proposed by *Deming* [1994a] from literature data on anisotropy, or a correction developed from measurements on sedimentary rocks from the North Slope of Alaska by *Deming et al.* [1992]. We estimate that the overall error due to uncertainties introduced by the anisotropy correction is ±10% but lack a definitive basis for quantifying uncertainties related to anisotropic effects.

It is similarly difficult to precisely estimate the uncertainties inherent in temperature corrections. There is a moderately extensive literature [*Birch and Clark*, 1940; *Sugawara and Yoshizawa*, 1961; *Kawada*, 1964, 1966; *Anand et al.*, 1973; *Kappelmeyer and Haenel*, 1974; *Sibbitt et al.*, 1979; *Roy et al.*, 1981; *Cermak and Rybach*, 1982; *Mongelli et al.*, 1982; *Robertson*, 1988; *Seipold*, 1990] on the change of thermal conductivity with temperature, but most data are for crystalline rocks. A comparison of the *Sekiguchi* [1984] correction with what data are available for sedimentary rocks led us to estimate the average error level as ±10%.

We do not believe the error due to uncertainty in porosity is significant in comparison to other errors. For example, if the estimated porosity is as high as 8% (Table 1), an error of ±25% leads to a porosity range of 6-10%. For a matrix conductivity of 2.0 W/m-K and fluid conductivity of 0.6 W/m-K, this leads to an error in conductivity of ±2-3%.

Propagation of errors. Propagating the above uncorrelated error sources using standard techniques [*Barry*, 1978, p. 73] leads to a total estimated error of ±18%.

7. Heat Production

Few heat production measurements have been made on basement rocks in Oklahoma, largely due to the fact that nearly the entire state is covered by substantial thicknesses of sedimentary rock [*Roy et al.*, 1968a, b; *Borel*, 1995]. The core library of the Oklahoma Geological Survey contained no core samples from basement rocks in the Arkoma Basin and northeast Oklahoma Platform. Therefore we estimated heat production (Table 3) in these areas from gamma ray logs using an empirical relationship introduced by *Bücker and Rybach* [1996]. The relationship is

$$A = 0.0158(\gamma - 0.8) \tag{6}$$

where A is heat production in $\mu W/m^3$ and γ is the gamma ray log reading in API units. *Bücker and Rybach* [1996] claimed that (6) yields an estimate of heat production that is valid for $0.03 \leq A \leq 7.0$ $\mu W/m^3$ and is accurate to within ±10%. The validity of the relationship depends upon "standard" Th/U and K/U ratios, and though the *Bücker and Rybach* [1996] relationship has been calibrated for a variety of rock types in numerous research wells, it is possible that it may not be valid for basement rocks in Oklahoma. However, owing to the general inaccessibility of basement rocks in the study area, the gamma ray method was the only viable alternative for estimating heat production. We do not know to what extent, if any, the heat production of basement rocks penetrated by logged wells may have been altered by paleoweathering.

Table 3. Heat Production Estimates, Southeastern Oklahoma

Section, Township and Range	Latitude °N	Longitude °W	Heat Production[a] $\mu W/m^3$	Depth to Basement m	Sampling Interval Below the Top of Basement m	Geological Unit[b]
21-8N-26E	35.15°	94.55°	2.3	3801	102	WVG
36-8N-22N	35.12°	94.93°	2.1	2928	53	WVG
26-8N-22E	35.14°	94.94°	2.4	2737	67	WVG
29-9N-18E	35.22°	95.42°	2.8	2512	51	COGG
16-12N-22E	35.51°	94.98°	1.9	837	31	COGG
24-15N-19E	35.76°	95.24°	3.1	1036	9	SGG
18-17N-17E	35.95°	95.55°	2.0	550	212	WVG
6-18N-18E	36.07°	95.43°	3.3	652	22	WVG
25-19N-17E	36.10°	95.45°	3.5	619	24	SGG
9-19N-20E	36.11°	95.19°	3.5	335	10	SGG
6-19N-18E	36.15°	95.43°	2.4	646	9	SGG
36-21N-16E	36.26°	95.55°	2.5	685	251	SGG
33-22N-21E	36.34°	95.05°	2.6	255	12	SGG
24-20N-11E	36.19°	96.08°	2.6	1108	17	WVG
9-21N-9E	36.31°	96.33°	2.5	796	19	OM
1-21N-7E	36.33°	96.51°	3.3	958	7	COGG
10-24N-18E	36.58°	95.37°	2.6	618	11	COGG

[a]Estimated from gamma ray logs as explained in text;average of 17 estimates is 2.67 ± 0.12 $\mu W/m^3$.
[b]WVG, Washington Volcanic Group; COGG, Central Oklahoma Granite Group; OM, Osage Microgranite; SGG, Spavinaw Granite Group. Average rock age is 1286 ± 18 Ma [*Denison*, 1981].

The average heat production of basement rocks estimated from 17 wells in east central and southeast Oklahoma (Figure 1) was 2.7±0.1 $\mu W/m^3$ (Table 3). Average heat production in the Arkoma Basin was 2.3±0.2 $\mu W/m^3$; average heat production in the Oklahoma Platform and Ozarks was 2.8±0.1 $\mu W/m^3$. The tectonothermal age of basement rocks in southeastern Oklahoma is Middle Proterozoic (1286±18 Ma) [*Denison*, 1981]. Worldwide the average heat production rates of basement rocks of Early Proterozoic and Late Proterozoic age are 1.9±0.6 and 2.4±1.2 $\mu W/m^3$, respectively; where the range indicates ±1 standard deviation [*Vitorello and Pollack*, 1980]. The measurements reported in this study thus appear to be slightly higher than average for continental rocks of similar ages but not so much as to be anomalous.

8. Heat Flow

8.1 Results

We estimated heat flow by multiplying average thermal gradients (Figure 7c) by the estimated average *in situ* thermal conductivities at 11 sites along cross section A-A' (Figures 1, 2, and 8, and Table 2). The 11 sites chosen were those wells from which we were able to obtain substantial quantities of drill cuttings for thermal conductivity measurements.

Estimated heat flow increases systematically from south to north, ranging from a low of 33 mW/m^2 in the deep part of the Arkoma Basin in the south to a high of 66 mW/m^2 on the Oklahoma Platform to the north (Figure 8). The relatively low thermal gradients in the deep section of the Arkoma Basin are thus not due solely to simple changes in thermal conductivity. Rather, there is a significant change in heat flow along cross section A-A'. The mean of 11 heat flow values is 50 mW/m^2. However, as more heat flow determinations were made in the south section of the Arkoma Basin where heat flow is lower, the mean background heat flow along the north-south profile is closer to 55 mW/m^2. On the southern end of cross section A-A', near the Ouachita Mountain front, the average heat flow value is 44±9 mW/m^2. In the central part of the basin, average heat flow is 58±12 mW/m^2, increasing to 66±13 mW/m^2 in the north near the stable midcontinent platform.

8.2 Error Analysis

Using standard propagation of error techniques for uncorrelated random variables [*Barry*, 1978, p. 73], we estimated heat flow errors by propagating the estimated error in thermal conductivity (±18%) with estimated errors in thermal gradients (±6-14%). Estimated error in heat flow ranges from ±19% along the south and central portions of profile A-A' to ±23% on the Oklahoma Platform to the north. In view of the uncertainties involved in estimating the magnitude of errors, it is useful to round the estimated error level off to a uniform ±20%.

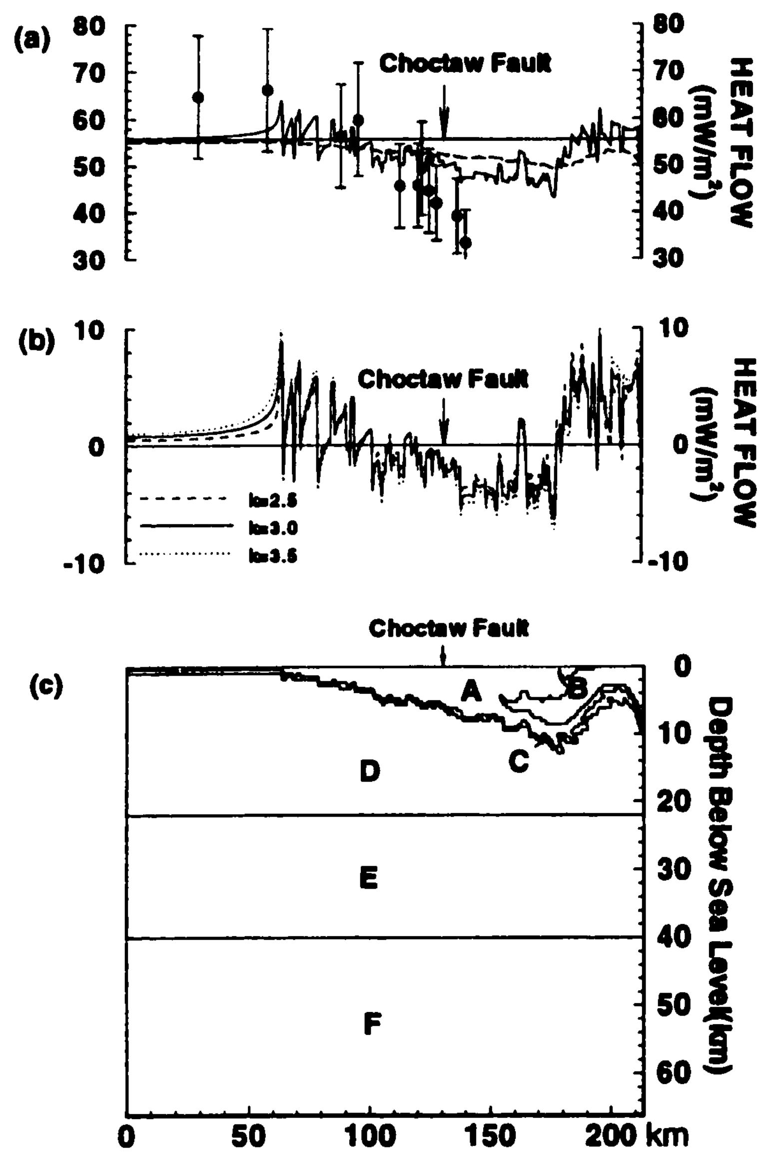
9. Hypotheses to Explain Heat Flow Variation

We considered and quantitatively evaluated three hypotheses to explain the observed variation in heat flow from north to south: heat refraction, lateral variations in crustal heat generation, and groundwater flow. It is difficult to explain the heat flow change by other mechanisms. The Ouachita Mountains and Arkoma Basin have been undergoing uplift and erosion for approximately the last 300 Myr (see earlier discussion and references). However, erosion rate decreases exponentially with increasing time [*Vitorello and Pollack*, 1980], and it is therefore unlikely that present-day heat flow has been significantly elevated by erosion. There is evidence that the eastern third of the Ouachitas experienced a Mesozoic heating event (about 100 Myr) related to rifting and intrusive activity in the Mississippi Embayment [*Houseknecht and Matthews*, 1985; *Arne*, 1992]. However, we know of no evidence for volcanic activity in the western Ouachitas and Arkoma Basin. Furthermore, both erosion and volcanic activity are not viable hypotheses to explain decreasing heat flow to the south, as both would tend to elevate heat flow in the Ouachita Front. Finally, it is possible that the observed trend of decreasing heat flow to the south could be due to some unknown, deep-seated cause below the base of the crust. However, we have no geologic evidence for the existence of such a cause. In contrast, the three hypotheses we evaluate in the following discussion are suggested by the known geologic structure and history.

9.1 Heat Refraction

Low heat flow in the deep section of the Arkoma Basin (Figures 8 and 9) could be due to refraction of heat from the relatively low thermal conductivity sedimentary rocks of the Arkoma Basin (~1.6 W/m-K as determined by measurements) into the more conductive crystalline rocks (estimated to be ~3.0 W/m-K) found in the Ouachita Mountains to the south and the basement of the Oklahoma Platform to the north.

To estimate the magnitude of heat refraction along cross section A-A', we used a two-dimensional finite difference model (Figure 9) to solve the equation which



(a)
HEAT FLOW (mW/m²)
Choctaw Fault
(b)
HEAT FLOW (mW/m²)
Choctaw Fault
k=2.5
k=3.0
k=3.5
(c)
Choctaw Fault
A
B
C
D
E
F
Depth Below Sea Level(km)
km

Figure 9. (a) Estimated heat flow (discrete points) and heat flow calculated from numerical model. Dashed line (no heat refraction) shows effect of basin depth. Heat flow estimated from the temperature difference from surface to top of basement decreases as sediment depth increases due to loss of heat flow from heat generation in sedimentary rocks (assumed to be 1 mW/m^3). Solid line shows combined effects of basin depth and heat refraction, assuming a basement conductivity (unit D) of 3.0 W/m-K at 25 °C. (b) Estimated heat flow change due solely to heat refraction for basement conductivities (unit D) of 2.5, 3.0, and 3.5 W/m-K. (c) Numerical model used to estimate magnitude of heat refraction. Units A, B, C, D, E, and F have thermal conductivities of 1.6, 3.2, 3.4, 3.0 (at 25°C), 2.6, and 3.5 W/m-K, respectively. Thermal conductivity is constant in each unit, with the exception of unit D for which it is calculated as a function of temperature as explained in text. Heat production in units A, B, C, D, E, and F is 1.0, 1.0, 1.0, 1.0, 0.45, and 0.02 $\mu W/m^3$, respectively. Thermal conductivity values for basement rocks follow those suggested by *Furlong and Chapman* [1987] and *Chapman and Furlong* [1992].

describes steady state heat conduction in a heterogeneous domain with heat production,

$$\frac{\partial}{\partial x}[\lambda_x \frac{\partial T}{\partial x}] + \frac{\partial}{\partial z}[\lambda_z \frac{\partial T}{\partial z}] + A^* = 0 \tag{7}$$

where T is temperature, x and z are the horizontal and vertical coordinates, respectively, λ_x and λ_z are the horizontal and vertical thermal conductivities, respectively, and A^* is radioactive heat generation in W/m^3. Boundary conditions were set as follows. Lateral boundaries were insulating, temperature (20°C) was fixed at the top of the model, and heat flow was held constant (25 mW/m^2) at 66 km depth.

Geologic cross section A-A' did not extend far enough into the Ouachita Mountains to allow us to constrain the geologic structure and thus heat refraction. Therefore we used information from geologic cross section B-B' (Figure 1) about 67 km to the east of geologic cross section A-A' as a template for constructing the mathematical model (Figure 9c). The geologic cross section (B-B') upon which our model is based can be found in the work by *Roberts* [1994]. Although this introduced some uncertainty due to lateral variations in structure, the uncertainty was tolerable as the entire modeling process was an imprecise exercise that sought to constrain the magnitude of heat refraction in an approximate sense only.

To isolate the magnitude of the refractive effect, we specified no lateral variation in heat production for a series of model simulations. We used a simple, four-unit model (A, B, C, D) for the thermal conductivity structure of the upper crust (Figure 9c); heat production was specified as a constant value of 1 μW/m^3 for each unit.

The thermal conductivity of each of the sedimentary layers was constrained by measurement and so held constant. Unit A (1.6 W/m-K) corresponded to clastic sedimentary rocks in the Arkoma Basin and the Oklahoma Platform younger than Lower Ordovician-Upper Cambrian, unit B (3.2 W/m-K) represented Upper Ordovician-Upper Cambrian sedimentary rocks in the Ouachita Mountains, and unit C (3.4 W/m-K) represented Lower Ordovician-Upper Cambrian sedimentary rocks. The thermal conductivities of unit A and C were constrained by measurements. We had no direct thermal conductivity measurements for rocks in unit B but used values determined for rocks of similar age and composition in the Arkoma Basin and the Oklahoma Platform to the north. We also considered the effect of introducing a more detailed thermal conductivity structure for individual geologic formations but found the results to not be significantly different from those obtained using the simpler four-unit model.

The thermal conductivity of crystalline rocks in the upper crust (unit D, Figure 9c) was estimated to be 3.0 W/m-K at 25°C [*Chapman and Furlong*, 1992] but was specified to decrease with increasing temperature according to the relationship given by *Sekiguchi* [1984]. We found that when the basement conductivity was specified as 3.0 W/m-K (at 25°C), the total magnitude of heat refraction was 8 mW/m^2 (±4 mW/m^2). When the basement conductivity (at 25°C) was varied from 2.5 to 3.5 W/m-K, reflecting a realistic range of uncertainty, the magnitude of heat refraction was found to vary from 7 to 9 mW/m^2 (Figure 9b). The relative insensitivity of the magnitude of heat refraction to the conductivity of basement rocks was due in part to the presence of a core of relatively high-conductivity (3.2 W/m-K) Upper Ordovician-

Upper Cambrian sedimentary rocks in the Ouachita Mountains that strongly influenced the refraction of heat out of the Arkoma Basin.

The total magnitude of heat flow change from north to south caused by heat refraction appears to be of the order of 5-10 mW/m^2. The range of uncertainty is due to uncertainties in thermal conductivity, as well as a realistic appreciation of the limitations of a simplified, two-dimensional model.

9.2 Heat Production

Some part of the observed north-south variation in heat flow is an artifact of the methodology we used to estimate heat flow. The average thermal gradients used in heat flow estimations at 11 sites were calculated from estimated surface temperature and the estimated temperature (Figure 7c) near the bottom of the Arkoma Basin and the Oklahoma Platform. The average heat flow calculated in this manner was lower than the surface heat flow owing to the loss of heat flow contributed by radioactive heat generation in sedimentary rocks. For an average heat production of 1 $\mu W/m^3$ [*Kappelmeyer and Haenel*, 1974; *Rybach*, 1988] and a basin depth of 5 km near the Choctaw Fault, the apparent decrease in heat flow was 2.5 mW/m^2 (Figure 9a).

Variations in the concentration of heat-producing radioactive isotopes in Precambrian basement rocks (granite and rhyolite) are likely to have also contributed to the observed decrease in heat flow from north to south (Figure 8a). Our estimates of heat production (Figure 1 and Table 3) were not as numerous or as detailed in their spatial coverage as might have been desirable, but what data were available indicated that mean heat production in basement rocks in the Arkoma Basin (2.3 $\pm$0.2 $\mu W/m^3$)

was lower than average heat production in the Oklahoma Platform and Ozarks to the north (2.8 ±0.1 μW/m^3).

The Arkoma Basin originally formed as an extensional basin in late Precambrian to middle Paleozoic time (~750-400 Ma) as evidenced by normal faults in basement rocks immediately underlying the Arkoma Basin (Figure 3). Geophysical evidence indicates that the crust underneath the Arkoma Basin has been thinned, compared to continental crust on the platform to the north, and thus the total heat flow due to radioactive heat generation should be lower even if the concentration of radioactive heat-generating elements is the same. Seismic refraction experiments and inversion of gravity data indicate that the upper crust underneath the Ouachita Front has been thinned by up to 50% [*Lillie et al.*, 1983]. Simple isostatic calculations lead to a somewhat smaller estimate of the total crustal thinning. Assuming an original crustal thickness of 40 km [*Lillie et al.*, 1983], a mean crustal density of 2900 kg/m^3, mean mantle density of 3300 kg/m^3, mean sediment density of 2600 kg/m^3 (constrained by laboratory measurements and porosity estimates), and depth of compensation at 100 km, a simple Airy isostatic balance [e.g., see *Turcotte*, 1980] suggests that the crust underneath the deep part of the Arkoma Basin has been thinned by about 17%. Considering that heat loss due to crustal thinning is partially mitigated by heat production in sedimentary rocks (estimated to be 1 μW/m^3) in the Arkoma Basin and assuming that radioactive elements in the upper crust (upper 10 km) are distributed according to an exponential model [*Lachenbruch*, 1970], the total change in heat flow from south to north due to change in crustal heat production is approximately 5-15 mW/m^2. The range of uncertainty reflects uncertainties in the amount of crustal thinning that has taken place, as well as uncertainties in factors such as the exponential decay constant in the *Lachenbruch* [1970] model of crustal heat production.

9.3 Groundwater Flow

Adding up the contributions from refractive effects and changes in heat production, about 10 to 25 mW/m^2 of the observed increase in heat flow (20-30 mW/m^2) from south to north can readily be explained by a combination of heat refraction and lateral changes in heat production, as constrained by both measurements and knowledge of the regional geologic history and structure. It is thus possible to entirely explain the observed change in heat flow without invoking groundwater movement. However, due to uncertainties, it is also possible that as much as 20 mW/m^2 of the total heat flow change from north to south could be due to advective heat transport by regional groundwater flow (or other unknown factors). Regional groundwater flow has been invoked as the probable cause of similar thermal anomalies found in other sedimentary basins. These include the Uinta Basin in Utah [*Chapman et al.*, 1984]; the Western Canadian Sedimentary Basin [*Majorowicz and Jessop*, 1981; *Majorowicz*, 1989], and the North Slope Basin in Alaska [*Deming et al.*, 1992].

Darcy's law implies there must be fluid flow between any two points with a finite permeability and nonzero potential energy gradient. The topographic gradient (~0.003) from the Ouachita Mountains northward to the Arkoma Basin and the

Oklahoma Platform (Figure 3), although not large, is also nonzero and thus must induce some fluid flow from south to north with concomitant heat transport.

Domenico and Palciauskas [1973] considered steady state groundwater flow in an idealized, two-dimensional basin with homogeneous and isotropic physical properties and showed that a relative gauge of the magnitude of convective to conductive heat transport (Peclet number, Pe) under these circumstances is given by the dimensionless group

$$P_e = \frac{\Delta h k \Delta z \rho^2 g C}{2 \mu L \lambda_{pr}} \tag{8}$$

where Δh is the total head drop across a basin of length L and depth Δz, k is permeability, ρ is fluid density, C is fluid specific heat, g is the acceleration due to gravity, μ is fluid dynamic viscosity, and λ_{pr} is the thermal conductivity of a porous rock (matrix and fluid). If we take the head drop to be equal to the elevation drop across a distance L from the Ouachita Mountains through the Arkoma Basin, $\Delta h/L \sim$ 330/115,000. For Δz ~3000 m, ρ = 1000 kg/m^3, g = 9.8 m/s^2, C = 4200 J/kgK, λ_{pr} = 1.6 W/m-K, μ = 5 x 10^{-4} kg/ms (pure water at 50°C), an average permeability of k = 9.0 x 10^{-16} m^2 will result in P_e = 0.2, implying an advective component of ~10 mW/m^2, sufficient to reduce the conductive heat flow in the recharge area by 10 mW/m^2 and increase the conductive heat flow in the discharge area by 10 mW/m^2. Thus a maximum estimate of the regional scale permeability is 10^{-15} m^2.

Permeability data from rocks in the Arkoma Basin are scarce, especially on the regional scale relevant to basin-wide heat transport. *Jorgensen et al.* [1993] estimated the permeability of Jurassic through Upper Mississippian age rocks in and near the Arkoma Basin to range from 10^{-14} m^2 to 10^{-15} m^2 [*Jorgensen et al.*, 1993]. Verbal reports of low-permeability rocks in the Arkoma Basin are common (D. Blackwell, personal communication, 1996); however, we have been unable to find published data to back up these assertions. However, porosity is correlated with the logarithm of permeability [*Nelson*, 1994; *Neuzil*, 1994], and rocks in the Arkoma Basin do tend to have low porosities (Table 1).

10. Discussion

Our results are generally not supportive of theories which seek to explain the genesis of Mississippi Valley-type ore deposits in the North American midcontinent by topographically driven fluid flow from the Ouachita Mountains and Arkoma Basin in Late Pennsylvanian-Early Permian time (~290 Ma). These theories generally invoke relatively high permeabilities and heat flows. For example, *Deming and Nunn* [1991] found that model simulations of fluid temperatures in discharge areas were generally lower than that indicated from studies of fluid inclusions, even when the background heat flow was as high as 90 mW/m^2. *Bethke and Marshak*'s [1990] calculation of heat and fluid transport assumed a background heat flow of 100 mW/m^2 at the time of brine migration (~290 Ma). One study [*Garven et al.*, 1993] has claimed that higher temperatures can be reached at lower heat flows due to thermal transients, but such transients have since been shown to be a numerical artifact related to improper boundary conditions (Nunn et al., 1996).

Could the background heat flow in the Arkoma Basin area have decreased by ~50 mW/m^2 in 290 Myr? The time constant for thermal relaxation of the lithosphere is ~ 50 Myr [*McKenzie*, 1978], thus sufficient time has elapsed to make such a cooling theoretically feasible. Studies of how the magnitude of heat flow changes with tectonic age also show that young tectonic provinces are hotter than older ones [*Lee and Uyeda*, 1965; *Polyak and Smirnov*, 1968; *Vitorello and Pollack*, 1980]. Heat flow decreases over time as tectonic activity wanes due to the decay of a transient thermal perturbation of unknown origin. Erosion also leads to a loss of radioactive heat-generating elements with a concomitant reduction in the surface heat flow [*Vitorello and Pollack*, 1980]. The average magnitude of heat flow decrease over 300 Myr due to these factors is about 30 mW/m^2. It is thus conceivable that heat flow in the Arkoma Basin area 300 Myr ago could have been as high as 100 mW/m^2 but perhaps unlikely. It is therefore difficult to describe our results as supportive of brine-migration theories which invoke topographically driven regional groundwater flow, as there is no evidence from present day thermal studies that the Arkoma Basin was an area of unusually high heat flow in Late Pennsylvanian-Early Permian time (~290 Ma).

Most simulations of topographically driven fluid flow through the Arkoma Basin have invoked a detailed hydrostratigraphy, although the permeability structure proposed has not been supported by measurement [e.g., *Garven et al.*, 1993]. It is not possible to make detailed inferences concerning the average permeability of specific geologic units or formations from a study of the thermal field. However, an understanding of how heat flow changes spatially does allow us to constrain the average permeability structure in a gross sense if unexplained heat flow differences are attributed to fluid flow. As the ratio of advective to conductive heat transport along the studied section appears to be no greater than 1/5 (Pe ~0.2), the implied permeability is lower than 10^{-15} m^2. This result, again, is not generally supportive of numerical simulations of ore formation by regional groundwater flow for which the Peclet number may approach unity (P_e ~1). It is possible that rock permeability may have decreased through time, and the topographic gradient driving hypothetical fluid migrations in Paleozoic time was almost certainly much greater than today; however, the inferred low permeability remains an obstacle to groundwater flow models which typically invoke permeabilities as high as 10^{-12} to 10^{-13} m^2 to explain hypothetical fluid migrations [*Garven et al.*, 1993, p. 524].

11. Conclusions

1. The average geothermal gradient along a north-south cross section through the Arkoma Basin and the Oklahoma Platform in southeastern Oklahoma is 23°C/km. The average geothermal gradient in the southern part of the basin is 20°C/km, exceeds 30°C/km in the middle part of the basin, and is 24°C/km in the northern part of the basin.

2. Mean background heat flow along a north-south cross section through the Arkoma Basin and the Oklahoma Platform in southeastern Oklahoma is 55±11 mW/m^2. On the southern end of the cross section, near the Ouachita Mountain front, the average heat flow value is ~44±9 mW/m^2. In the central part of basin, heat flow is

~58±12 mW/m^2, increasing to ~66±13 mW/m^2 on the northern end near the stable midcontinent platform.

3. Mean heat production of basement rocks in southeastern Oklahoma estimated from 17 gamma ray logs is 2.7±0.1 $\mu W/m^3$. Average heat production in the Arkoma Basin basement is 2.3±0.2 $\mu W/m^3$; the average heat production in basement rocks from the Oklahoma Platform is 2.8±0.1 $\mu W/m^3$.

4. About 10-25 mW/m^2 of the observed increase in heat flow (20-30 mW/m^2) from the Ouachita Front in the south to the Oklahoma Platform in north can be explained readily by a combination of heat refraction and lateral changes in heat production, as constrained by both measurements and knowledge of the regional geologic history and structure. Advective heat transport by regional groundwater flow, if present, is responsible for no more than 20 mW/m^2 of the observed heat flow increase from south to north. Application of a simple model constrains the regional scale permeability to be lower than 10^{-15} m^2.

5. Although it is probable that the thermal state has changed significantly over geologic time, present-day thermal data are generally not supportive of theories which invoke topographically driven groundwater flow through the Arkoma Basin in Late Pennsylvanian-Early Permian time (~290 Ma) to explain the genesis of Mississippi Valley-type lead-zinc deposits and regional diagenesis in the North American midcontinent. The present-day heat flow is relatively low compared to that apparently needed to satisfy constraints from fluid inclusions; and the present-day ratio of advective to conductive heat transport in the basin (Peclet number) is lower than 0.2.

12. References

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CHAPTER 2

Evaluation of thermal conductivity temperature corrections applied in terrestrial heat flow studies

1. Abstract

We collated 117 sets of temperature-dependent thermal conductivity measurements on minerals and rocks for the purpose of evaluating the comparative accuracy of different types of temperature corrections commonly applied in terrestrial heat flow studies. Data used consisted of measurements on 10 minerals, 59 igneous rocks, 19 sedimentary rocks, and 29 metamorphic rocks with a thermal conductivity range from 1.4 to 9.0 W/m-K at room temperature (20-25°C). The data set included thermal conductivites measured at temperatures from 0 to 500°C. When accuracy was bench-marked against the entire data set, corrections proposed by *Sekiguchi* [1984], *Somerton* [1992], *Sass et al.* [1992], *Funnell et al.* [1996], and *Chapman et al.* [1984] were found to have mean absolute relative errors of 6.9%, 7.7%, 8.5%, 14.2%, and 17.6%, respectively. The correction proposed by *Somerton* [1992] was found to have the lowest mean absolute relative error for minerals (11.3%) and sedimentary rocks (6.5%), while the correction proposed by *Sekiguchi* [1984] was found to have the lowest mean absolute relative error for igneous (5.4%) and metamorphic rocks (4.9%). The correction schemes were also found to differ in their systematic errors. Corrections proposed by *Sekiguchi* [1984], *Somerton* [1992], *Sass et al.* [1992], *Funnell et al.* [1996], and *Chapman et al.* [1984] were found to have mean relative errors of -3.3%, +2.6%, -5.8%, -11.0%, and +15.5%, respectively. Most of the temperature-dependent thermal conductivity data used to evaluate correction schemes were measurements on air-saturated rocks; however, the invariance of the relative rankings with respect to rock porosity suggests the rankings may be valid with respect to *in situ* conditions.

2. Introduction

In terrestrial heat flow studies, thermal conductivity measurements are usually made at room temperature and extrapolated to *in situ* temperature using some type of correction [*Kappelmeyer and Haenel*, 1974; *Sass et al.*, 1992; *Deming*, 1994; *Lee et al.*, 1996]. However, *in situ* thermal conductivities estimated by different types of temperature corrections can be quite different, even though the thermal conductivity at room temperature is the same (Figure 1). These differences can be large enough to cause serious errors in the estimation of heat flow values. For example, an average thermal conductivity of 3.0 W/m-K at 20°C is estimated to be 2.8 W/m-K at 200°C, if the correction proposed by *Funnell et al.* [1996] is used, but if the correction proposed by *Chapman et al.* [1984] is used, the estimated value at 200°C is 1.9 W/m-K. The object of this study is to evaluate temperature corrections frequently applied in terrestrial heat flow studies by comparing them with an extended suite of existing measurements.

3. Temperature Effect on Thermal Conductivity

The change of thermal conductivity with temperature depends on various factors. These include opacity, composition, and the shape, size, and orientation of crystals and their fragments [*Buntebarth*, 1991; *Clauser and Huenges*, 1995].

Thermal conductivity can be mainly divided into lattice conductivity and radiative conductivity. Lattice conductivity (or phonon conductivity) is produced by the diffusion of thermal vibration in a crystalline lattice, while radiative conductivity is

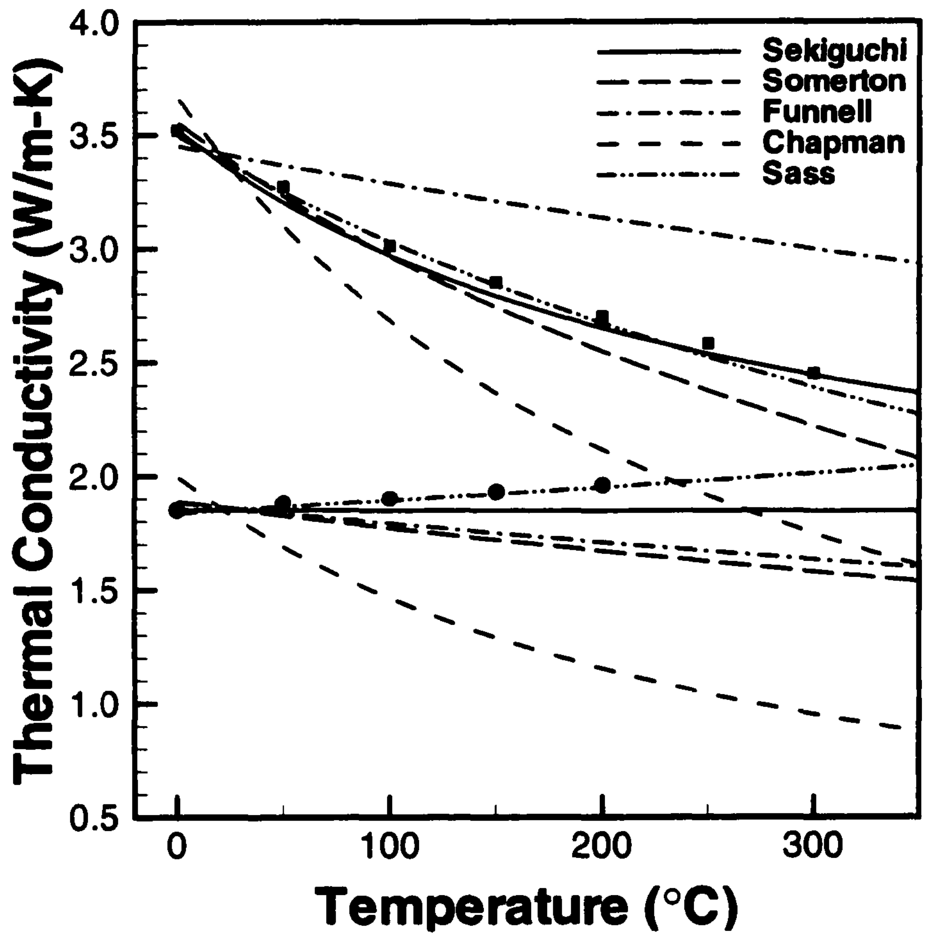

Figure 1. Temperature-dependent thermal conductivity measurements on a granite sample (squares, Birch and Clark [1940]) and an anorthosite sample (circles, Birch and Clark [1940]), and corresponding temperature corrections proposed by Sekiguchi [1984], Somerton [1992], Sass et al. [1992], Funnell et al. [1996], and Chapman et al. [1984].

produced by infrared electromagnetic waves [*Clark*, 1969; *Schatz and Simmons*, 1972; *Anderson*, 1989; *Poirier*, 1991]. Radiative conductivity is negligible in comparison to lattice conductivity at temperatures less than about 500°C [*Sibbitt et al.*, 1979]. Theoretically, lattice conductivity tends to be inversely proportional to absolute temperature ($\lambda \propto T^{-1}$), while radiative conductivity tends to be proportional to the cube of the absolute temperature ($\lambda \propto T^3$) [*Clark*, 1969; *Schatz and Simmons*, 1972; *Anderson*, 1989; *Poirier*, 1991]. Some authors have suggested that the change of lattice conductivity with increasing temperature is proportional to $T^{-5/4}$ [*Lubimova*, 1958; *McBirney*, 1963]. However, it is difficult to discriminate between theoretical models due to experimental errors [*Beck et al.*, 1978]. Strictly speaking, the T^{-1} relationship for lattice conductivity holds only for structurally perfect isotropic single crystals [*Roy et al.*, 1981; *Buntebarth*, 1991]. However, many rocks are composed of mixtures of highly disordered crystals of different compositions. For that reason, the thermal conductivity of rocks tends to decrease more slowly than T^{-1} and may tend to actually increase, in some cases, with increasing temperature [*Birch and Clark*, 1940; *Somerton*, 1992]. For example, the thermal conductivities of feldspar aggregates increase with increasing temperature [*Birch and Clark*, 1940], and the thermal conductivities of glasses and vitreous materials also increase with increasing temperature [*Ratcliffe*, 1959; *Somerton*, 1992]. Our survey of experimental data shows that the thermal conductivities of rocks with thermal conductivities less than about 2.0 W/m-K at room temperature tend to increase with increasing temperature, while those with thermal conductivities higher than about 2.0 W/m-K at room temperature tend to decrease with increasing temperature. Examples of rocks that behave in this manner include anorthosite [*Birch and Clark*, 1940], obsidian [*Cermak and Rybach*, 1982], porphyrite [*Cermak and Rybach*, 1982], shale [*Cermak and Rybach*, 1982], and sandstone [*Sugawara and Yoshizawa*, 1961, 1962; *Anand et al.*, 1973].

4. Data collection

We collated temperature-dependent thermal conductivity data reported by *Birch and Clark* [1940], *Sugawara and Yoshizawa* [1961, 1962], *Kawada* [1964], *Kanamori et al.* [1968], *Anand et al.* [1973], *Beck et al.* [1978], *Sibbitt et al.* [1979], *Cermak and Rybach* [1982], *Mongelli et al.* [1982], and *Petrunin and Popov* [1985]. A total of 117 sets of data, consisting of measurements on 10 minerals, 59 igneous rocks, 19 sedimentary rocks, and 29 metamorphic rocks measured at temperatures in the range 0 to 500°C, were used to evaluate different types of temperature corrections (see Figure 2). In some cases, documentation is sufficient so as to instill a high degree of confidence in the thermal conductivity data reported in the literature [e.g., *Birch and Clark*, 1940; *Anand et al.*, 1973; *Sibbitt et al.*, 1979; *Mongelli et al.*, 1982]. In other cases, however, methods may have been poorly documented, with inadequate or missing error discussions. Most of the thermal conductivity data used in this study were derived from measurements on dry (air or gas-saturated) rock samples. However, in some few cases, the condition (air or water-saturated) of the rock sample was not reported. It is conceivable that these types of uncertainties could have some effect on our conclusions. However, the possibility of some spurious data is offset by the averaging of errors through aggregation.

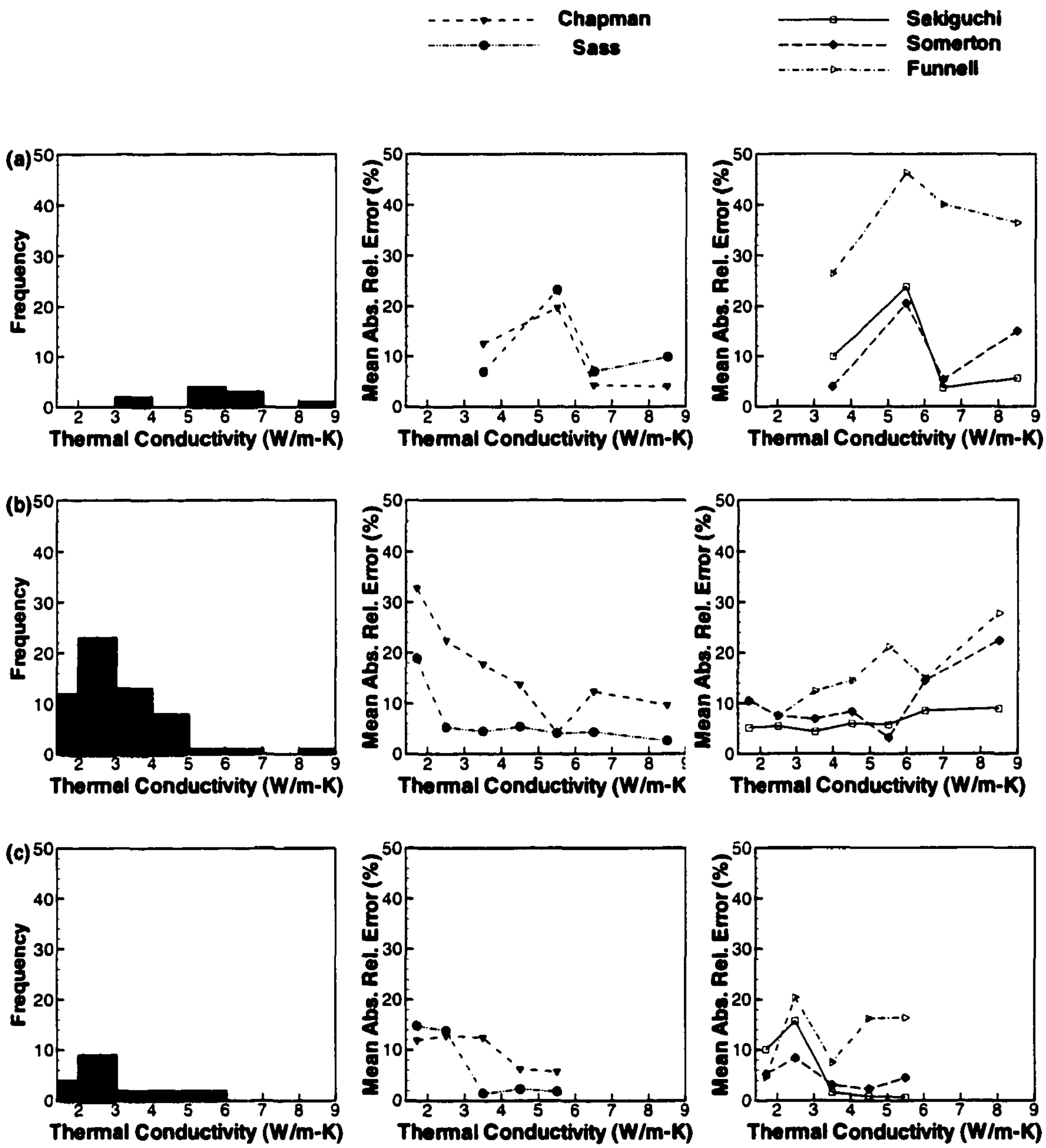

Figure 2. Histogram of thermal conductivity at 20 °C and mean absolute relative errors for temperature corrections proposed by Sekiguchi [1984], Somerton [1992], Sass et al. [1992], Funnell et al. [1996], and Chapman et al. [1984] for (a) minerals, (b) igneous rocks, (c) sedimentary rocks, (d) metamorphic rocks, and (e) all mineral and rocks. Errors calculated over the temperature range from 0 to 500 °C with respect to thermal conductivity (at 20-25 °C) and averaged over intervals of 1.4-2.0W/m-K, 2.0-3.0 W/m-K, 3.0-4.0 W/m-K, etc.

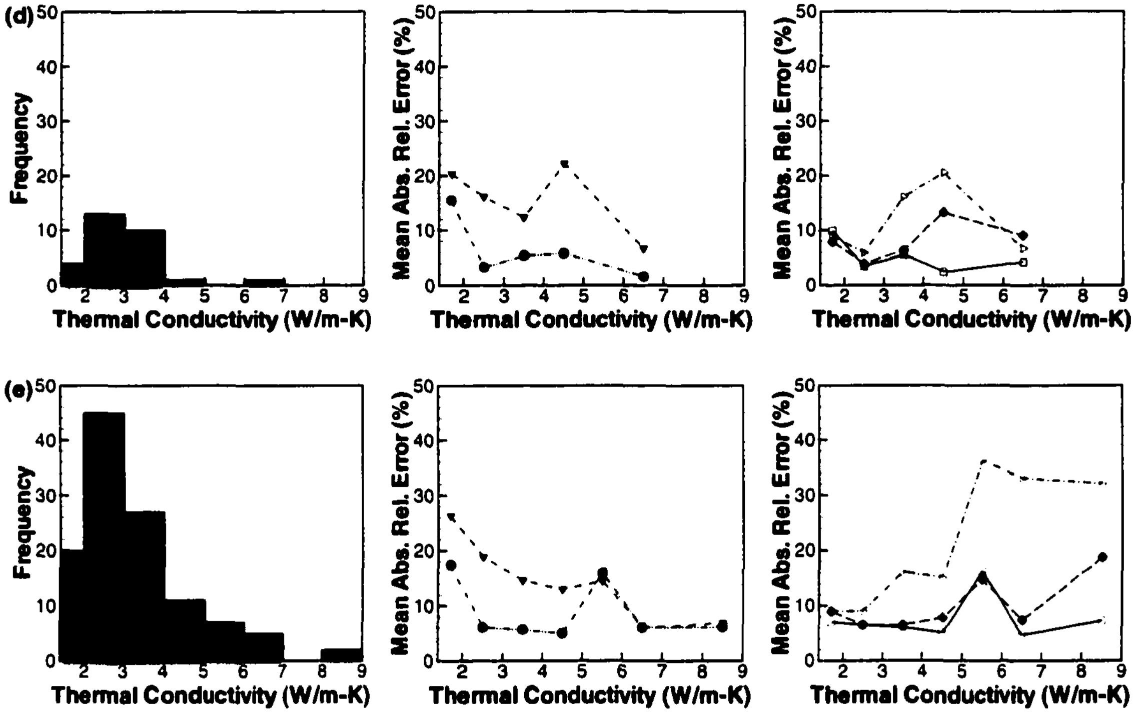

Figure 2. continued

5. Temperature Corrections

5.1 Funnell et al. [1996]

Funnell et al. [1996] proposed a T^{-1} type temperature correction,

$$\lambda(T) = \lambda_{20} \frac{1}{1 + 0.0005(T - 20)} \tag{1}$$

where λ_{20} is the thermal conductivity in W/m-K at 20°C and T is the estimated *in situ* temperature in degrees Celsius. This correction is for matrix thermal conductivity.

5.2 Sass et al. [1992]

Sass et al. [1992] suggested a temperature correction,

$$\lambda_0 = \lambda_{25}[1.007 + 25(0.0037 - \frac{0.0074}{\lambda_{25}})] \tag{2}$$

$$\lambda(T) = \frac{\lambda_0}{1.007 + T(0.0036 - \frac{0.0072}{\lambda_0})} \tag{3}$$

where λ_{25} is the conductivity in W/m-K at 25°C, λ_O is the inferred conductivity at 0°C using λ_{25}, and $\lambda(T)$ is the estimated conductivity in W/m-K at the estimated *in situ* temperature T in degrees Celsius. This correction is calibrated from some measurements on crystalline rocks made by *Birch and Clark* [1940] that have thermal conductivities in the range of 2.0 to 4.0 W/m-K at room temperature. For that reason, the correction proposed by *Sass et al.* [1992] is suitable only for crystalline rocks which have thermal conductivities greater than 2.0 W/m-K at 25°C, and is valid only up to 300°C.

5.3 Somerton [1992]

Somerton [1992] suggested a temperature correction modified from *Tikhomirov*'s [1968] correction after he found that the results of *Tikhomirov*'s [1968] correction were not satisfactory.

$$\lambda(T) = \lambda_{20} - 10^{-3}(T - 293) \times (\lambda_{20} - 1.38)$$

$$\times [\lambda_{20} \times (1.8 \times 10^{-3}T)^{-0.25\lambda_{20} + 1.28}] \times \lambda_{20}^{-0.64} \quad (4)$$

where $\lambda(T)$ is the thermal conductivity in W/m-K at the estimated *in situ* temperature T in Kelvins, and λ_{20} is the thermal conductivity in W/m-K at 20°C. The correction proposed by *Somerton* [1992] works only for rocks with conductivities less than about 9.0 W/m-K at a temperature of 20°C. For thermal conductivities greater than 9.0 W/m-K at 20°C, this correction shows unrealistic thermal conductivity variations.

5.4 Chapman et al. [1984]

Chapman et al. [1984] suggested a T^{-1} type of correction,

$$\lambda(T) = \lambda_{20} \frac{293}{273 + T} \quad (5)$$

where $\lambda(T)$ is the conductivity in W/m-K at the *in situ* temperature T in degrees Celsius, λ_{20} is the thermal conductivity in W/m-K at 20°C, and T is the estimated *in situ* temperature in degrees Celsius. This correction was proposed for matrix thermal conductivity. It has been used in heat flow studies by *Deming and Chapman* [1988a, b, 1989], *Correia et al.* [1990], and *Huang and Williamson* [1994].

5.5 Sekiguchi [1984]

Sekiguchi (1984) proposed the following temperature correction:

$$\lambda(T) = \lambda_m + [\frac{T_0 T_m}{T_m - T_0} \times (\lambda_0 - \lambda_m) \times (\frac{1}{T} - \frac{1}{T_m})] \tag{6}$$

where T is the estimated *in situ* temperature in Kelvins, λ(T) is the estimated thermal conductivity at the *in situ* temperature T, λ_m and T_m are the thermal conductivity and absolute temperature (Kelvins) at what *Sekiguchi* [1984] refers to as "the assumed point," and λ_o is the thermal conductivity at room temperature T_o. We used 1.8418 W/m-K and 1473°K for λ_m and T_m, respectively, as suggested by *Sekiguchi* [1984].

5.6 Zoth and Haenel [1988]

Zoth and Haenel [1988] suggested a temperature correction which follows the T^{-1} rule:

$$\lambda(T) = A + \frac{B}{350 + T} \tag{7}$$

where λ(T) is the thermal conductivity in W/m-K at *in situ* temperature T in degrees Celsius and A and B are constants determined by a least squares regression. *Zoth and Haenel* [1988] derived seven different coefficients (A and B) for different rock types. However, unlike some other correction procedures, the correction suggested by *Zoth and Haenel* [1988] has a single correction curve for a certain rock type, no matter what conductivity value the rock sample has at room temperature. For example, the temperature correction for limestones is

$$\lambda(T) = 0.13 + \frac{1073}{350 + T} \tag{8}$$

where $\lambda(T)$ is the thermal conductivity in W/m-K at the estimated *in situ* temperature T in degrees Celsius. The thermal conductivity from correction (8) at 20°C is 3.03 W/m-K. However, actual limestone conductivities used to determined constants A and B range from 1.7 to 4.3 W/m-K at room temperature [*Zoth and Haenel*, 1988, p. 451]. Therefore the correction proposed by *Zoth and Haenel* [1988] cannot be compared to corrections which are designed to extrapolate measurements made at room temperature (20-25°C) to *in situ* conditions.

6. Comparison of Temperature Corrections

The corrections proposed by *Sekiguchi* [1984], *Chapman et al.* [1984], *Sass et al.* [1992], *Somerton* [1992], and *Funnell et al.* [1996] were compared using thermal conductivity data measured in the temperature range corresponding to conditions in the upper crust (0~500°C). To estimate the errors inherent in these temperature corrections, the mean absolute value of the relative error (MARE) was calculated as

$$MARE(\%) = \frac{100 \times \dfrac{|\lambda_T - \lambda_C|}{\lambda_T}}{N} \tag{9}$$

where λ_T is the thermal conductivity measured at temperature T, λ_C is the thermal conductivity inferred from a temperature correction at the same temperature, and N is the total number of measurements.

When the entire data set was used to benchmark accuracy, the corrections proposed by *Sekiguchi* [1984] *Somerton* [1992] *Sass et al.* [1992], *Funnell et al.* [1996], and *Chapman et al.* [1984] were found to have MAREs of 6.9±0.4%, 7.7±0.3%, 8.5±0.7%, 14.2±0.7%, and 17.6±0.5%, respectively (Table 1), where the uncertainty is given as the standard error of the mean. The correction proposed by *Somerton* [1992] was found to have the lowest MARE for minerals (11.3±1.8%) and for sedimentary rocks (6.5±0.7%), while the correction proposed by *Sekiguchi* [1984] was found to have the lowest MARE for igneous (5.4±0.5%) and metamorphic rocks (4.9±0.7%). Relative errors varied between and amongst various rock types and correction procedures (Figure 2 and Table 2).

The error involved in applying all corrections increases with increasing temperature (Figure 3). Corrections proposed by *Chapman et al.* [1984] and *Sass et al.* [1992] apparently have less error when applied to rocks with relatively high thermal conductivities (>3.0 W/m-K at 20-25°C, see Figure 2e). The correction proposed by *Funnell et al.* [1996] has less error when applied to rocks with relatively low thermal-conductivities (<3.0 W/m-K at 20-25°C, see Figure 2e), while corrections proposed by *Sekiguchi* [1984] and *Somerton* [1992] tend to have relatively low errors over the entire range of rock thermal conductivity (1.4-8.0 W/m-K at 20-25°C, see Figure 2e).

The correction methods also differed in their systematic errors. The mean relative error (MRE) was calculated as

$$MRE(\%) = \frac{100 \times \frac{(\lambda_T - \lambda_c)}{\lambda_T}}{N} \qquad (10)$$

The corrections proposed by *Sekiguchi* [1984], *Somerton* [1992] *Sass et al.* [1992], *Funnell et al.* [1996], and *Chapman et al.* [1984] were found to have MREs of

Table 1. Mean Absolute Relative Error (MARE, ±S.E.) of Different Temperature Corrections of 0 to 500°C

Rock type	Minimum of k_{20}, [a]W/m-K	Maximum of k_{20}, [a]W/m-K	N[b]	*Sekiguchi* [1984]	*Somerton* [1992]	*Sass et al.* [1992]	*Funnell et al.* [1996]	*Chapman et al.* [1984]
Minerals	3.26	8.53	10	12.7±2.2	11.3±1.8	13.2±2.2	38.7±4.0	11.9±1.7
Igneous Rocks	1.42	8.16	59	5.4±0.5	8.5±0.4	8.0±1.2	10.8±0.6	21.9±0.8
Sedimentary Rocks	1.42	5.36	19	11.2±1.4	6.5±0.7	11.5±1.4	15.2±1.8	11.4±0.8
Metamorphic Rocks	1.69	6.04	29	4.9±0.7	5.5±0.5	5.6±0.8	10.5±1.1	15.1±0.8
All of the Minerals and Rock Types	1.42	8.53	117	6.9±0.4	7.7±0.3	8.5±0.7	14.2±0.7	17.6±0.5

[a]Thermal conductivity at 20°C.

[b]Number of data.

Table 2. Mean Absolute Relative Error (MARE, ±S.E.) of Different Temperature Corrections of 0 to 500°C by Rock Types

Rock	Minimun of k_{20},[a] W/m-K	Maximum of k_{20},[a] W/m-K	N[b]	*Sekiguchi* [1984]	*Somerton* [1992]	*Sass et al.* [1992]	*Funnell et al.* [1996]	*Chapman et al.* 1984]
Limestone	1.63	3.33	8	17.0±2.7	9.0±1.3	16.9±2.6	20.4±3.4	12.1±1.2
Dolomite	2.28	4.71	3	9.2±3.2	7.3±2.0	10.4±3.1	17.3±2.9	7.1±1.3
Sandstone	1.42	5.36	8	5.7±1.3	3.7±0.4	6.3±1.2	9.0±1.6	12.2±1.5
Granite	2.17	3.67	11	6.6±1.8	7.1±1.2	6.7±1.9	13.8±1.9	15.9±1.3
Diorite	2.05	3.48	6	4.1±0.9	5.1±1.1	4.2±1.0	6.0±1.0	20.7±2.2
Gabbro	1.92	3.21	4	3.5±0.6	11.0±1.7	2.6±1.1	5.2±1.1	29.7±3.4
Dunite	3.61	8.16	9	5.5±0.7	8.1±1.1	3.7±0.4	18.9±1.6	10.5±1.1
Amphibolite	2.23	2.82	4	0.88±0.1	3.32±0.4	0.92±0.2	2.80±0.4	17.5±1.7
Slate	1.90	2.61	3	6.8±2.5	4.9±1.6	8.8±2.9	5.0±1.7	20.2±3.5
Marble	2.85	2.91	3	8.0±1.5	4.9±0.6	7.1±1.3	13.9±2.3	7.8±1.6
Quartzite	1.89	6.04	3	14.2±3.2	10.6±2.1	15.5±3.7	20.6±3.3	7.7±1.1
Gneiss	2.13	3.66	8	1.3±0.1	2.7±0.3	0.7±0.1	8.1±0.8	12.6±1.0

[a] Thermal conductivity at 20°C.

[b] Number of data.

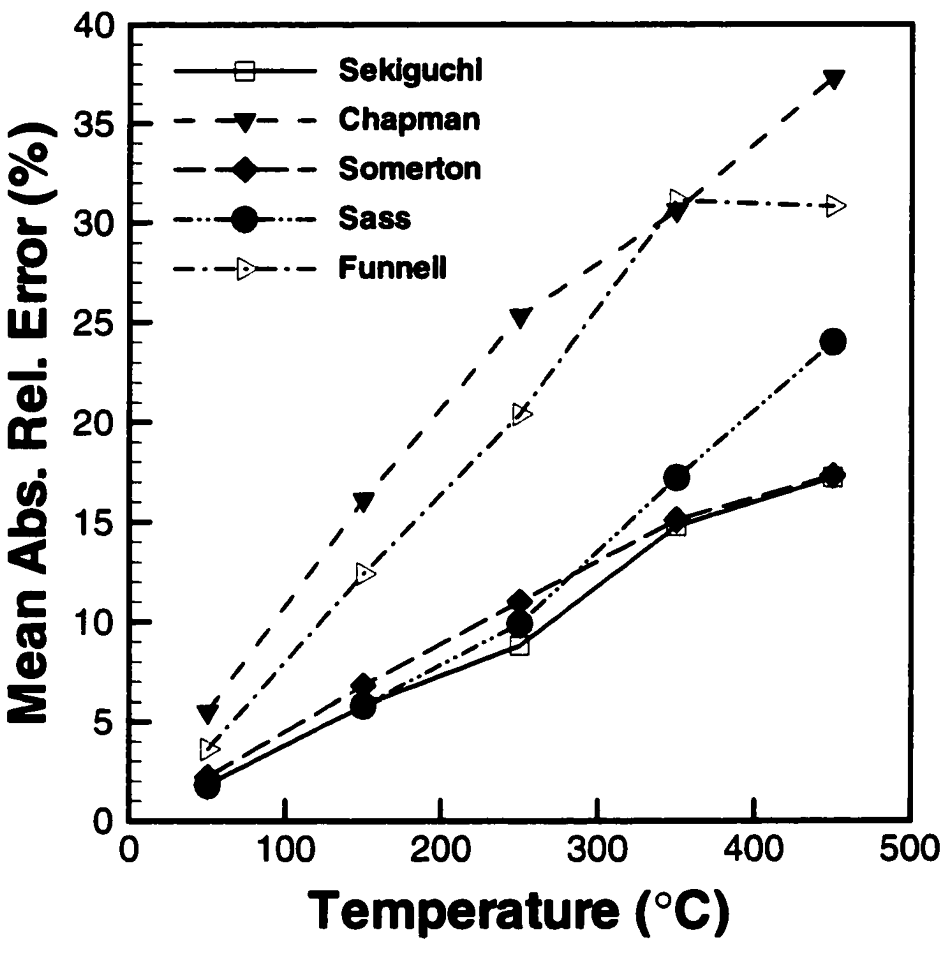

Figure 3. Mean absolute relative error for different types of temperature corrections over temperature range from 0 to 500 °C averaged over 100 °C temperature intervals. Errors calculated by comparing proposed temperature corrections with the entire data set, representing a total of 117 samples consisting of minerals, as well as sedimentary, metamorphic, and igneous rocks.

-3.3±0.5%, +2.6±0.4%, -5.8±0.7%, -11.0±0.8%, and +15.5±0.6%, respectively, where the uncertainties are given as the standard error of the mean (Figure 4). The temperature correction proposed by *Funnell et al.* [1996] tends to systematically overestimate *in situ* thermal conductivity, while the temperature correction proposed by *Chapman et al.* [1984] tends to systematically underestimate *in situ* thermal conductivity (Figure 4).

7. Discussion

The errors evaluated in this study represent both limitations of the generalized corrections and measurement errors in the data set. In consideration of the probable variation of thermal conductivity between and among rock types, it is perhaps unlikely that any generalized temperature correction would be a significant improvement over the better corrections that were evaluated.

We considered the possibility that our relative rankings may have been skewed by a few outlying data points. In a few cases, there are some large discrepancies between measured and predicted thermal conductivities. However, when these data were removed from the bench-marking evaluation, the magnitude of the errors involved did not change significantly, nor did the relative ranking of the different corrections change.

The data set used in this study as a basis for evaluating temperature corrections consists nearly entirely of measurements on gas or air-saturated rocks (J. Sass, personal communication, 1997). There is thus some question as to the applicability of

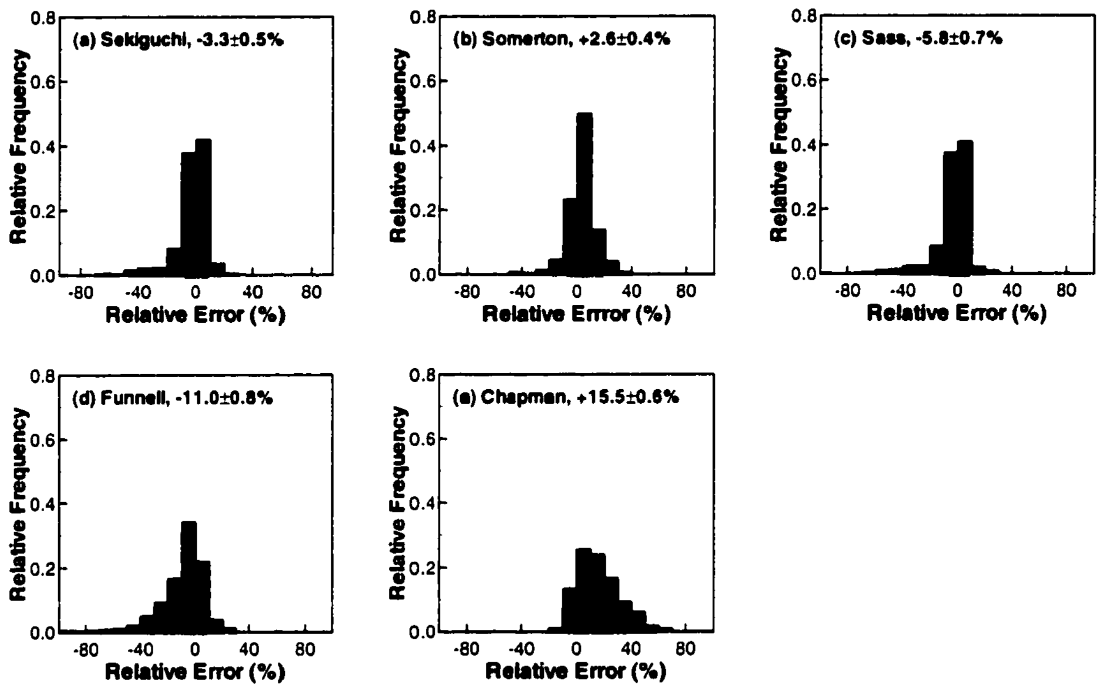

Figure 4. Histogram of relative errors of temperature corrections proposed by (a) Sekiguchi [1984], (b) Somerton [1992], (c) Sass et al. [1992], (d) Funnell et al. [1996], and (e) Chapman et al. [1984] for all minerals and rocks. Errors (%) represent mean relative errors and corresponding standard errors of means of temperature corrections.

our results to *in situ* conditions where rock pores are usually saturated with an aqueous fluid. It would have been preferable to use measurements on water-saturated rocks to bench-mark the correction schemes evaluated in this study. However, there exist only a handful of such measurements, too few to be representative of nature's lithological spectrum [*Pribnow et al.*, 1996; *Williams and Sass*, 1996]. At this point in time we do not believe sufficient data exist to justify the derivation of a generalized empirical correction or verify a theoretical model for water-saturated rocks.

A consequence of temperature-dependent measurements on air-saturated rocks is that heating stresses induce fractures which artificially increase porosity during the measurement process [*Pribnow et al.*, 1996]. As the additional porosity is filled with low-conductivity air (0.02 W/m-K at 0°C, 0.06 W/m-K at 500°C), this experimental artifact will tend to lead to measured thermal conductivities at high temperatures that are too low. Thus empirical corrections derived from air-saturated rocks may overcorrect at high temperatures. However, at the present time we do not believe enough data exist to determine the significance of this effect. *Pribnow et al.* [1996] found that when dry rocks were heated from 20 to 300°C, the porosity increase due to thermal cracking was less than 1%. The single rock sample for which *Pribnow et al.* [1996] reported thermal conductivity measurements in both the air- and water-saturated states exhibited a larger change of thermal conductivity in the air-saturated state, but the difference was relatively small. The water-saturated conductivity of this rock decreased by 10% from 20 to 300°C, while the air-saturated conductivity decreased by 12% over the same temperature range.

Because our conclusions and relative rankings are based on data for air-saturated rocks, it might also be argued that those correction schemes that fared poorly in our comparisons might actually work better for pure matrix or for water-saturated rocks and therefore be more representative of *in situ* conditions. However, none of the correction schemes we reviewed were theoretically formulated to explicitly include water versus air saturation. Empirically, the argument also breaks down. The relative importance of pure matrix versus air versus water saturation increases with rock porosity. Therefore a correction scheme that works with pure matrix or water-saturated rocks but not with air-saturated rocks should perform best with low-porosity igneous and metamorphic rocks and worst with sedimentary rocks. However, the relative ranking of the correction schemes we evaluated (Table 1) is nearly invariant between rock types. Those schemes that fare the worst when compared with data from high-porosity sedimentary rocks also perform relatively poorly when tested against data from low-porosity igneous and metamorphic rocks.

8. Conclusions

1. Rock and mineral thermal conductivities higher than about 2.0 W/m-K at room temperature (20-25°C) tend to decrease with increasing temperature, while those with thermal conductivities less than about 2.0 W/m-K at room temperature (20-25°C) tend to increase with increasing temperature.

2. Thermal conductivity temperature corrections differ in their accuracy. When all available temperature-dependent thermal conductivity data are used as an accuracy benchmark, the corrections proposed by *Sekiguchi* [1984], *Somerton* [1992], *Sass et*

al. [1992], *Funnell et al.* [1996], and *Chapman et al.* [1984] were found to have mean absolute relative errors of 6.9%, 7.7%, 8.5%, 14.2%, and 17.6%, respectively.

3. The magnitude of error increases with increasing temperature for all correction procedures.

4. Thermal conductivity temperature corrections differ in their systematic error. The corrections proposed by *Sekiguchi* [1984], *Somerton* [1992], *Sass et al.* [1992], *Funnell et al.* [1996], and *Chapman et al.* [1984] were found to have mean relative errors of -3.3%, +2.6%, -5.8%, -11.0%, and +15.5%, respectively.

5. Nearly all of the temperature-dependent thermal conductivity measurements which exist at the present time represent measurements on air-saturated rocks which may not be representative of the *in situ* fluid-saturated state. However, the relative rankings of the correction schemes evaluated in this study are probably valid, because the relative rankings do not change with respect to rock porosity.

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CHAPTER 3

Heat flow and Thermal History of the Anadarko Basin and the Western Oklahoma Platform

1. Abstract

The average geothermal gradient in the Anadarko Basin and the Oklahoma Platform estimated from 856 corrected bottom-hole temperatures (BHTs) is 21 °C/km. Analysis of previously published thermal maturation data indicates that the Anadarko Basin has undergone from 1 to 3 km of erosion starting about 40 to 50 Ma. The average thermal gradient at the time of maximum burial was in the range of 22 to 25 °C/km. There is no evidence that the thermal state of the Anadarko Basin has changed significantly since the late Paleozoic (250 Ma).

To estimate heat flow in the Anadarko Basin and the Oklahoma Platform, we made 652 thermal conductivity measurements on drill cuttings from 9 oil and gas wells. All measurements were corrected for the effects of anisotropy, temperature, and porosity. Matrix conductivities perpendicular to bedding at 22 °C range from 1.7 W/m-K for Permian age sandstone and shale, to 2.6 W/m-K for Devonian-Silurian age limestone and shale. *In situ* thermal conductivities vary from 1.3 W/m-K to 1.9 W/m-K. Estimated heat flow (±20%) ranges from 30 mW/m^2 to 42 mW/m^2, with a mean of 36 mW/m^2. We found no evidence for heat flow to increase significantly from the Anadarko Basin in the south to the Oklahoma Platform to the north. These results contradict a previous study which found heat flow increases from south to north, with heat flow in the Oklahoma Platform as much as 50% greater than heat flow in the Anadarko Basin. At the present time, we do not know if these differences are related to spatial variations or reflect methodological errors. If our results are averaged with those reported in previous work, mean heat flows in the Anadarko Basin and Oklahoma Platform are 39 mW/m^2 and 51 mW/m^2, respectively.

2. Introduction

The Anadarko Basin in southwest Oklahoma (Figure 1) is noteworthy for several reasons. The Anadarko is the deepest sedimentary basin located on the North

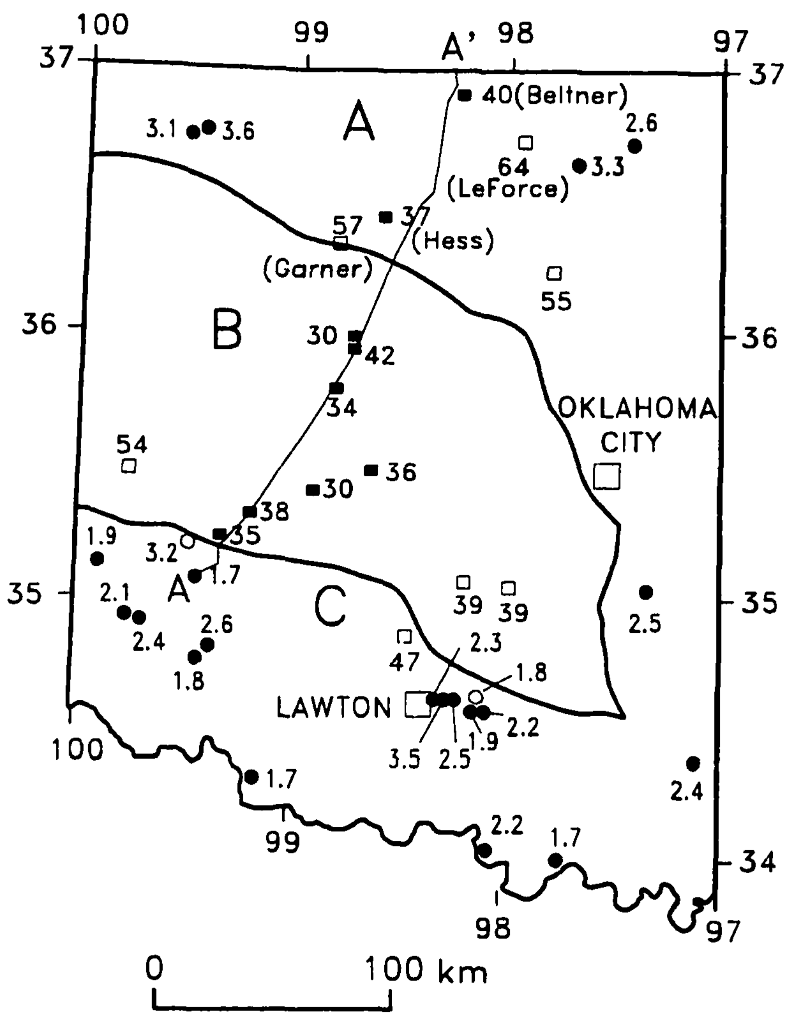


Figure 1. Heat flow and heat production estimates in the Anadarko Basin and Oklahoma Platform. Solid rectangles represent heat flow estimates in this study; open rectangles represent heat flow estimates made by Carter et al. (1998); solid circles represent heat production estimates made from gamma ray logs (Cranganu et al., 1998); open circles represent heat production measurements made by Borel (1995). "A" represents the Oklahoma Platform; "B" represents the Anadarko Basin; "C" represents the Wichita Mountains. Cross section A-A' refers to Figure 2.

American craton, with an existing in-place sedimentary section that is more than 12 km thick. It is also a prolific producer of both petroleum, and natural gas. Cumulative hydrocarbon production as of 1985 was > 5 million barrels of oil and 82 trillion cubic feet of natural gas (Davis and Northcutt, p. 13, 1989).

In addition to being an important oil and gas province, the Anadarko Basin is also a hydrologic anomaly. Although it is an old and tectonically senescent basin, the Anadarko is known to contain abnormally high fluid pressures (Hunt, 1990; Al-Shaieb et al., 1994). The most rapid phase of sedimentation in the basin ended nearly 250 million years ago, and it has been in a state of more-or-less continuous erosion for the last 50 million years. However, fluid pressures approaching lithostatic are commonly encountered in the deep Anadarko Basin. These high fluid pressures are anomalous. Overpressured fluids are more commonly associated with young, actively subsiding sedimentary basins undergoing compaction disequilibrium (e.g., Bredehoeft and Hanshaw, 1968; Sharp and Domenico, 1976; Bethke, 1986; Mello et al., 1994).

Both hydrocarbon generation and overpressuring in the Andarko Basin are partially controlled or affected by its thermal history. The thermal maturation of kerogen into oil and gas as represented by standard kinetic models exhibts an exponential dependence upon temperature. Similarly, if oil or gas generation is implied in overpressuring, the rate at which fluid pressuring occurred in the Anadarko Basin may be temperature-dependent. Temperature is also a critical variable for diagenetic processes such as mineral precipitation which may change the permeability of rocks in the Anadarko Basin. Obtaining an understanding of the thermal history of the Anadarko Basin is therefore crucial both for reasons related to the economic exploitation of oil and gas, as well as to improve our theoretical understanding of how anomalous fluid pressures in the continental crust are generated and dissipated.

The logical starting point in reconstructing the thermal history of a sedimentary basin is the present day. The present day is the only point in time at which temperature may be measured rather than estimated. If the present day thermal state cannot be accurately estimated, it is unlikely that the thermal history can be reconstructed with any greater degree of reliability. The most useful and fundamental parametization of the thermal state of the crust is in terms of heat flow. Unlike the thermal gradient, heat flow is conserved. However, heat flow studies in sedimentary basins may be more difficult than in areas where crystalline rocks are exposed at the surface. High-quality temperature data may be rare to nonexistent, and sedimentary rocks tend to have higher permeabilities than crystalline rocks and therefore be more susceptible to hydrologic disturbances.

In this study, we characterize the present day thermal state of the Anadarko Basin using abundant, but low-quality temperature data from well logs made in oil and gas wells. Our results are constrasted with recent studies by Carter et al. (1998) and Gallardo and Blackwell (1999) based upon different methodologies. We explore inconsistencies and apparent discrepancies which arise when contrasting results are compared. Although we derive some important constraints on both the present and past thermal state of the Anadarko Basin, one of our primary conclusions is to underscore the uncertainties which are present in these types of studies.

3. Temperature

To estimate subsurface temperatures (Figure 2) in the Anadarko Basin and the Oklahoma Platform, we collected 856 bottom-hole temperatures (BHTs) from log headers of 589 oil and gas wells located in the Anadarko Basin and the Oklahoma Platform to the north. The temperature data were corrected using a Horner plot and average depth-time correction (Deming, 1989). Our methodology was identical to that used by Lee et al. (1996) and discussed in several other papers (Deming, 1989; Deming et al., 1990; Lee et al., 1994). The average annual air temperature in the

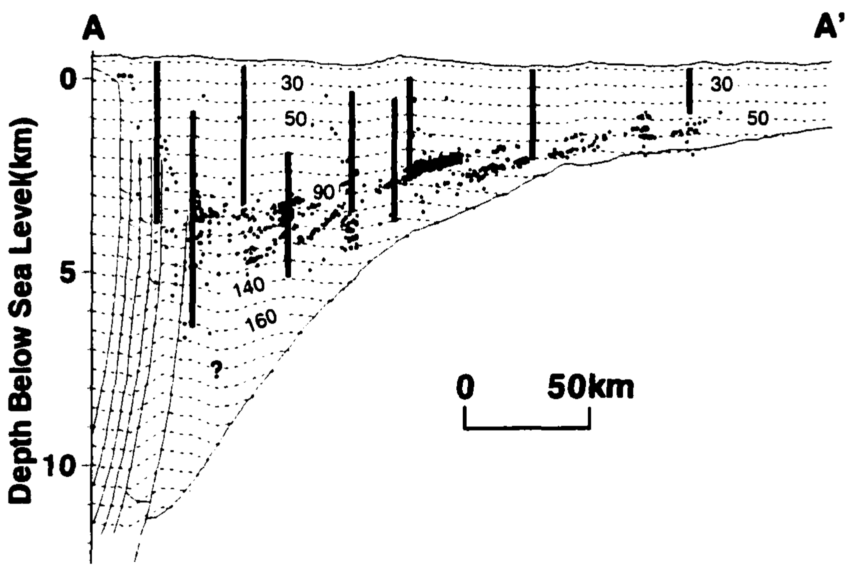


Figure 2. Estimated temperature (degrees Celsius) along a north-south cross section (see figure 1) though the Anadarko Basin and Oklahoma Platform. Dashed lines show estimated temperature in degrees Celsius. Dots show location of temperature measurements, vertical lines show sampling intervals for thermal conductivity measurements.

vicinity of the Anadarko Basin is 15.8 °C. We assumed that the average ground temperature was ~4 °C warmer than the air temperature, as has been found in numerous geothermal studies (e.g., Blackwell et al., 1980; Harris and Chapman, 1995). Our estimate of average ground temperature may be in error by perhaps 1-3 °C. However, this magnitude of error is lower than that usually associated with corrected BHTs. It is thus useful to use the mean annual ground temperature as a "known" boundary condition for calculating average geothermal gradients.

We found no significant lateral temperature variation throughout the Anadarko Basin and the Oklahoma Platform (Figure 2). The average geothermal gradient in the south of the Anadarko Basin is 21 °C/km; the average geothermal gradient to the north of the Anadarko Basin in the Oklahoma Platform is 23 °C/km. When data from the entire basin are averaged together, a least-squares regression on 856 corrected BHTs yields a basin-wide average thermal gradient of 21 °C/km. The overall average is strongly influenced by data from the deep basin in the south where thermal gradients are 21 °C/km.

The relatively monotonic thermal state we observe (Figure 2) in the Anadarko Basin is not an artifact of our correction or analysis methods. If lateral variation in thermal state were evident in the data, our analysis would have picked it up. A recent analysis of a similar data set from the Arkoma Basin in southeastern Oklahoma using identical methods found a pronounced variation in thermal gradients along a north-south cross-section (Lee et al., 1996). The lack of a similar variation in the Anadarko Basin is a reflection of the thermal state, not an artifact of over zealous averaging.

It is difficult to verify our estimates of *in situ* temperature. High-precision, equilibrium temperature logs in sedimentary basins are rare. Thus there is no absolute yardstick against which to compare estimates made by different methods. Confidence in our temperature estimates comes from an understanding of the physics of heat transport and fluid movement during drilling and BHT equilibration, as well as statistical assumptions regarding the nature and approximate magnitude of the error in our data (see discussion in Lee et al., 1996).

Carter et al. (1998) and Carter (1993) report precision temperature logs from seven sites in the Anadarko Basin that were made sometime after drilling operations ceased. Although these logs may not have measured thermal conditions which were completely in thermal equilibrium, they were regarded by Carter et al. (1998) as of sufficient quality to allow moderately precise calculations of interval thermal gradients and heat flows. A comparison of these temperature profiles with temperature predicted from BHTs reveals differences of ±5 °C (Figure 3). It is difficult to draw any meaningful conclusions from these comparisons. The temperature logs made and reported by Carter (1993) and Carter et al. (1998) are tens of kilometers away from the cross-section along which we estimated subsurface temperature (Figure 1). Differences may therefore be related to geographic variation. Vertical differences may be due to the averaging of BHT data through a model which does not allow vertical variation in thermal gradient, or they may reflect non-equilibrium conditions in the precision temperature logs. For example, the precision temperature logs reported by Carter et al. (1998) both show curvature near the bottom of each well suggestive of a lingering cooling effect in the bottom half of these boreholes.

4. Thermal conductivity

To estimate heat flow in the Anadarko Basin and the Oklahoma Platform, we made 652 thermal conductivity measurements on drill cuttings from 9 oil and gas wells (Figure 2) using the cell method of Sass et al. (1971). All measurements were corrected for the effects of anisotropy (Deming, 1994), temperature (Lee and Deming, 1998), and porosity, to estimate *in situ* thermal conductivity (Table 1). Our methods were identical to those used and reported by Lee et al. (1996). Matrix conductivities

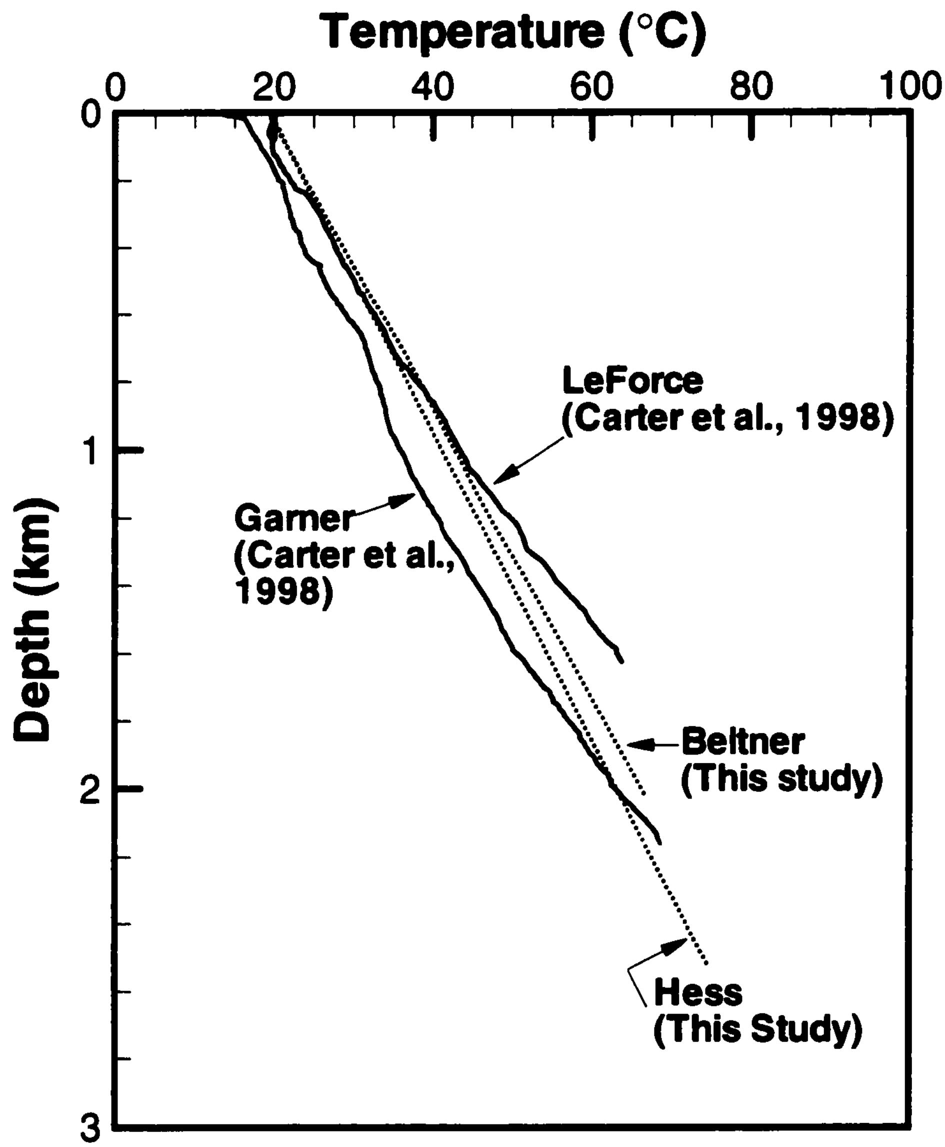


Figure 3. Temperature-depth profiles from corrected BHTs (this study) and measured by Carter et la. (1998). Dotted lines show estimated temperatures at the Beltner and the Hess well (this study) from corrected BHTs; solid lines show temperatures measured at the LeForce and the Garner wells by Carter (1993) and Carter et al. (1998). See Figure 1 for well locations.

Table 1. Thermal Conductivity and Porosity Data

Geologic Unit	Lithologies	N[a]	Matrix Conductivity[b] W/m-K	Standard Deviation[c] W/m-K	Standard Error[d] W/m-K	Porosity[e] %
Permian System	Sandstone, shale, Evaporite	223	1.67	0.86	0.06	11.2
Pennsylvanian System	Shale, Limestone, Granite Wash	292	1.90	0.60	0.04	10.5
Mississippian System	Limestone, shale	61	1.94	1.08	0.14	3.7
Devonian-Silurian Systems	Limestone, Shale	32	2.61	0.67	0.12	3.6
Ordovician-Cambrian Systems	Limestone, Sandstone, Shale	44	2.55	1.36	0.21	2.4

[a] Number of measurements

[b] Harmonic mean perpendicular to bedding at 22°C

[c] From the arithmetic mean

[d] Of the arithmetic mean

[e] Estimated from density logs and matrix density measurements

perpendicular to bedding at 22 °C range from 1.7 W/m-K for Permian age sandstone and shale, to 2.6 W/m-K for Devonian-Silurian age limestone, and shale (Table 1). Estimated *in situ* thermal conductivities vary from 1.3 W/m-K to 1.9 W/m-K (Table 2).

It is difficult to compare our results with those reported by Carter et al. (1998). Carter et al. (1998, p. 298) reported results categorized by lithology, not geologic age. This approach is not practical in our case, as drill cuttings typically contain a mixture of lithologies when collected from all but the most homogeneous formations. We measured aggregates of rock chips with random orientations, while Carter et al. (1998) measured consolidated core samples which in most cases were water-saturated. Core measurements are generally preferable, but are subject to limited availability and sampling bias. Shales (generally, the most common sedimentary rock) and other incompetent, low-permeability rocks are usually not cored when oil and gas wells are drilled.

5. Heat flow

5.1 Methodology

We estimated heat flow (Figure 1 and Table 2) by multiplying estimated *in situ* thermal conductivities (Figure 2) by average thermal gradients estimated from temperatures at the top and the bottom of the interval where thermal conductivities were measured. Heat flow was estimated only in areas where we had overlapping temperature and thermal conductivity data.

5.2 Results

Heat flow in the Anadarko Basin and Oklahoma Platform estimated from corrected BHTs varies from 30 to 42 mW/m^2, with an average of 36 mW/m^2 (Figure 1, Table 2). When these new heat-flow estimates are combined with previous

Table 2. Heat Flow Summary

Well name	Section, Township and Range	Latitude °N	Longitude °W	In-situ Conductivity[a] W/m-K	Geothermal Gradient °C/km	Heat Flow[b] mW/m^2
Counts #1	29-9N-20W	35.22°	99.33°	1.70	21	35
Bertha Rogers	27-10N-19W	35.31°	99.19°	1.82	21	38
Goeringer #1	28-11N-16W	35.40°	98.90°	1.45	21	30
Helena Dick #36-1	36-12N-14W	35.47°	98.63°	1.68	22	36
Dobbins #1	9-15N-15W	35.79°	98.80°	1.58	22	34
Marshall #1-19	19-17N-14W	35.94°	98.73°	1.90	22	42
Feikes #1	31-18N-14W	35.99°	98.73°	1.33	22	30
Hess #1	33-23N-13W	36.43°	98.59°	1.67	22	37
Beltner #1	13-28N-10W	36.91°	98.23°	1.72	23	40

[a] Harmonic mean of measurements after corrections for anisotropy, temperature, and porosity;

[b] Estimated error is ±20%

estimates made by Carter et al. (1998), the average heat flow in the Oklahoma Platform is 51 mW/m^2, while that in the Anadarko Basin and the Wichita Mountains is 39 mW/m^2.

We estimate the error in our heat flow estimates to be ±20%, using assumptions and methods identical to those used by Lee et al. (1996). Globally, the average heat flow in terrains with basement rocks of Middle Proterozoic Age (1300-1600 Ma, see Denison et al., 1984, and Johnson, 1989) is 50 mW/m^2 (Vitorello and Pollack, 1980). Heat flow in the Anadarko Basin and the Oklahoma Platform is thus somewhat low when compared to other areas of similar tectonothermal age.

6. Interpretation

If we accept the validity of both this and previous work, simple averaging of available heat flow estimates indicates there is an increase in heat flow of 12 mW/m^2 from the Anadarko Basin in the south to the Oklahoma Platform in the north. We agree with Carter et al. (1998) that the most likely explanation for this change is a difference in crustal heat production. No measurements or estimates of heat production for basement rocks beneath the Anadarko Basin are available due to their burial underneath an extraordinary thickness of sedimentary rocks (> 12 km). However, average heat production in basement rocks to the south of the Anadarko Basin is 2.2 ±0.1 $\mu W/m^3$, while average heat production in the basement rocks of the Oklahoma Platform is 3.0 ±0.2 $\mu W/m^3$. This difference may be sufficient to account for most of the observed 12 mW/m^2 increase in heat flow from south to north. For example, extrapolation of a heat-production difference of 0.8 $\mu W/m^3$ over an enriched layer 10 km thick would result in a net difference in surface heat flow of 8 mW/m^2. As discussed by Carter et al. (1998) and Cranganu et al. (1998), there is also anciliary geological and geophysical evidence for mafic basement rocks (which tend to have low heat production) in the Wichita Mountains just to the south of the Anadarko Basin.

Heat refraction from the relatively low thermal-conductivity rocks of the Anadarko Basin (most less than 2.0 W/m-K) into the more conductive crystalline rocks to the north and south could also have some small effect on the north-south distribution of heat flow. Carter et al. (1998) modeled heat refraction in the Anadarko Basin, and concluded that the magnitude of any possible effect was about 2-3 mW/m^2. This is not terribly large, but when a refraction effect (3 mW/m^2) is combined with the estimated effect of heat production (8 mW/m^2), the resulting change in heat flow (11 mW/m^2) is close to that observed (12 mW/m^2).

Groundwater movement can influence the thermal regime of a sedimentary basin (e.g., Deming et al., 1992). However, there is no geological evidence for regional groundwater flow in the vicinity of the Anadarko Basin. The permeability of the Upper Mississippian-Upper Cambrian aquifer system was inferred by Jorgensen et al., (1993, p. 48) to be relatively low. Furthermore, the documented existence of regional overpressuring in the deep Anadarko Basin (e.g., Al-Shaieb et al., 1994) apparently precludes the possibility of significant fluid flow.

7. Burial History of the Anadarko Basin

The Anadarko Basin has a complex history, starting as a rift basin with the formation of the Southern Oklahoma Aulacogen in late Proterozoic-early Cambrian time (~900-540 Ma). During the Wichita orogeny in Pennsylvanian time (~300-275 Ma), a foreland basin developed by compression and tectonism with the relatively rapid accumulation of ~10 km of sediments in ~10-20 million years (Gilbert, 1992). During Permian time (~286-245 Ma), as much as ~2 km of sediments were deposited in the Anadarko Basin (Johnson, 1989).

The history of Paleozoic sedimentation is relatively well known, because most of the sediments deposited during this era are retained today. However, the post-Paleozoic burial history of the Anadarko Basin is more controversial and less well constrained, because most Mesozoic and Cenozoic strata have been eroded. Different authors have made different estimates of the magnitude and the timing of sedimentary strata deposited and subsequently eroded during Mesozoic and/or Cenozoic time (e.g., Waples, 1982; Schmoker, 1986; Johnson, 1989; Gilbert, 1992; Carter et al., 1998).

There is geologic evidence for sedimentary deposition in the Anadarko Basin during the Mesozoic, but it is difficult to precisely constrain the magnitude and timing of sedimentary events. Extension of the Cretaceous seaway across the Anadarko Basin during the last great inundation of the western interior of the United States implies that sedimentary strata were deposited in a marine environment during Cretaceous time (Garner and Turcotte, 1984; Johnson, 1989). Remnants of Triassic, Jurassic, Cretaceous, and Tertiary sedimentary strata (~200-500 m) are found today in the western parts of Anadarko Basin and the Oklahoma Panhandle (Johnson et al., 1972; Johnson, 1989). The Denver and Raton Basin, which are located between the Rocky Mountains and the Anadarko Basin, probably had the same Cretaceous sediment provenance as the Anadarko Basin, and these basins contain preserved Cretaceous sedimentary strata with thicknesses that range from 560 to 2890 m (Baars et al., 1988).

When the thickness of eroded strata is uncertain, paleothermal indicators of various types can be utilized to obtain quantitative constraints on burial histories. Both vitrinite reflectance (R_O) and apatite fission-track (AFT) data from the Anadarko Basin have been published and have been used to interpret the erosional history.

A few studies have attempted to estimate the magnitude of erosion in southwest Oklahoma from vitrinite reflectance data (e.g., Waples, 1982; Schmoker, 1986; Cardott, 1989; Pawlewicz, 1989), however these studies are in poor agreement with each other. Results are different for different data from different areas, and may change significantly if some data are removed from larger sets. Possible problems include the reworking of older vitrinite, contamination of drill cuttings by caving (Feazel and Aram, 1990), and the suppression of vitrinite reflectance in strata such as the Woodford Shale (Lewan, 1985).

Cardott (1989) analyzed vitrinite reflectance data from the Woodford Shale in the Anadarko Basin and the Oklahoma Platform, and suggested that ~630 m of erosion occurred. Schmoker's (1986) analysis of vitrinite reflectance data indicated that ~790 m of erosion had occurred, while Pawlewicz (1989) interpreted vitrinite reflectance data from the eastern Anadarko Basin and the Oklahoma Platform as representing ~2.5 km of post-Paleozoic erosion.

Some of the differences between the above estimates may be related to suppression of vitrinite reflectance in the Woodford Shale (Buchardt and Lewan, 1990). Suppression of vitrinite reflectance refers to a phenomenon whereby the maturation state in hydrogen-rich kerogens (as measured by vitrinite reflectance) is lower than would otherwise be the case for other kerogens. Suppression is usually significant only for R_0 values less than 2.0% and tends to increase as the hydrogen index increases (Lo, 1993).

We used the model of Lo (1995) to correct previously reported measurements of vitrinite reflectance in the Woodford Shale. The Lo (1995) correction depends upon the hydrogen index of immature kerogen. For this reason, the hydrogen index of the Woodford Shale had to be estimated from exposed or shallow immature samples at the margins of the Anadarko Basin. The hydrogen index of the Woodford Shale varies from 610 mg/g for samples from the Arbuckle Mountains to the south of the Anadarko Basin, to 470 mg/g for samples from the western part of the Oklahoma Platform to the north of the Anadarko Basin (Comer, 1992; Kirkland et al., 1992; Cardott and Chaplin, 1993). In our corrections, we used an average value of 540 mg/g for the hydrogen index, and calculated the uncertainty introduced by the hydrogen index varying between the full range of 470 to 610 mg/g.

After making corrections for suppression effects, we reinterpreted the vitrinite data reported by previous researchers. We considered Cardott's (1989) data set to be one of the best, because it represented measurements on vitrinite from only one lithological unit (Woodford Shale), and possible noise related to lithological variations was thus reduced. After excluding data from the Oklahoma Platform and mountainous areas to the south, and correcting for the suppression effect described above, we found that a regression on vitrinite data published by Cardott (1989) yielded an estimate for post-Paleozoic erosion in the Anadarko Basin of 2.5 km. Varying the hydrogen index and the suppression correction as described above resulted in estimates that ranged between 2.0 and 3.0 km. These estimates are significantly higher than the estimate made by Cardott (1989) (630 m) which was based on uncorrected data from a wider area. In revisiting Schmoker's (1986) data, we excluded data from the Oklahoma Platform and corrected for vitrinite-reflectance suppression in the Woodford Shale. Our interpretation of Schmoker's (1986) data yields a "best" erosion estimate of 2.0 km, while a consideration of uncertainties leads to estimates in the range of 1.8 to 2.2 km. These estimates are substantially higher than Schmoker's (1986) original estimate of 790 m. A similar reinterpretation of Pawlewicz's (1989) vitrinite reflectance data changed our erosion estimate from his value of 2.5 km to 2.4 km.

Analysis of apatite fission-tracks (AFT) is a quantitative method for estimating erosion thicknesses that also yields an estimate of the timing of erosional events. Carter et al. (1998) interpreted AFT data from both the Wichita Mountains and eastern deep parts of the Anadarko Basin as representing total erosion of ~1.5 km that started 40~50 Ma. AFT data from uplifts along the Ouachita trend indicate total erosion of ~1.0 km occurred in the epeirogenically uplifted southern midcontinent during the last ~40 Ma (Corrigan et al., 1998).

Although there is a wide variance in the estimates of post-Paleozoic erosion made by different methods, all of the estimates fall in the range of 1.0 to 3.0 km. AFT data indicate erosional cooling started about 40 to 50 Ma. It thus appears likely that maximum sediment thicknesses in the Anadarko Basin were reached in late Cretaceous or Tertiary time and have been followed by a few kilometers of subsequent erosion.

8. Thermal History

During Cambrian-Ordovician (~525-435 Ma) heat flow in the present day Anadarko Basin was likely high due to rifting (Feinstein, 1981). After Cambrian igneous activity ceased, thermal subsidence followed in the aulacogen trough (~500-300 Ma). There has been no metamorphic, igneous activity in the Anadarko Basin for approximately the last 500 Ma (Johnson, 1989). Garner and Turcotte (1984) proposed that heat flow in the Anadarko Basin was high in late Paleozoic time (325-305 Ma) due to lithospheric thinning caused by extension. However, there doesn't appear to be geological evidence of significant extensional deformation in the Anadarko Basin or Wichita Mountains during the late Paleozoic (Ye et al., 1996). The widely accepted tectonic model for the Anadarko Basin during late Paleozoic time is a foreland basin model, with compressional tectonics and basement-involved thrust faulting (Brewer et al., 1983; Gilbert, 1983; Goldstein, 1984; Perry, 1989; Gilbert, 1992; Ye et al., 1996).

Different types of paleothermal indicators provide different types of information. Analysis of apatite fission-track (AFT) data provides estimates of the timing and magnitude of cooling. However, cooling can be due to erosion or reductions in heat flow, and AFT data alone are insufficient to distinguish between the two. If vitrinite reflectance data are available over a significant depth interval, the rate at which reflectance increases with depth can be used to estimate the paleothermal gradient. However, it may be difficult to relate the present day maturation state of vitrinite to a single, precise time period in the past.

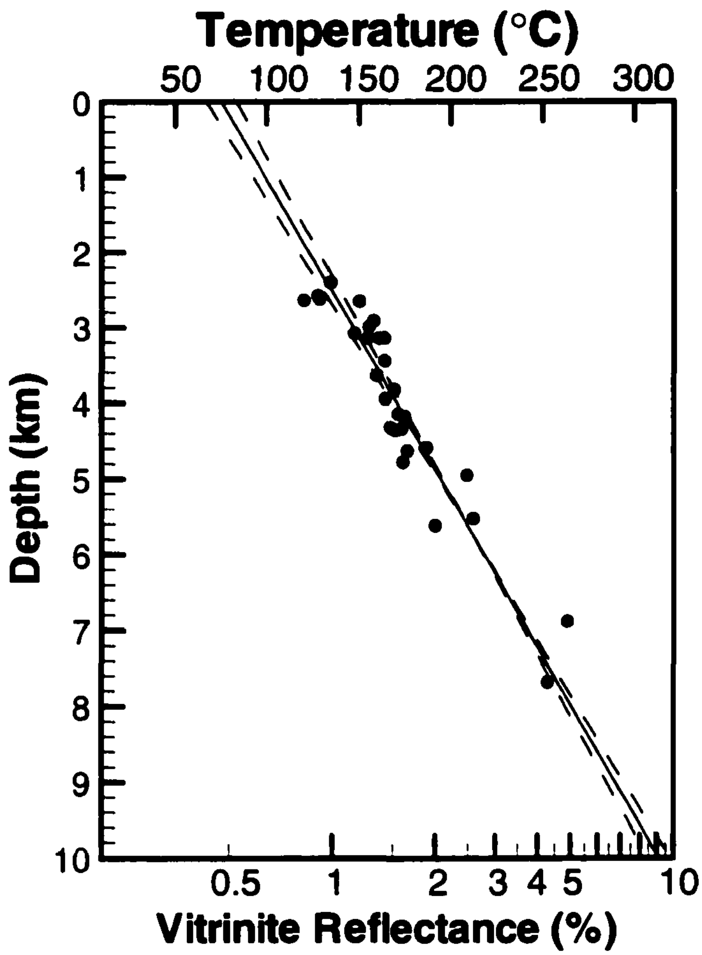
Paleothermal gradients can be estimated from an empirical relationship between maximum temperature (T_{max}) and vitrinite reflectance (R_o) as (Barker and Pawlewicz, 1994)

$$T_{max} = \frac{\ln R_o + 1.68}{0.0124} \qquad (1)$$

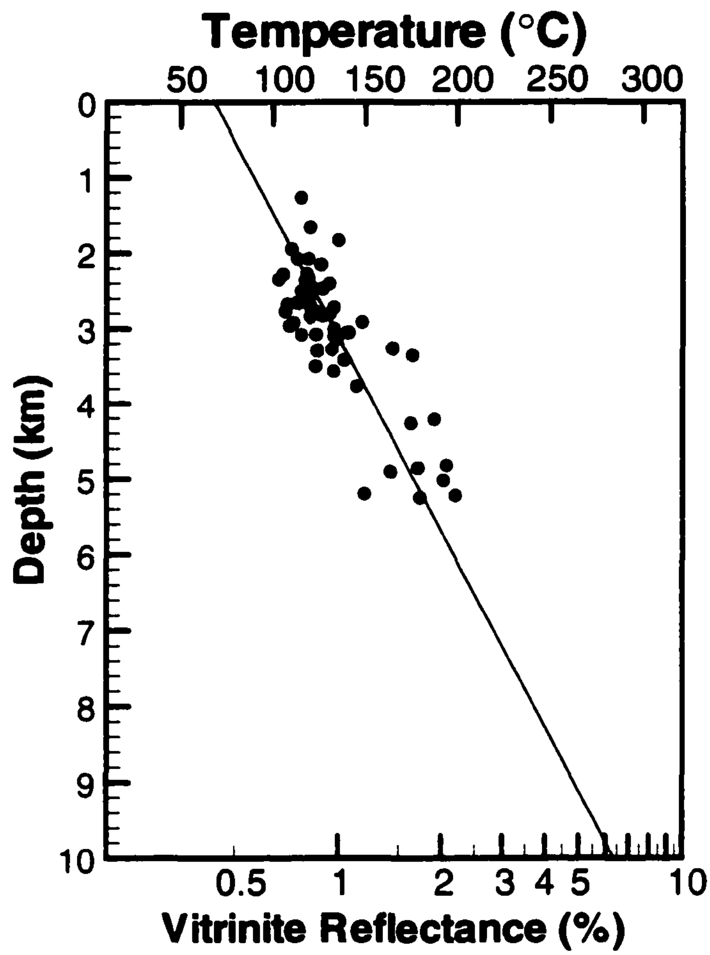
where R_o is vitrinite reflectance in % and T_{max} is maximum temperature in degree Celsius. We are aware of more sophisticated methods of relating vitrinite reflectance to temperature which incorporate time (e.g., Sweeney and Burnham, 1990). However, we have poor constraints on timing, and believe any possible refinement in methodology would be small in comparison to likely errors from numerous sources. We therefore consider a simple method appropriate to the task at hand.

Applying equation (1) above, we estimated paleothermal gradients in the Anadarko Basin by analyzing a suite of vitrinite reflectance data collected by various researchers (Schmoker, 1986; Cardott, 1989; Pawlewicz, 1989) (Figure 4). We winnowed these data to remove data which were geographically close to, but not actually in, the Anadarko Basin. We also made corrections for suppression as described above. In the case of Pawlewicz's (1989) data, we threw out two Permian-age samples, as these relatively shallow samples appeared to be anomalously mature and may have been reworked.

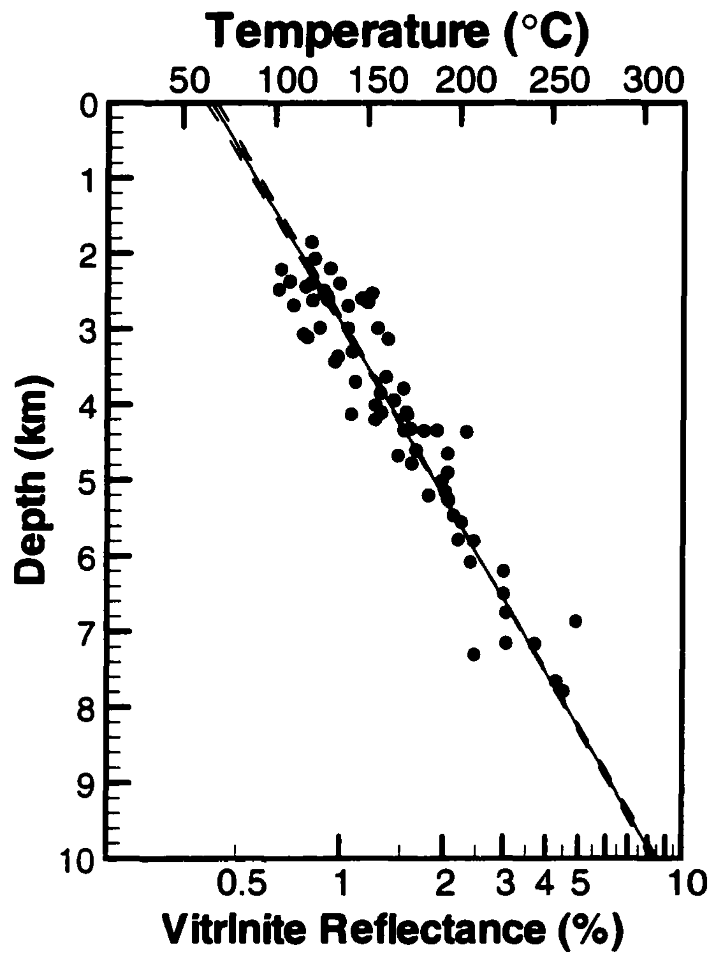
A reinterpretation of vitrinite reflectance data published by Schmoker (1986), Cardott (1989), and Pawlewicz (1989), resulted in estimated paleothermal gradients of 24 (±1) C/km, 24 (±2) °C/km, and 22 °C/km, respectively, where the range of uncertainty reflects uncertainty in the suppression correction applied to Woodford


Temperature (°C)
50 100 150 200 250 300
Depth (km)
0 1 2 3 4 5 6 7 8 9 10
0.5 1 2 3 4 5 10
Vitrinite Reflectance (%)

(a)


Temperature (°C)
50 100 150 200 250 300
Depth (km)
0 1 2 3 4 5 6 7 8 9 10
0.5 1 2 3 4 5 10
Vitrinite Reflectance (%)

(b)


Temperature (°C)
50 100 150 200 250 300
0 1 2 3 4 5 6 7 8 9 10
Depth (km)
0.5 1 2 3 4 5 10
Vitrinite Reflectance (%)

(C)

Figure 4. Semilogarithmic plots of vitrinite reflectance and equivalent maximum temperature versus depth. Maximum temperatures were calculated using equation (1) in the text. Filled circles represent vitrinite reflectance data. Solid lines are best-fit exponential regressions; dashed lines indicate range of uncertainty introduced by corrections for suppression of vitrinite reflectance in the Woodford Shale. (a) Cardott's vitrinite reflectance measurements (Cardott, 1989) on the Devonian-age Woodford Shale from the Anadarko Basin. Regressions indicate ~2.5 (±0.5) km of erosion and a paleothermal gradient of 24 (±2) °C/km. (b) Pawlewicz's vitrinite reflectance measurements (Pawlewicz, 1989) on Ordovician, Silurian, and Pennsylvanian age rocks from the Anadarko Basin. Regressions indicate 2.4 km of erosion and a paleothermal gradient of 22 °C/km. (c) Vitrinite reflectance data (collected by Schmoker (1986)) from the Anadarko Basin. Regressions indicate 2.0 (±0.2) km of erosion and a paleothermal gradient of 24 (±1) °C/km.

Shale samples (Figure 4). The effect of vitrinite reflectance suppression in the Woodford Shale is significant for some data sets. For example, if Cardott's (1989) data are not corrected, they yield an estimated paleothermal gradient of 35 °C/km, because they consist purely of Woodford Shale samples. The effect of suppression is to make the vitrinite reflectance of shallow samples anomalously low. This increases the paleothermal gradient estimated from uncorrected data and decreases the estimated amount of erosion. However the interpretation of Schmoker's (1986) data, which consists only partly of Woodford Shale samples, is not significantly affected by a suppression correction. The overall effect of correcting for suppression of vitrinite reflectance in the Woodford Shale is to produce estimates of the paleothermal state which are more consistent with vitrinite obtained from other rocks, as well as other paleothermal indicators and the present day thermal state.

In consideration of the inherent uncertainties involved in estimating paleothermal gradients from vitrinite reflectance data (e.g., see discussion by Feazel and Aram, 1990), the three estimates we obtained from different vitrinite data sets from different locations are reasonably consistent. As vitrinite reflectance is believed to have an exponential dependence upon temperature (Sweeney and Burnham, 1990), the reflectance of vitrinite recovered from southwestern Oklahoma was probably determined at the time of maximum burial. If we assume that remnants of Triassic, Jurassic, Cretaceous, and Tertiary sedimentary strata found today in the western parts of Anadarko Basin and the Oklahoma Panhandle (Johnson et al., 1972; Johnson, 1989) indicate continuous Mesozoic sedimentation, then maximum burial and maximum temperatures were probably reached between the end of the Cretaceous Period (65 Ma) and early Tertiary time (40-50 Ma) when AFT data (Carter et al., 1998) indicate uplift and cooling began.

The paleothermal gradient inferred from the vitrinite reflectance (~22-25 °C/km) is slightly higher than a present geothermal gradient of 21 °C/km, and appears to be reasonably consistent with geological evidence that the Anadarko Basin has not experienced heating events since the Late Paleozoic (250 Ma) or before. The youngest volcanic rocks and igneous intrusions are middle Cambrian age (~525-550 Ma) (McConnell and Gilbert, 1990). As the half-lives of radioactive heat-generating isotopes in the crust are of the order of 10^9 years, it is not possible for significant changes in crustal heat-production to have occurred within the last few hundred million years. We therefore believe it is reasonable to conclude that the thermal state of the Anadarko Basin has probably not changed significantly since late Paleozoic time.

9. Discussion

The heat flow estimates made in this study differ in some respects from those reported by Carter et al. (1998). In the Anadarko Basin (see Figure 1), differences may be readily attributable to spatial variations. However, apparent discrepancies in the Oklahoma Platform are more difficult to explain. Carter et al. (1998) estimated heat flow at two sites in the Oklahoma Platform (Garner and LeForce wells) to be 57 and 64 mW/m^2, while we estimate heat flow at two nearby (25-30 km distant) sites to be 37 and 40 mW/m^2. The magnitude of this discrepancy is troubling; the average heat flow estimated by Carter et al. (1998) at the Garner and LeForce wells is 56% higher than that found in this study.

Differences may be due to the use of different methodologies. We used corrected BHTs to estimate *in situ* temperature, and measured thermal conductivities on drill cuttings. Carter et al. (1998) made continuous, high-precision temperature logs in idle oil and gas wells and made thermal conductivity measurements on core samples. Continuous temperature-measurements are more desirable than BHTs, especially if they are made in boreholes at thermal equilibrium with the wall rock. However, it is questionable if the temperature logs used in Oklahoma Platform heat flow estimates by Carter et al. (1998) were completely at thermal equilibrium. The Garner well had equilibrated for only one month, and the LeForce well had been produced for two months just a few months before logging. Carter et al. (1998) also utilized core samples for thermal conductivity measurements. These are usually considered to be preferable to the drill cuttings we utilized. However, cuttings provide a more complete sampling of the geologic section over a greater depth range. The limited availability of core samples meant that Carter et al. (1998) had to rely upon thermal conductivity measurements made on cores from wells other than the wells in which temperatures were logged. The possibility thus exists that facies changes could have introduced considerable errors. In other cases, measurements were made on outcrops or general basin-wide averages substituted for measurements ("the shale thermal conductivity used for the Mississippian interval and the Woodford Formation was an average value from several locations in the basin", Carter, 1993). On the other hand, the estimation of *in situ* thermal conductivities from measurements on drill cuttings requires an anisotropy correction which is not needed for measurements on cores. Drill cuttings are also subject to sampling biases. For example, the Permian section in the Anadarko Basin contains about 15% evaporites. The thermal conductivity of these rocks was estimated by Gallardo and Blackwell (1999) to be 4.65 W/m-K. If Permian evaporites were dissolved during the drilling process, their absence in our measurements may have biased average thermal conductivities downward. This factor could possibly have contributed to the lower heat flow values reported in this study. However, if present, the magnitude of the effect would be to lower heat flow estimates only about 10% below their true values. This factor is thus insufficient by itself to explain why heat flows estimated by Carter et al. (1998) were 56% higher than the estimates made in this study.

At the present time, we do not know if differences between our results and those of Carter et al. (1998) are due to geographic variations in heat flow between sites, or if they can be attributed to inherent uncertainties in different methodologies.

10. Conclusions

1. The average geothermal gradient in the Anadarko Basin and the Oklahoma Platform estimated from 856 corrected BHTs (bottom-hole temperatures) is 21 °C/km, but varies from 21 °C/km in the south to 23 °C/km in the Oklahoma Platform to the north.

2. Heat flow (±20%) at 9 sites in the Anadarko Basin and Oklahoma Platform estimated from corrected bottom-hole temperatures ranges from 30 mW/m^2 to 42 mW/m^2, with a mean value of 36 mW/m^2. The average heat flow at 7 sites in this area as estimated in previous studies is 51 mW/m^2.

3. Combining new and previously published data, average heat flow in the Anadarko Basin and the Wichita Mountains is 39 mW/m^2, in the Oklahoma Platform the average heat flow is 51 mW/m^2.

4. The north-south variation of heat flow across the Anadarko Basin and the Oklahoma Platform appears to be mainly controlled by differences in basement heat production.

5. There is no evidence that the thermal state of the Anadarko Basin has changed significantly since the late Paleozoic (250 Ma). Estimates of the average paleothermal gradient obtained from analysis of vitrinite reflectance samples range from 22 to 25 °C/km. The dominant thermal change over the last tens of millions of years has been cooling due to erosion.

6. The total amount of erosion in the Anadarko Basin that has occurred since early to middle Cenozoic time appears to be in the range of 1.0 to 3.0 km as estimated from analysis of apatite fission-trick and vitrinite reflectance data.

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CHAPTER 4

Overpressures in the Anadarko Basin, Southwestern Oklahoma: Static or Dynamic ?

1. Abstract

Fluid pressure and temperature data were collected along a north-south linear trend through the Anadarko Basin in southwest Oklahoma. Subsurface temperature was estimated from corrected bottom-hole temperature data; the average thermal gradient was 21 °C/km. Fluid pressures were estimated from mud weights and well-head shut-in pressures. Fluid pressures above hydrostatic start at a depth of about 2.5 km. On a normalized scale where hydrostatic pressure = 0 and lithostatic pressure = 1, estimates of fluid pressures were found to range from 0.4 to 0.8 over a depth range of 4 to 7 km.

Evaluation of stratigraphic data and paleothermal indicators constrains the Anadarko Basin to have undergone from 1 to 3 km of Mesozoic and/or Cenozoic sedimentation followed by subsequent erosion which started about 50 Ma. Analysis of vitrinite reflectance and apatite fission-track data suggests that the average thermal gradient in the Anadarko Basin has been in the range of 21 to 25 °C/km over the last 50 Ma. Geologic data indicate unambiguously that the Anadarko Basin has cooled over the last tens of millions of years.

Using scale analyses and a simple numerical model, we evaluated two end-member hypotheses (compaction disequilibrium and hydrocarbon generation) as possible causes of overpressuring in the Anadarko basin. If compaction disequilibrium is the primary cause of present-day overpressures in the Anadarko Basin, the Anadarko Basin is required to have an average permeability of 10^{-23} m^2 or lower. If geopressures in the Anadarko Basin result from hydrocarbon generation, average basin permeabilities can be as high as 10^{-20} m^2. Scaling these average permeabilities down to thicknesses of 100 m or less implies permeabilities lower than 10^{-24} m^2 are required in the most optimistic scenario. The lowest permeabilities ever measured on sedimentary rocks are in the neighborhood of 10^{-22} to 10^{-23} m^2. Thus it

is not possible to reconcile the existence of overpressures in the Anadarko Basin with classic hydrodynamic theories which maintain that aquicludes do not exist. Either some unknown process is generating excess fluid pressures, or the Anadarko Basin contains pressure seals.

Geopressure models which invoke compaction disequilibrium and are commonly applied to Cenozoic basins undergoing rapid sedimentation (e.g., Gulf Coast Basin of southeastern U.S.) do not appear to apply to the Anadarko Basin. The Anadarko Basin belongs to a group of cratonic basins which are tectonically quiescent and are characterized by the association of abnormal pressures with natural gas.

2. Introduction

Fluid pressures above hydrostatic are common in sedimentary basins throughout the world (Fertl, 1976; Hunt, 1990; Bigelow, 1994; Osborne and Swarbrick, 1997), and are commonly referred to as "geopressures" or "overpressures". Abnormally high fluid pressures in sedimentary basins can potentially be caused by a number of different geologic mechanisms. These include the rapid accumulation of low-permeability sediments, hydrocarbon generation, diagenic dehydration, and aquathermal pressuring (Bethke, 1986; Shi and Wang, 1986; Spencer, 1987; Luo and Vasseur, 1992; Bredehoeft et al., 1994; Audet, 1995; Gordon and Fleming, 1998). The degree to which overpressuring develops depends upon competition between the geologic pressuring-mechanism and pressure dissipation as determined by the hydraulic diffusivity of rocks within a basin (Bredehoeft and Hanshaw, 1968; Neuzil, 1995; Osborne and Swarbrick, 1997).

As discussed by Bredehoeft et al. (1994), there are two distinct schools of thought on the creation and maintenance of anomalous fluid pressures in the Earth. The static school (Bradley, 1975; Bradley and Powley, 1994) believes that abnormal pressures are maintained by pressure seals (Hunt, 1990; Ortoleva, 1994). A pressure seal was defined by Hunt (1990, p.2) as

> "...a zone of rocks capable of hydraulic sealing, that is, preventing the flow of oil, gas, and water. The term does not refer to capillary seals...the term refers to seals that prevent essentially all pore fluid movement over substantial intervals of geologic time".

The above quote clearly shows that the original intention was to apply the term to rock units that essentially behave as if they have zero permeability over substantial periods of geologic time (~10^7-10^8 years). However, in the literature in recent years one more commonly finds the term "pressure seal" loosely defined to be any low-permeability unit which is capable of confining abnormal fluid pressures over any time period. This is better defined as an aquitard. In this paper, we prefer to follow Hunt's (1990) original intention and reserve the term "pressure seal" to designate a layer which has zero or near-zero permeability.

In contrast to the static paradigm, the dynamic school embraces the classical hydrogeologic tenet that "there are no totally impermeable geologic materials" (Bredehoeft et al., 1982, p. 297). Although exceptions can be made for unusual materials such as permafrost and salt, the dynamic view is that pressure seals do not exist and are certainly not common in sedimentary basins. In the dynamic paradigm, abnormal formation pressures arise from transient disturbances related to an ongoing geologic processes such as rapid sedimentation (Bethke, 1986), time-dependent compaction (Dewers and Ortoleva, 1994), clay dehydration (Audet, 1995; Gordon and Fleming, 1998), or hydrocarbon generation (Spencer, 1987; Bredehoeft et al., 1994). The existence of abnormal fluid pressures implies as a corollary the parallel existence of a geologic process which is actively creating a pressure gradient.

The existence of overpressures in young, rapidly subsiding, basins such as the Gulf Coast Basin of the southeast U.S. or the South Caspian Basin of the former USSR (Bredehoeft et al., 1988), presents no problems for the dynamic school. Overpressures in these settings are readily explainable as transient compaction disequilibrium related to high sedimentation rates and a predominance of low-permeability shales (Bredehoeft and Hanshaw, 1968; Sharp and Domenico, 1976; Bethke, 1986; Mello et al., 1994). For example, Bethke (1986, p. 6538) showed that anomalous formation pressures in the Gulf Coast basin could be maintained at sedimentation rates of 100 - 10,000 m/Ma if average shale permeabilities were in the range of 10^{-18} - 10^{-20} m^2. These values are consistent with laboratory measurements of shale permeabilities from the Gulf Coast (Neglia, 1979) as well as estimated Plio-Pleistocene sedimentation rates.

What is not well understood at the present time is how overpressures can exist in older basins that have not undergone active subsidence for tens or hundreds of millions of years. The Anadarko Basin in southwestern Oklahoma is known to have areas of extensive overpressuring (Figure 1) (Hunt, 1990; Al-Shaieb et al., 1992; Al-Shaieb et al., 1994a, 1994b), yet the Anadarko Basin has not experienced significant sedimentation for more than 200 million years, and has undergone uplift and erosion for about the last 40 to 50 million years (Gilbert, 1992; Carter et al., 1998: Corrigan et al, 1998; Lee and Deming, 1999).

It is in areas such as the Anadarko Basin that the divergence of static and dynamic school hypotheses indicates how little we understand concerning the origin and maintenance of abnormal pressures in older sedimentary basins. The fundamental question is this: what is the cause of the overpressuring, and how is it maintained? In this paper, we approach this problem from the perspective of testing two end-member hypotheses. We first consider as a static hypothesis what average permeability will allow overpressuring in the Anadarko to be explained as a remnant of Paleozoic

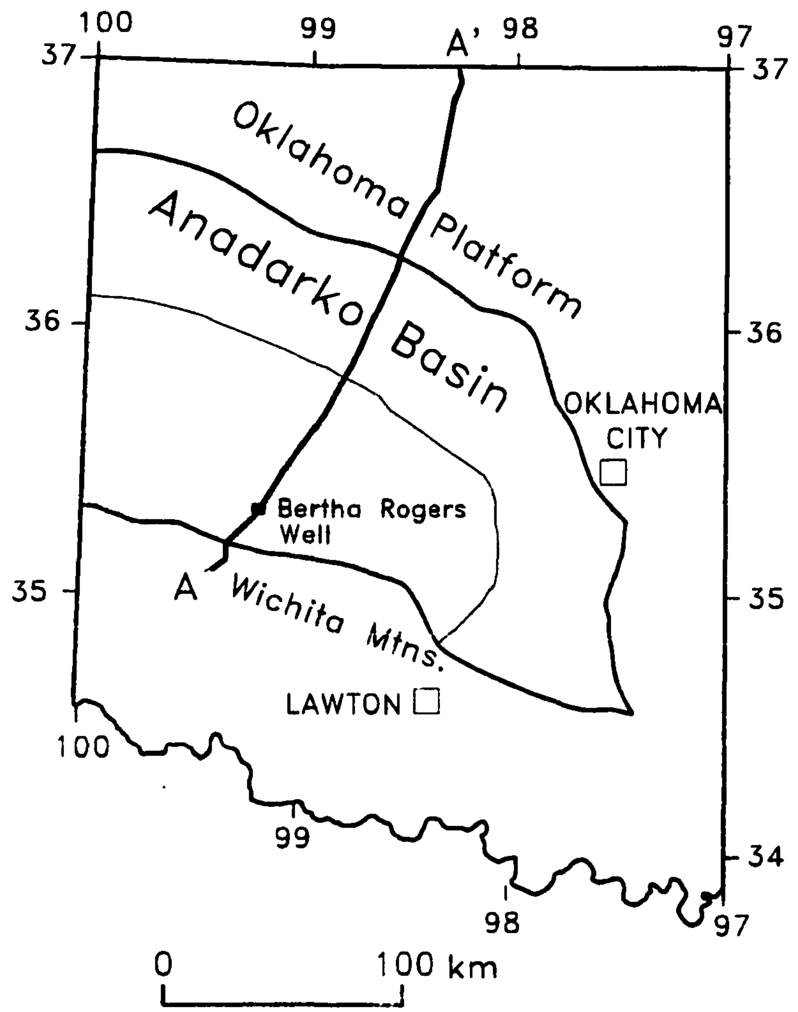


Figure 1. Location of the Anadarko Basin and cross-section A-A' in western Oklahoma (after Johnson, 1989). Stippled area represents overpressure zone.

compaction disequilibrium. Alternatively, we then consider under what circumstances hydrocarbon generation can explain overpressuring.

3. Geologic Setting

The Anadarko Basin (Figure 1) bounded by the Wichita-Amarillo uplift on the south, the Nemaha uplift on the East, and the Cimarron arch on the west, is the deepest sedimentary basin located on the North American Craton. Total sediment thickness in the deep Anadarko Basin exceeds 12 km, and consists of sandstone, limestones, and shale of Cambrian to Permian age (Johnson, 1989) (Figure 2). The Anadarko Basin is a prolific producer of both petroleum, and, more importantly, natural gas. Cumulative hydrocarbon production as of 1985 was > 5 million barrels of oil and 2.3 trillion cubic meter of natural gas (Davis and Northcutt, 1989). The Anadarko Basin contains ~50 major fields and hundreds of minor ones; production is from both stratigraphic and structural traps. Identified source rocks in the Anadarko Basin and surrounding areas mainly consist of Ordovician (Simpson Group shales), Devonian-Mississippian (Woodford shale), and Pennsylvanian (thick, dark shales) (Johnson and Cardott, 1992). Summaries of the geologic history, stratigraphy, and tectonic setting of the Anadarko Basin are given by Gilbert (1992) and in the volume edited by Johnson (1989).

4. Fluid Pressure Data

The Anadarko Basin is reported in the geologic literature to have areas of overpressuring (Jorgensen, 1989; Hunt, 1990; Al-Shaieb et al., 1992, 1994a, b). Al-Shaieb et al. (1992, 1994a, b) stated that fluid pressures exceeding hydrostatic generally start at about 2.3 to 3.0 km depth, but return to near-hydrostatic below the Woodford Shale. The three-dimensional overpressured zone was identified by Al-Shaieb et al. (1992, 1994a, b) in map view as oval in shape, extending about 267 km

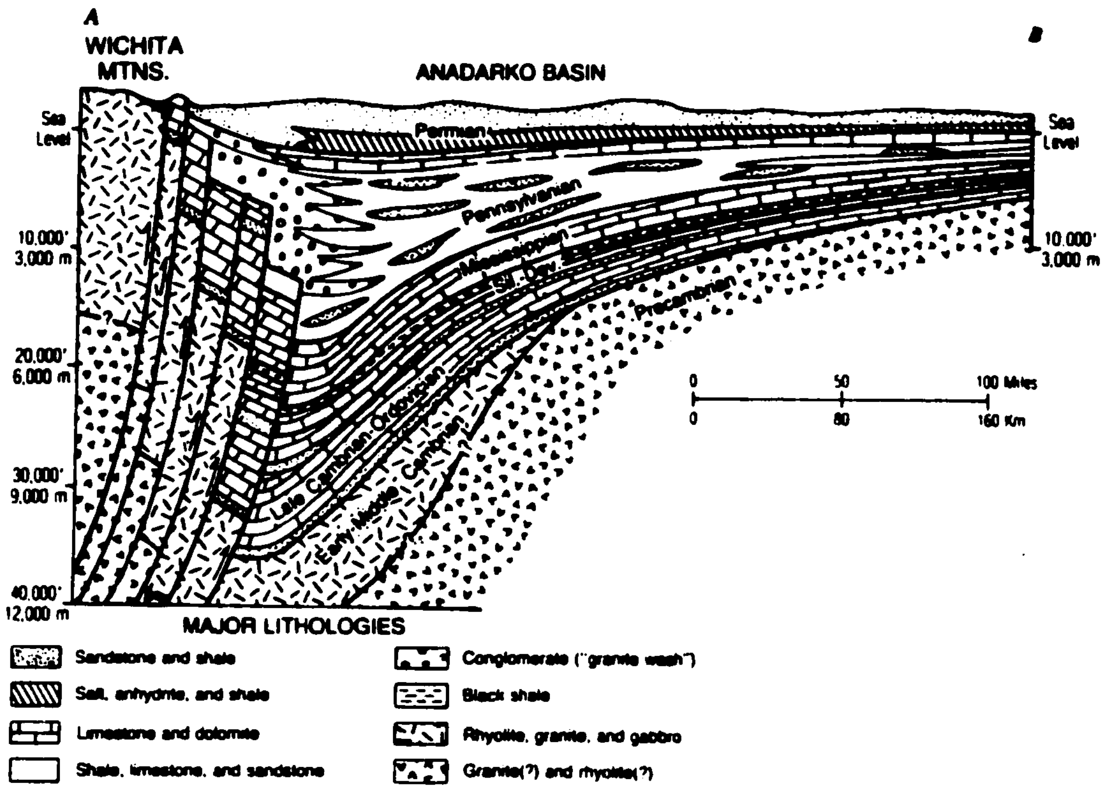


Figure 2. Geological cross-section (A-A') through the Anadarko Basin (from Johnson, 1989).

along a northwest-southeast trend, and about 100 km from southwest to northeast. Within this three-dimensional volume, high fluid-pressures are not ubiquitous. Rather, they are evidently confined to distinct zones termed compartments. Al-Shaieb et al. (1992, 1994a, b) termed the entire volume occupied by pressurized compartments a "megacompartment complex".

To present our data and modeling results, we use a dimensionless normalized pressure (P_n) defined as

$$P_n = \frac{P - P_{hydro}}{P_{lith} - P_{hydro}} \tag{1}$$

where P is fluid pressure, P_{hydro} is hydrostatic pressure (9.8 MPa/km), and P_{lith} is lithostatic pressure (22.5 MPa/km). P_n ranges from 0 for hydrostatic pressure to 1 for lithostatic pressure.

We estimated fluid pressures in the Anadarko Basin from two types of data: drilling mud weights reported on well log headers and well-head shut-in pressures. Due to the size and complexity of the Anadarko Basin, we chose to concentrate our data collection and analysis along a southwest-northeast line perpendicular to the strike of the basin structure (Figures 1, 2).

Mud-weight data (Figure 3a) were collected from the Oklahoma Geologic Survey well log library. The ability to estimate *in situ* pressures from drilling mud weights has limitations. However, the method has been successfully applied in the Gulf Coast Basin by Hanor and Sassen (1990). Mud weights represent an upper limit on formation pressure, as the weight of the mud column in the borehole must be greater than formation pressure to avoid the possibility of a blowout. We also estimated fluid pressures from well-head shut-in pressure (WHSIP) donated by Dwight's Energy Data (Figure 3b). WHSIPs represent minimum estimates of true formation pressures,

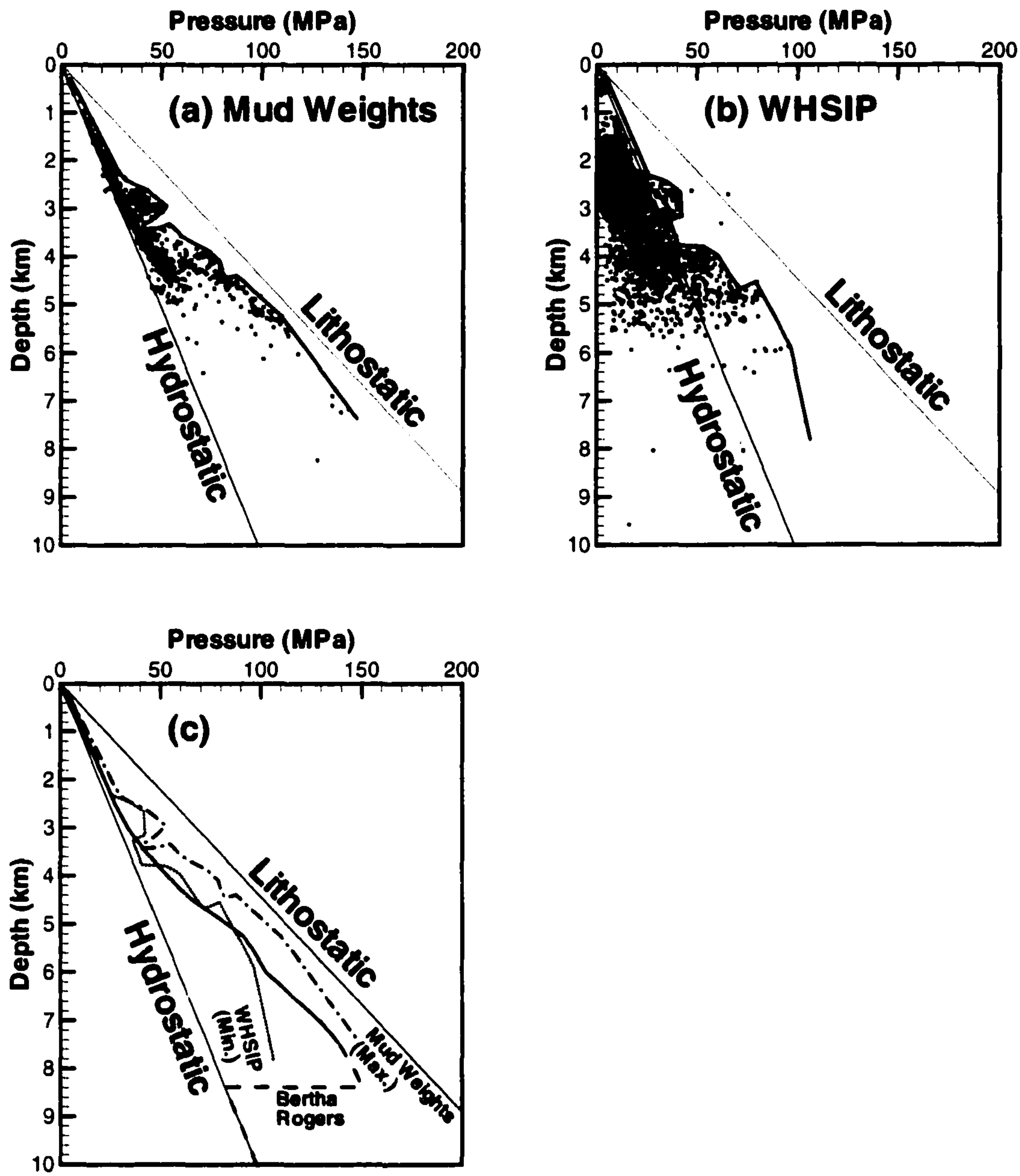


Figure 3. Fluid pressure in the Anadarko Basin estimated from (a) mud weights, and (b) well-head shut-in pressures (WHSIPs). Mud weights generally yield maximum estimates of true fluid pressures, while WHSIPs are minimum estimates of fluid pressures. (c) Fluid pressure trends estimated from mud weights and WHSIPs plotted along with fluid pressure in the Bertha Rogers well estimated from mud weights.

as reservoir pressures may be drawn down by production. Comparing the two data types (Figure 3c) allows us to estimate the limits of probable resolution.

Normalized fluid pressure (P_n) ranges from 0.4 (from WHSIPs) to 0.8 (from Mud weights) for the depth range from 4 to 7 km (Figure 3c). A contour plot (Figure 4) of normalized fluid-pressure (P_n) estimated from mud weights shows that overpressuring is essentially confined to the southern half of the Anadarko Basin. A maximum fluid pressure (P_n) of about 0.6 is reached at depths of about 5 to 6 km. Deeper data are insufficient to allow accurate discernment of the fluid pressure regime. However, Al-Shaieb et al. (1992, 1994a, b) indicate that the Devonian-age Woodford Shale constitutes a bottom seal to the overpressured zone and that underlying Cambrian-Ordovician carbonates contain fluid pressures that are near hydrostatic. Jorgensen (1989) and Jorgensen et al. (1993) described the Cambrian-Ordovician carbonates of the central US as constituting part of the Western Interior Plains aquifer system. However, Jorgensen et al. (1993) also said that in "deep basins" the permeability is "greatly reduced" and cited a permeability range of 10^{-12} to 10^{-18} m^2. Although a minimum permeability of 10^{-18} m^2 would be too low to qualify as an aquifer, it is also too high to effectively retain abnormal formation pressures over geologic time (Deming, 1994). It thus seems likely that the geopressured zone in the Anadarko Basin does not extend into the basal Cambrian-Ordovician carbonates.

5. Mathematical Modeling

5.1 General Approach and Philosophy

Mathematical models should be used adjunctively to understand the physics of geologic processes. Most geologic settings, including sedimentary basins, are far too complex to reproduce accurately in modern digital computers. As the degree of model complexity grows, so does the problem of obtaining meaningful constraints

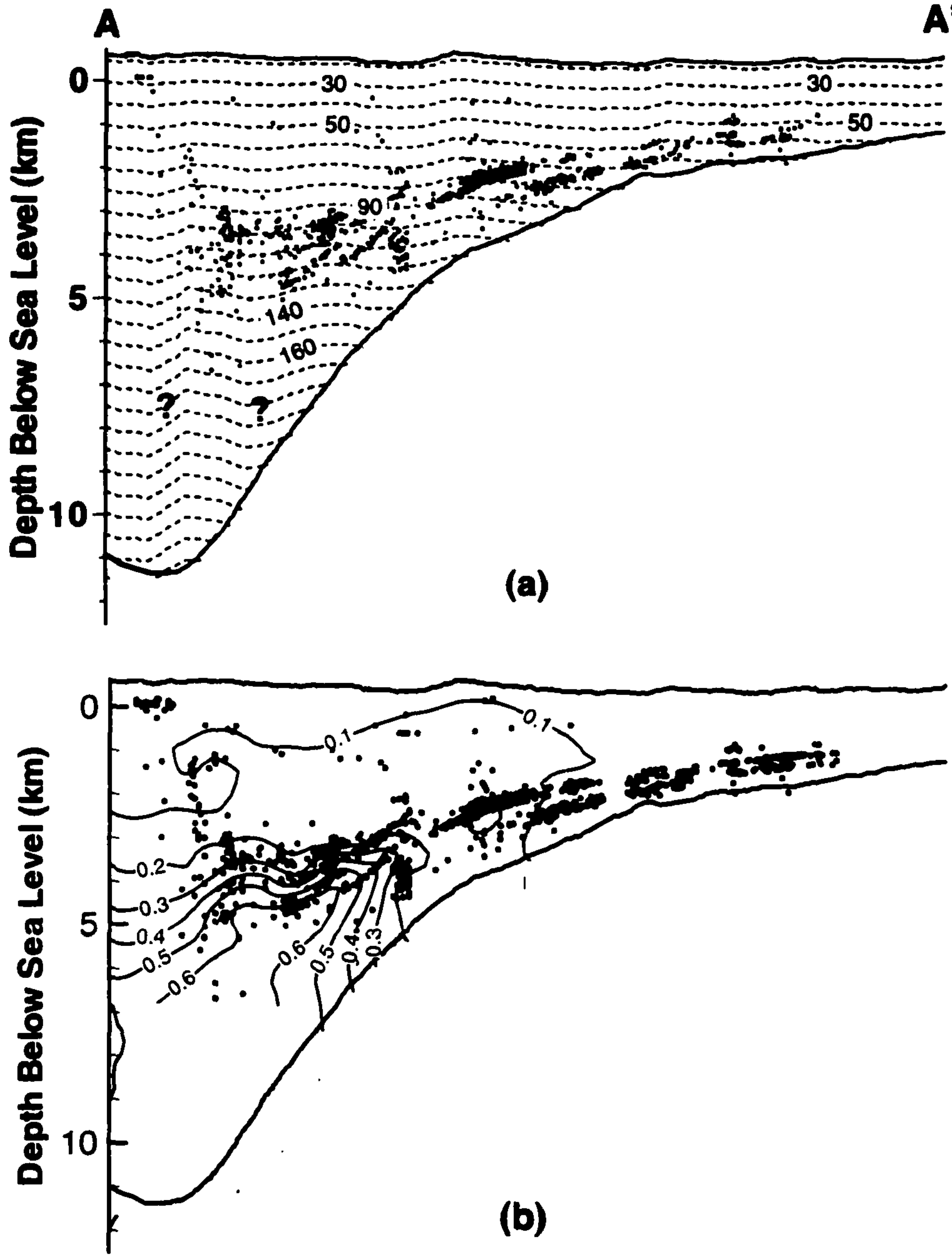


Figure 4. (a) Estimated temperature (degrees Celsius) along the cross-section A-A' through the Anadarko Basin and the Oklahoma Platform. Dashed lines show estimated temperature in degrees Celsius, dots show location of bottom-hole temperature measurements. (b) Fluid pressure estimated along the cross-section A-A' through the Anadarko Basin. Contour lines represent normalized pressure (P_n) estimated from mud weight, dots show location of mud weight measurements.

and results. We therefore favor simple models, as these provide the most insight into the physics of complex geologic problems.

The simplest approach that can be taken with diffusion problems is to conduct a scale analysis which explores the relationship between characteristic times and lengths for different values of a key physical property such as hydraulic diffusivity. For example, Deming (1994) used this approach to demonstrate that in order for geologic units thinner than 100 m to effectively confine anomalous fluid pressures for more than a million years their permeability would have to be lower than 10^{-23} m^2.

In this study, we found it useful to construct a simple numerical model for evaluating alternative hypotheses concerning the generation of overpressures in the Anadarko Basin. Our approach was to identify one variable, permeability, as the largest unknown, and pose the following question : for what value or values of permeability is this hypothesis viable? In this approach, other variables (Table 1) were generally held fixed. The justification for this approach was that permeability was unknown by several orders of magnitude while variables such as fluid viscosity were well known to vary within a predictable range.

In keeping with our philosophy of employing the simplest methods available, we used a one-dimensional model with uniform permeability. However, we included a realistic representation of source rocks in this model, as we had reasonably good constraints on depths, kerogen types, and total organic carbon (see below). In contrast, permeability data were essentially missing.

5.2 Numerical Model

A generalized description of fluid flow in one dimension is

$$\frac{\partial}{\partial z}(K\frac{\partial h}{\partial z}) = S_s\frac{\partial h}{\partial t} + \Gamma \qquad (2)$$

Table 1. Model parameters

Parameter	Value
Specific storage	1×10^{-6} m^{-1}
Permeability	variable
Surface temperature	17.5 °C
Thermal gradient at present	21 °C/km
Thermal gradient at 250 Ma	21-25 °C/km
Water viscosity	2.121×10^{-4} kg $m^{-1}s^{-1}$
Matrix density	2650 kg/m^3
Water density	1000 kg/m^3
Kerogen density	1150 kg/m^3
Oil density	850 kg/m^3
Gas density	0.718 kg/m^3

where K is hydraulic conductivity, h is hydraulic head, S_S is specific storage, t is time and Γ is a geologic forcing term (Neuzil, 1995, p.748). Fluid pressure (P) can be estimated from hydraulic head (h) as

$$P = \rho g(h - z) \tag{3}$$

where ρ is fluid density, g is gravitational acceleration, and z is elevation. In this study, where we dealt with fluid pressures of the order of 10-100 MPa, we assumed that atmospheric pressure on the surface (~0.1 MPa) was negligible.

In general, there are two different methods that can be used for evaluating pressures due to hydrocarbon generation: single and multiphase modeling. Multiphase modeling (e.g., Luo and Vasseur, 1996) is more rigorous than single phase modeling because the hydraulic properties of three different phases (water, oil, and gas) can be separately treated. However, single phase modeling can provide feasible results for the distribution of fluid pressures, even though it can not provide the details of oil migration, hydrocarbon saturation, and capillary pressure (Bredehoeft et al., 1994). Therefore, we used single-phase modeling as the simplest approach which would provide insight into the physics of pressure generation and dissipation in this setting.

Our numerical code was tested against an analytical solution to equation (2) for a homogeneous domain with similar spatial and temporal scales. Carlslaw and Jaeger (1959, p. 131) provide an analytical solution to equation (2) for the case where the forcing function Γ is equal to $A_0\exp(-\beta t)$, where A_0 and β are constants and t is time. In comparison to the analytical solution, the numerical solution provided a maximum relative error of 0.7%.

In nature, when fluid pressures exceed lithostatic values, hydraulic fracturing occurs and overpressures are dissipated. Fracturing can be simulated in a numerical model by increasing local permeability until the local pressure returns to lithostatic or below (e.g., Bredehoeft et al., 1994; Mello and Karner, 1996). An alternative method of dealing with this problem is to simply create a local negative source term. In other words, fluid pressures which exceed lithostatic are numerically set back to lithostatic. This procedure mimics the hydraulic fracturing and pressure dissipation which must occur in nature. In this study, we chose the latter approach as the simplest. Most of our critical simulations dealt with studying the circumstances under which the present day observed maximum fluid pressure (P_n = ~0.6) could be produced, and therefore did not result in lithostatic pressures.

6. Hydrocarbon Generation

Overpressures in sedimentary basins can potentially be generated from a number of difference geologic processes. These include aquathermal pressuring, active flow, diagentic alteration, tectonic stresses, and clay dehydration (Bethke, 1986; Shi and Wang, 1986; Spencer, 1987; Luo and Vasseur, 1992; Bredehoeft et al., 1994; Audet, 1995; Gordon and Fleming, 1998). Recent reviews by Osborne and Swarbrick (1997) and Neuzil (1995) indicate, however, that hydrocarbon generation may be one of the most potentially effective mechanisms for generating high fluid pressures in sedimentary basins. Due to the large change in volume, the conversion of solid kerogen or liquid oil to gas is an especially potent mechanism (Barker, 1990). The Anadarko Basin is a prolific producer of natural gas. It therefore follows that gas generation may be a viable hypothesis for explaining the origin of high fluid pressures in this setting.

Our treatment assumes that oil and gas in the Anadarko Basin were generated from the thermal maturation of organic-rich source rocks. The biogenic theory of the origin of oil and gas is well accepted, although abiogenic theories are resurgent (Gold, 1993; 1999).

Modeling of the hydrocarbon generation process must not only take into account the increase in volume that takes place due to gas and/or oil generation, but also the porosity increase due to kerogen/oil consumption (Bredehoeft et al., 1994; Luo and Vasseur, 1996). Oil or gas generation from can be estimated from a first order kinetic model,

$$\frac{dV_k}{dt} = -kV_k \tag{4}$$

where V_k is volume of kerogen per volume of rock. The kinetic rate constant for oil or gas generation from kerogen (k), is given by Arrhenius equation as

$$k = A \exp(-E/RT) \tag{5}$$

where A is the Arrhenius constant (Table 2), E is the activation energy (Table 2), R (8.314 J/mole-K) is the universal gas constant, and T is temperature in K. There are no experimental constraints on appropriate values for either the Arrhenius constant (A) or activation energy (E) for Anadarko Basin source rocks. Therefore, we used generic values (Table 2) for kerogen types II and III as given by Quigley et al. (1987). We assumed that kerogen type II is converted to oil even though a part of kerogen type II is converted to gas (Mackenzie and Quigley, 1988), while kerogen type III is converted to gas. For oil or gas generation from kerogen, or porosity increase due to kerogen consumption, the geologic forcing term of equation (2) was derived by the conservation of mass as

$$\Gamma = -(\frac{\rho_k}{\rho_p} - 1)kV_k \tag{6}$$

Table 2. Kinetic parameters for hydrocarbon generations [Quigley et al., 1987]

	A (1/sec)	E (kJ/mol)	σ^1 (kJ/mol)
Kerogen type II	1.58×10^{13}	208	5
Kerogen type III	1.83×10^{18}	279	13

[1] Standard deviation

where ρ_k is kerogen density (1150 kg/m^3), ρ_p is petroleum density (oil density = 850 kg/m^3; gas density = 0.718 kg/m^3), V_k is volume of kerogen per volume of rock that can be estimated from kerogen density, rock density, and TOC, and k is the rate constant for petroleum generation (oil or gas) from kerogen. V_k can be estimated from equation (5) and (6) as

$$V_k = V_k^o \exp(-kt) \tag{7}$$

where V_k^o is the initial volume of kerogen per volume of rock and t is time.

7. Model Constraints

7.1 Burial History.

The history of Paleozoic sedimentation in the Anadarko Basin (Figure 5) is relatively well known, because most of the sediments deposited during this era are retained today (Johnson, 1989; Gilbert, 1992). However, the post-Paleozoic burial history of the Anadarko Basin is more controversial and less well constrained because most Mesozoic and Cenozoic strata have been eroded.

There is geologic evidence for sedimentary deposition in the Anadarko Basin during the Mesozoic. Extension of the Cretaceous seaway across the Anadarko Basin during the last great inundation of the western interior of the United States implies that sedimentary strata were deposited in a marine environment during Cretaceous time (Garner and Turcotte, 1984; Johnson, 1989). Remnants of Triassic, Jurassic, Cretaceous, and Tertiary sedimentary strata (~200-500 m) are found today in the western parts of Anadarko Basin and the Oklahoma Panhandle (Johnson et al., 1992; Johnson, 1989). The Denver and Raton Basin, which are located between the Rocky Mountains and the Anadarko Basin, probably had the same Cretaceous sediment provenance as the Anadarko Basin, and these basins contain preserved Cretaceous

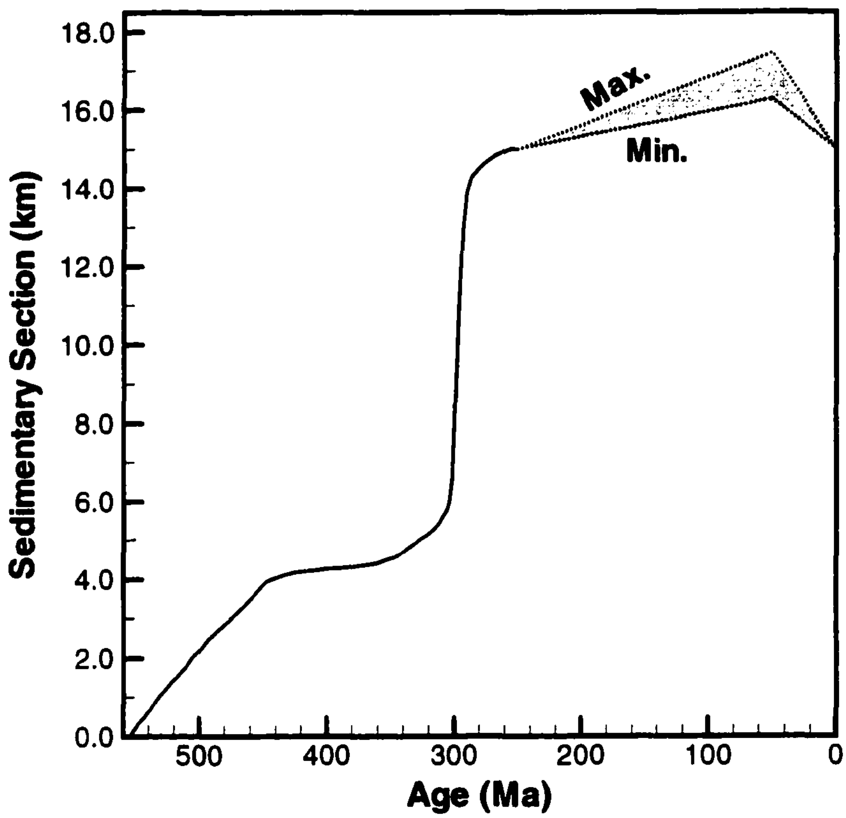


Figure 5. Burial history of the Anadarko Basin (after Gilbert, 1992). Maximum and minimum burial histories from 250 to 0 Ma were estimated from stratigraphic relationships and paleothermal indicators as discussed in the text.

sedimentary strata with thicknesses that range from 560 to 2890 m (Baars et al., 1988).

The primary tectonic activity in the vicinity of the Anadarko Basin since late Cretaceous time has been erosion. The amount and timing of erosion in the Anadarko Basin and surrounding areas is not known with precision. Neither are regional variations in erosion amounts or timing well known. However, analysis of apatite fission-track (AFT) and vitrinite reflectance (R_o) data allow us to place some approximate constraints on the total amount and timing of Cenozoic erosion.

AFT data can be used to estimate both the amount and timing of erosion. Carter et al. (1998) interpreted AFT data from both the Wichita Mountains and eastern deep parts of the Anadarko Basin as representing total erosion of ~1.5 km that started about 40~50 Ma. Corrigan et al. (1998) interpreted AFT data from uplifts along the Ouachita trend as indicating total erosion of ~1.0 km in an epeirogenically-uplifted southern midcontinent during the last ~40 Ma (Corrigan et al., 1998).

Vitrinite reflectance (R_o) data can be used to estimate total erosion by extrapolating semi-log plots of R_o vs. depth to presumed surface values of $R_o = 0.2$ (Figure 6b). Three previous researchers (Schmoker, 1986; Cardott, 1989; Pawlewicz, 1989) used vitrinite reflectance data to estimate the total amount of post-Paleozoic erosion in southwest Oklahoma to be in the range of 630 m to 2.5 km. Lee and Deming (1999) reviewed vitrinite reflectance data collected by previous workers, and concluded that varying erosion estimates could have been affected by regional variations and suppression of vitrinite reflectance in the Woodford Shale (Buchardt and Lewan, 1990). After excluding data from outside the Anadarko Basin (e.g., the Ouchita Mountains and the Oklahoma Platform) and correcting for suppression of reflectance in the Woodford Shale, Lee and Deming (1999) interpreted vitrinite reflectance data from the Anadarko Basin as indicating from 1.8 to 2.5 km of total Cenozoic erosion (Figure 6b).

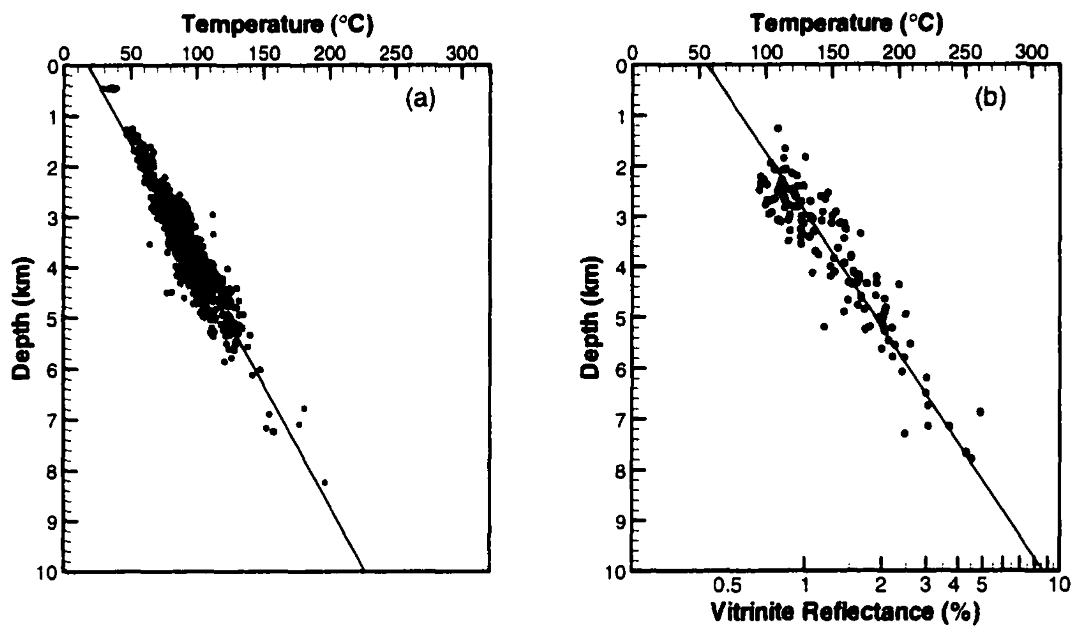


Figure 6. (a) Corrected bottom-hole temperature data from wells lying along cross-section A-A'. Average thermal gradient is 21 °C/km. (b) Vitrinite reflectance data from the Anadarko Basin and equivalent maximum temperature. Average paleothermal gradient is 24 °C/km.

Although there is some variance in the estimates of post-Paleozoic erosion made by different methods, all of the estimates fall in the range of 1.0 to 3.0 km (Figure 5). AFT data indicate erosional cooling started about 40 to 50 Ma. We thus conclude that existing data allow us to reasonably conclude that maximum sediment thicknesses in the Anadarko Basin were reached in late Cretaceous or Tertiary time, and have been followed by 1 to 3 kilometers of subsequent erosion.

7.2 Present-day Temperature

We estimated subsurface temperature in the Anadarko Basin from 856 corrected bottom-hole temperatures (BHTs) collected from log headers of 589 oil and gas wells located along cross section A-A' (Figure 1) through the Anadarko Basin and Oklahoma Platform (Figure 6a). The temperature data were corrected using a Horner plot and average depth-time correction (Deming, 1989). The methodology we used is identical to that employed by Lee et al. (1996) and discussed in several other papers (Deming, 1989; Deming et al., 1990; Lee et al., 1994; Lee and Deming, 1999). We believe our estimates of subsurface temperature (Figure 4b) have an average accuracy of ±5 °C (see discussion in Lee et al., 1996). There are a few direct temperature measurements in the Anadarko Basin (e.g., Carter et al., 1998). However, the oil and gas wells used for temperature measurements by Carter et al. (1998) are not close enough to our cross section A-A' (Figure 1) to allow meaningful comparisons.

Based on corrected BHT data, the average geothermal gradient in the deep, southern part of the Anadarko Basin in the south is 21 °C/km, while the average geothermal gradient in the northern part of the Anadarko Basin and the Oklahoma Platform is 23 °C/km (Figure 6a). There is little regional variation in the thermal state from south to north (see discussion by Lee and Deming, 1999) (Figure 4a). A simple least-squares regression on all temperature data yielded an average thermal gradient of 21 °C/km.

7.3 Thermal History

If vitrinite reflectance data are available over a significant depth interval, paleothermal gradients can be estimated from an empirical relationship between maximum temperature and vitrinite reflectance as (Barker and Pawlewicz, 1994) as

$$T = \frac{\ln R_o + 1.68}{0.0124} \tag{8}$$

Using vitrinite reflectance data previously reported by other researchers (Schmoker, 1986; Cardott, 1989; Pawlewicz, 1989), Lee and Deming (1999) corrected for reported suppression of reflectance in the Woodford Shale (Buchardt and Lewan, 1990), and also excluded data from outlying areas. After corrections, existing data yielded estimates of the paleothermal gradient that ranged from 22 to 25 °C/km. When all data are lumped together, a regression yields an estimated paleothermal gradient of 24 °C/km (Figure 6b).

As vitrinite reflectance is believed to have an exponential dependence upon temperature (Sweeney and Burnham. 1990), the reflectance of vitrinite recovered from the Anadarko Basin was probably determined at the time of maximum burial. If we assume that remnants of Triassic, Jurassic, Cretaceous, and Tertiary sedimentary strata found today in the western parts of Anadarko Basin and the Oklahoma Panhandle (Johnson and Cardott, 1992; Johnson, 1989) indicate more-or-less continuous Mesozoic sedimentation, maximum burial was probably reached between the end of the Cretaceous Period (65 Ma) and early Tertiary time (40-50 Ma) when AFT data (Carter et al., 1998) indicate uplift and cooling began. In the absence of any evidence that the Anadarko Basin has experienced significant tectonothermal events for the last several hundred million years (e.g., the youngest volcanic rocks are more than 500 million years old, McConnell and Gilbert, 1990), we concluded that the paleothermal state recorded by vitrinite reflectance data was probably established at the time of maximum burial, about 40 to 50 Ma.

We recognize that there are limitations inherent in the analysis of vitrinite reflectance data. Vitrinite can be reworked, reflectance may depend upon lithology, and determinations on the same samples from different laboratories may vary by as much as 0.4%. Oxidation can affect vitrinite reflectance, resulting in estimates that are too high by 0.1 to 0.3 %. Samples obtained from drill chips may not represent the actual depth of record, but may be contaminated by vitrinite introduced by borehole caving at more shallow levels (Heroux et al., 1979; Barker and Pawlewicz, 1986; Robert, 1988; Feazel and Aram, 1990). These limitations and possible sources of error are all serious, and should be taken into account in the interpretation of vitrinite reflectance data. However, there is no doubt that a strong causal relationship exits between vitrinite reflectance and temperature. Despite all of the problems, vitrinite is one of the best paleothermal indicators available. Lacking better constraints, we used the best indicator available to us.

We also recognize that more sophisticated methods of relating vitrinite reflectance to time and temperature exist (e.g., Sweeney and Burnham, 1990). However, in the absence of better constraints on timing, we chose to utilize a simple method which recognizes the strong relationship between R_0 and maximum temperature. Under different circumstances (e.g., forward modeling) it might have been appropriate to model the possible influence of time on maturation profiles.

7.4 Geochemical Constraints

Primary source rocks in the Anadarko Basin (Table 3) are Ordovician Simpson-Group shales (Ordovician Age), the Devonian-Mississippian Woodford shale, and certain thick, dark shales of Pennsylvanian age (Johnson and Cardott, 1992). We

Table 3. Source rock data from the Bertha Rogers well (from Kareem, 1992)

	Fm.	Fm. Top (m)	Fm. Bottom (m)	Thickness (m)	Rock type	Kerogen type	TOC (%)[2]
Pennsylvanian	Virgilian	1341	2592	1251	Shale	II, III[1]	1-25
Pennsylvanian	Atoka	3920	5335	1415	Shale(70%) Limestone(30%)	III	1-25
Pennsylvanian	Morrowan	5335	6565	1230	Shale	III	0.5-3.4
Pennsylvanian	Springer	6565	7134	569	Shale	III	0.5-3.4
Devonian	Woodford	8388	8465	77	Shale	II[1], III	1-14
Ordovician	Sylvan	8802	8855	53	Shale	II	1-9
Ordovician	Simpson	9072	9521	449	shale	I, II[1]	1-9

[1] dominant kerogen

[2] from Johnson and Cardott (1992)

estimated dominant kerogen types and total-organic-carbon (TOC) content for these source rocks from published analyses of samples obtained from the Anadarko Basin (Table 3). The depth and thickness of source rocks were constrained by analysis of samples from the Bertha Rogers well (Table 3) in the southcentral part of the Anadarko Basin (Figure 1). The Bertha Rogers well is noteworthy as the deepest petroleum well in the world (9.59 km). Core samples from the Bertha Rogers well have been subjected to intensive geologic and geochemical studies (e.g., Price et al., 1981; Kareem, 1992). In our modeling, we assumed that initial TOC prior to the initiation of hydrocarbon generation was equal to maximum present day values and 50% of initial TOC was inert.

7.5 Permeability

Average reservoir permeabilities in the Anadarko Basin are in the neighborhood of 10^{-13} to 10^{-15} m^2 (Harrison and Routh, 1981; Bebout et al., 1993). In the context of our problem of understanding the origin and evolution of overpressures, these numbers imply little to nothing. In a discussion of midcontinent hydrogeology, Jorgensen et al. (1993) stated that "most oil and gas reservoirs are anomalous subsurface zones". The critical permeabilities are those of the non-reservoir units, the aquitards. Generally speaking, these data do not exist.

We arranged for a limited number of permeability measurements on two geologic units identified by Al-Shaieb et al. (1992) as pressure seals. We submitted 2 samples from each of 4 sites, taking care to submit samples that exhibited diagenetic banding or otherwise appeared to suggest low permeability. The results were inconclusive. One of the supposed seal layers (the Simpson sandstone) had a permeability too high to allow it to function as a pressure seal ($\sim 10^{-17}$ m^2). However, measurements on the Devonian Woodford Shale (Figure 7) yielded permeabilities that are amongst the

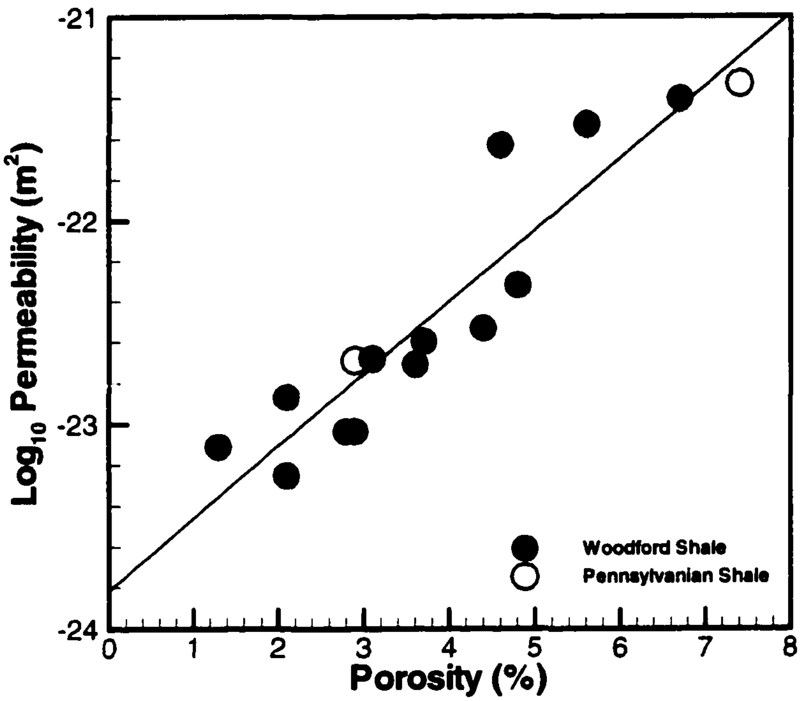


Figure 7. Porosity versus $\log_{10}$ permeability for Woodford Shale and upper Pennsylvanian shale samples.

lowest ever measured (~10^{-22} to 10^{-24} m^2). We also measured two samples of a Pennsylvanian-age shale which yielded very low permeabilities (Figure 7).

8. Hypothesis Testing

8.1 Compaction Disequilibrium

As discussed earlier, we conceptualize two end-member hypotheses for the cause of overpressuring in the Anadarko Basin. The first of these is compaction disequilibrium, wherein high fluid pressures were originally developed during rapid Paleozoic sedimentation (see Figure 5) about 250-300 Ma and have been sustained for several hundred million years by pressure seals. Compaction disequilibrium is "an important, if not the prime, source of overpressures in many basins" (Kooi, 1997). However, Kooi (1997) concluded that it was difficult to explain high fluid pressures in pre-Cenozoic sedimentary basins with this mechanism.

Extremely low rock permeabilities are necessary if compaction disequilibrium is to be successfully invoked as a mechanism to explain overpressuring in the Anadarko Basin. The simplest way to demonstrate this is with a scale analysis (Deming, 1994). It is well known that for diffusion problems, the characteristic time (t) for a hydraulic disturbance to diffuse through a layer of thickness z is (Carslaw and Jaeger, 1959; Bredehoeft and Hanshaw, 1968; Neuzil, 1986; Bethke, 1988; Deming, 1994)

$$t = z^2 \alpha \mu / 4k \tag{9}$$

where t is time, z is thickness, α is compressibility (Pa^{-1}), μ is fluid viscosity (kg/m-s), and k is permeability (m^2). If we take μ to be the dynamic viscosity of pure water at 95 °C (~3.0×10^{-4} kg/m-s) (Touloukian et al., 1975) and shale compressibility (α) to be 10^{-9} Pa^{-1}, then equation 9 reduces to

$$t\,(\mathrm{Ma}) = (z^2/k) \times 2.4 \times 10^{-27} \tag{10}$$

where k is the permeability (m^2) of a layer of thickness z (m) necessary to confine anomalous pressures for a time t (Ma). If overpressures in the Anadarko Basin are due to compaction disequilibrium resulting from Paleozoic sedimentation (Figure 5), then they have persisted for 250 Ma. Thus, according to equation 10, a pressure seal 100 m thick would of necessity have a permeability lower than about 10^{-25} m^2. If pressure seals are thinner zones of diagenic alteration, even lower permeabilities are required. Seals of 10-m and 1-m-thickness would require permeabilities of 10^{-27} m^2 and 10^{-29} m^2, respectively. These numbers are so low as to essentially be equivalent to zero.

We also evaluated the compaction disequilibrium hypothesis using a one-dimensional numerical model of pressure diffusion. The initial condition was specified as lithostatic pressure through a homogeneous section which varied in thickness from 6 to 12 km. Boundary conditions were prescribed as constant hydrostatic pressure. We then determined present day maximum fluid pressure as a function of permeability (Figure 8). To satisfy observed constraints, simulations had to result in a normalized fluid pressure P_n of 0.6 at the depth of maximum absolute pressure. This requirement demands an average basin permeability of the order of 10^{-23} m^2 (Figure 8).

8.2 Hydrocarbon Generation

The second hypothesis we evaluated as a possible cause of overpressuring was hydrocarbon generation. The most critical variables in these simulations were temperature and permeability. The enormous volume change associated with gas generation has a strong overpressuring effect, and gas generation has an exponential dependence upon temperature. Our approach was therefore to determine the

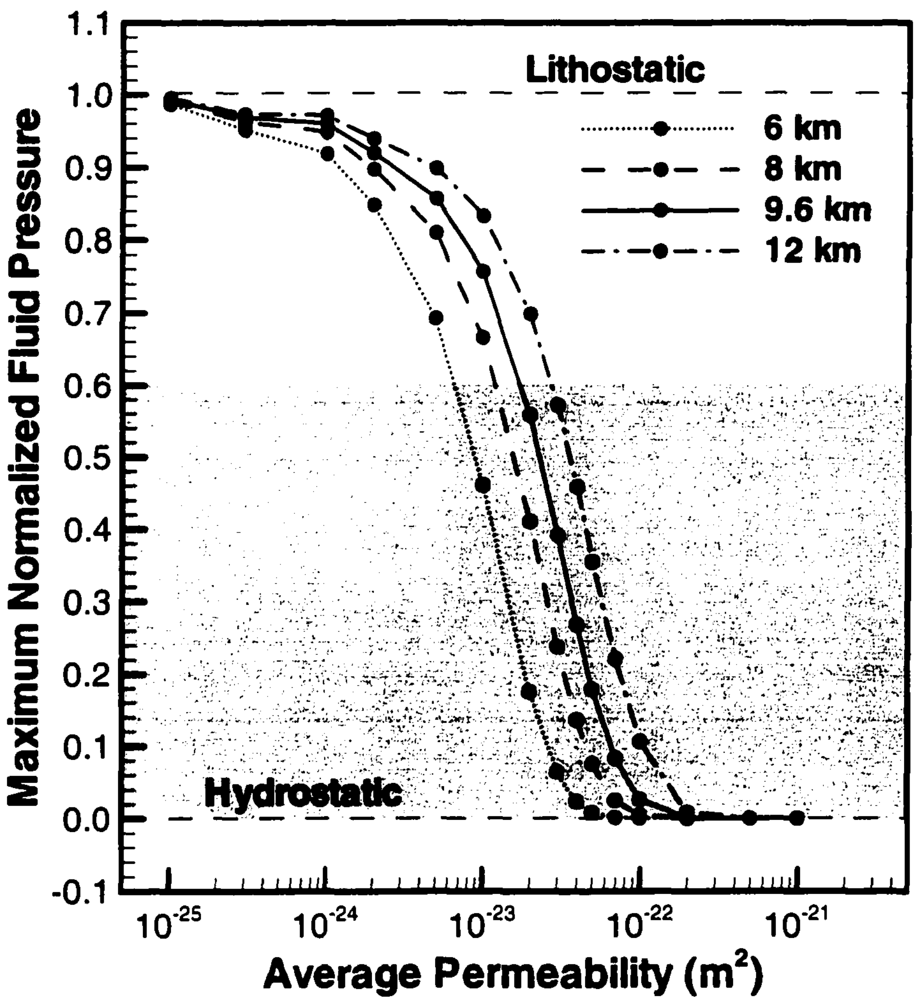


Figure 8. Maximum normalized pressure (P_n) at present day as calculated by a homogeneous numerical model to test a compaction disequilibrium hypothesis. Shaded area corresponds to present day observed fluid pressures in the Anadarko Basin. Modeling results show that average basin permeabilities in the neighborhood of 10^{-23} m^2 are required to explain present day overpressures as resulting from Paleozoic compaction disequilibrium.

permeability necessary to satisfy present day fluid pressures by varying parameters affecting the thermal state through ranges constrained by geologic data. Lacking any constraint on the pre-Cenozoic thermal state, we held the thermal gradient at a constant value ranging from 21 to 25 °C/km up to 50 Ma, and then linearly decreased it to the present day average value of 21 °C/km. Similarly, the burial history was varied through a "maximum" and "minimum" range of Mesozoic sedimentation and Tertiary erosion (Figure 5). The confluence of different burial and thermal histories had the effect of combining to produce "hot" and "cold" histories (Figures 9 and 10).

In general, we found that overpressuring can be achieved with the highest permeabilities if the thermal gradient has been constant through time with the least erosion. In other words, colder thermal histories allow present day hydrocarbon generation (Figure 10). Under the "cold" end case scenario, the thermal gradient is constant and only 1 km of Mesozoic sedimentation and subsequent erosion occur. In this case, an average basin permeability of about 10^{-21} m^2 is required to explain present day overpressuring (Figure 9), and hydrocarbon generation is ongoing for Pennsylvanian-age source rocks between depths of about 4.5 and 6.5 km. These depths correspond reasonably well to the depths at which maximum overpresssures are observed today (Figure 10).

In the "hot" end case, the thermal gradient cooled from 25 to 21 °C/km, and 3 km of Mesozoic sedimentation and Tertiary erosion occurred. If hotter conditions prevailed in the past, an average basin permeability near 10^{-23} m^2 is required to produce high fluid pressures (P_n = 0.6). In this scenario, both Pennsylvanian-age, Devonian-age, and Ordovician-age source rocks are totally exhausted. Upper Pennsylvanian-age (Virgilian) source rocks remain viable, but are not currently at high enough temperatures to generate hydrocarbons. There is no present day mechanism for overpressuring, and the very low inferred permeabilities (10^{-23} m^2) are required to preserve anomalous pressures generated tens of millions of years previous.

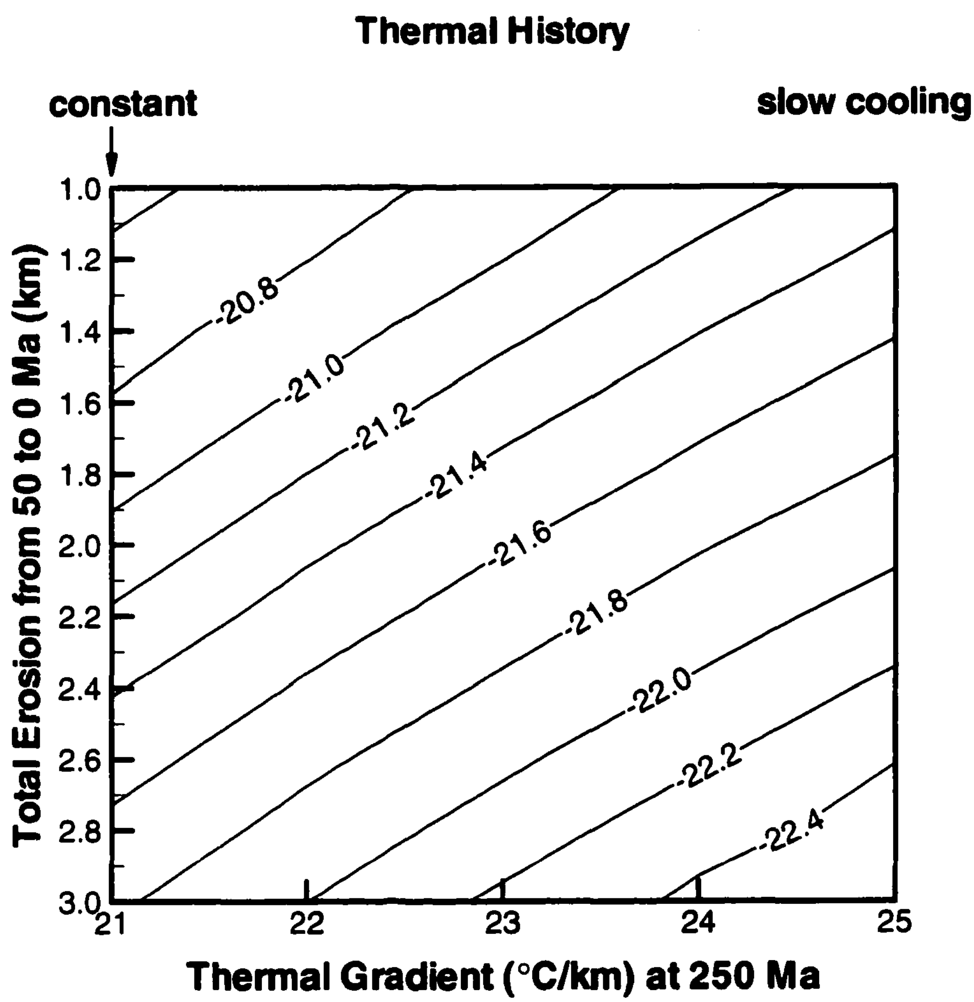
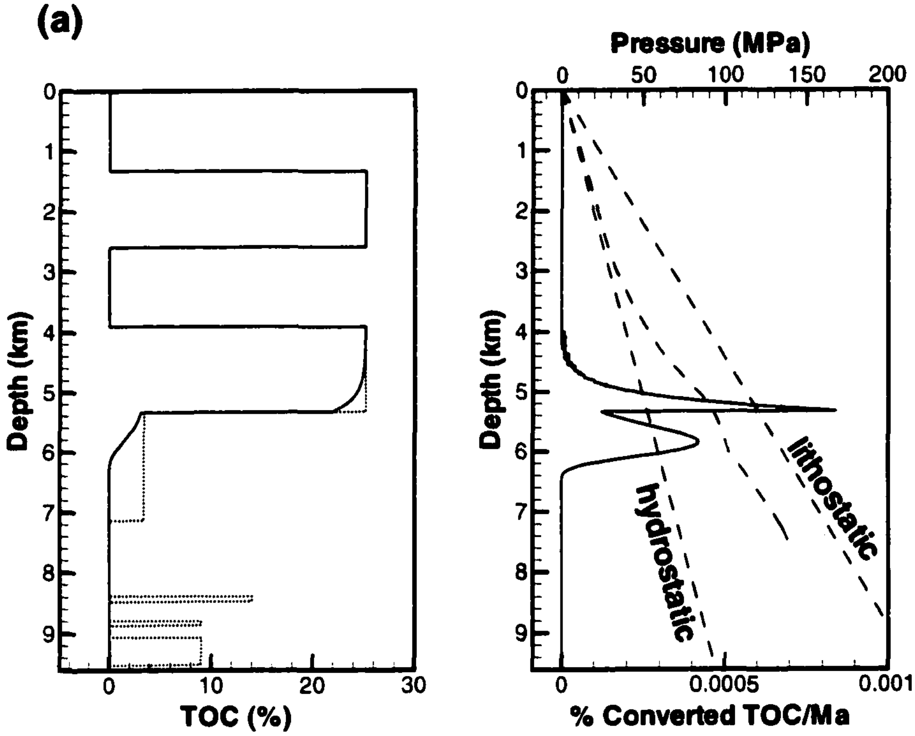


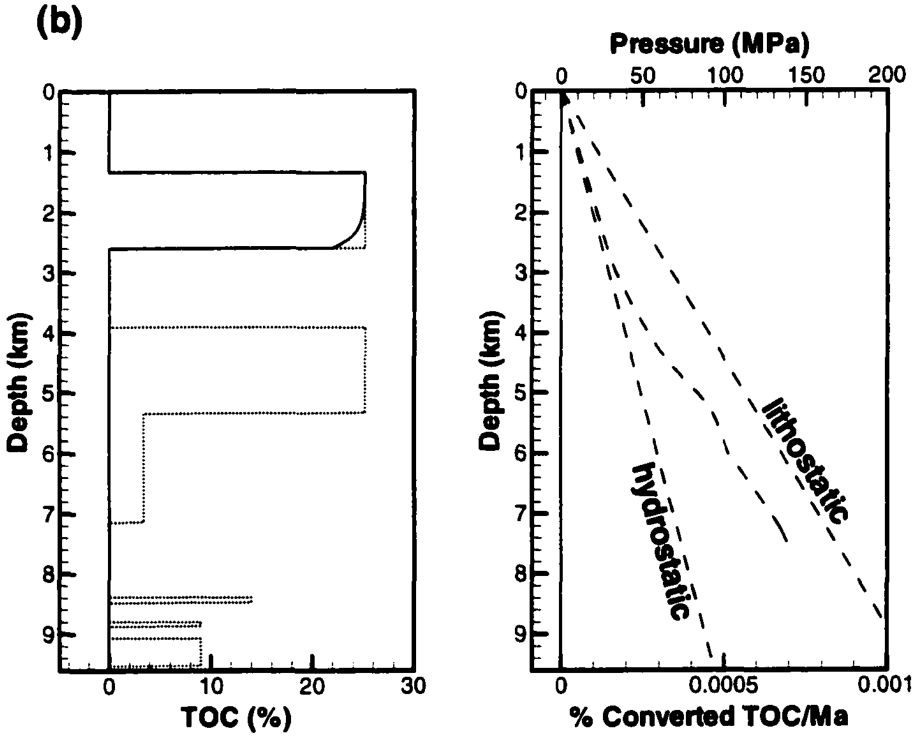
Figure 9. Log_{10} of average basin permeability required to reproduce observed present day overpressures ($P_n = 0.6$ at depth of maximum pressure) as calculated by a homogeneous numerical model for different thermal and burial histories. In these simulations, overpressuring is induced by hydrocarbon generation. In general, lower permeabilities (10^{-23} m^2) are required for hotter thermal histories (lower right corner). Colder thermal histories (upper left corner) allow overpressuring to exist at higher average permeabilities (10^{-21} m^2).

(a)



Pressure (MPa)
0 50 100 150 200
Depth (km)
0 1 2 3 4 5 6 7 8 9
TOC (%)
0 10 20 30
% Converted TOC/Ma
0 0.0005 0.001
hydrostatic
lithostatic

(b)



Pressure (MPa)
0 50 100 150 200
Depth (km)
0 1 2 3 4 5 6 7 8 9
TOC (%)
0 10 20 30
% Converted TOC/Ma
0 0.0005 0.001
hydrostatic
lithostatic

Figure 10. Distribution of total organic carbon (TOC) in source rocks, present day hydrocarbon generation rate, and observed present day fluid pressures for (a) "cold" end-case burial and thermal history and (b) "hot" end-case burial and thermal history as described in text. Solid line shows existing TOC, dotted line shows original TOC before hydrocarbon generation. In "hot" end-case scenario, Pennsylvanian-age source rocks are totally exhausted and there is no present day hydrocarbon generation. In "cold" end-case scenario, present day hydrocarbon generation occurs between 4.5 and 6.5 km, near the depths at which maximum overpressures are observed.

A comparison of the compaction disequilibrium and hydrocarbon generation hypotheses is shown in Figure 11. In both cases, permeability was held at a constant value of 5 x 10^{-22} m^2, and the thermal gradient decreased from 23 to 21 °C/km. For an initial condition of lithostatic pressure, high fluid pressures due to compaction disequilibrium dissipate by 200 Ma. However, for burial histories with between 1 and 3 km of sedimentation and erosion, model simulations with hydrocarbon generation can successfully reproduce present day maximum fluid pressures (P_n = 0.6). In general, the essential difference between the two hypotheses is that average permeabilities 1 to 2 orders of magnitude lower are required if compaction disequilibrium is to be invoked as a cause of present day overpressuring.

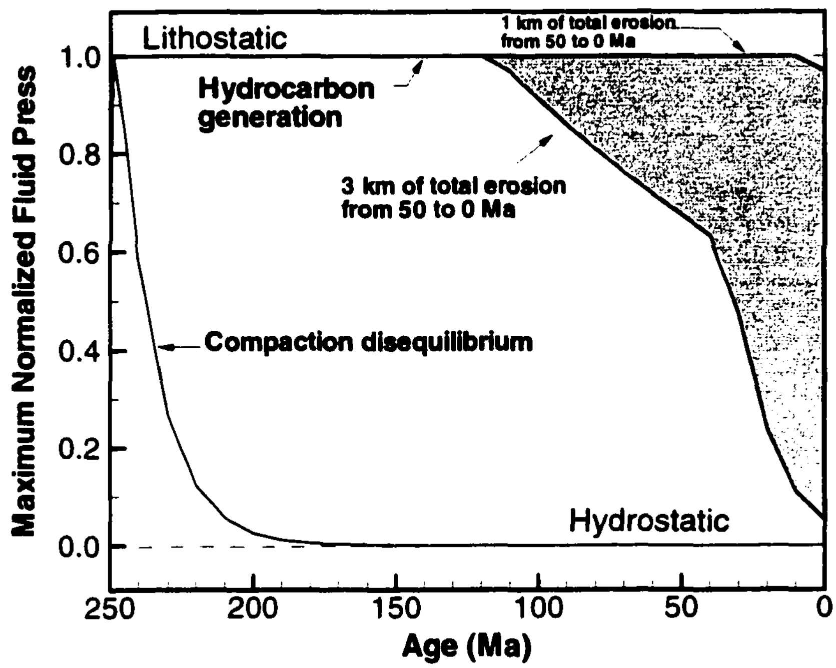
8.3 Sensitivity Analyses and Uncertainties

Different approaches have been taken in the use of numerical models to represent and study geologic processes. Some modelers use one preferred set of values for model parameters and do not explore the consequences of uncertainties. This approach is useful for the purpose of demonstrating the feasibility of a proposed process, but cannot be used to establish causality unless the difficult question of uncertainty is dealt with.

An ideal application of numerical modeling would rely upon physical parameters constrained by precise and exhaustive physical and chemical measurements. The uncertainties introduced by variations in model parameters could then be explored through a series of sensitivity analyses. In practice, this ideal is rarely approached. Geologic problems are usually complex and depend upon a wide range of physical and chemical variables which are often poorly constrained.

Our approach in this study is intermediate. We have reasonably good constraints on the thermal history and some source-rock characteristics. Recognizing that the most critical variable, permeability, is essentially unconstrained, we have defined the

(a)


Lithostatic
1 km of total erosion from 50 to 0 Ma
Hydrocarbon generation
3 km of total erosion from 50 to 0 Ma
Compaction disequilibrium
Hydrostatic
Maximum Normalized Fluid Press
1.0
0.8
0.6
0.4
0.2
0.0
250
200
150
100
50
0
Age (Ma)

(b)

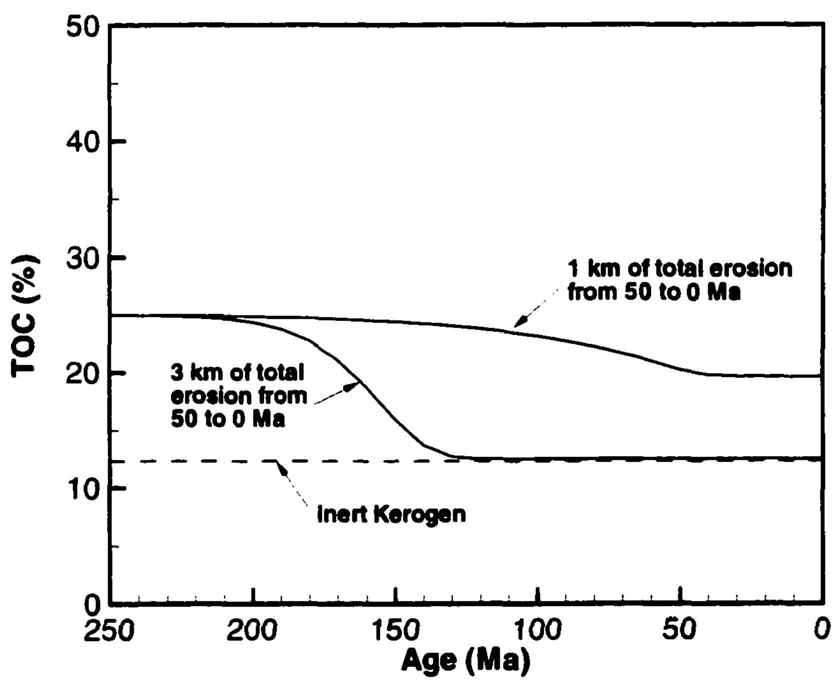

1 km of total erosion from 50 to 0 Ma
3 km of total erosion from 50 to 0 Ma
Inert Kerogen
TOC (%)
50
40
30
20
10
0
250
200
150
100
50
0
Age (Ma)

Figure 11. (a) Maximum normalized pressure (P_n) as a function of time at a fixed depth of 5.14 km as calculated by a homogeneous numerical model. Permeability is held constant at a value of 5 x 10^{-22} m^2. In these simulations, the thermal gradient is a constant 23 °C/km from 250 to 50 Ma, then decreases linearly to 21 °C/km at present day. Compaction disequilibrium is simulated by assuming lithostatic fluid pressures at 250 Ma, when rapid Paleozoic sedimentation ended. Hydrocarbon generation scenarios show that present day overpressures can be reproduced if total erosion was between 1 and 3 km.

(b) Total organic carbon (TOC) as a function of time as calculated by a kinetic model for a source rock in the Pennsylvanian Atoka formation at 5.14 km depth. Hotter thermal histories with 3 km of sedimentation and erosion result in reactive source-rock exhaustion by about 130 Ma and make it substantially more difficult to ascribe present day overpressuring to hydrocarbon generation.

problem by casting our results in terms of the permeability values necessary to validate hypotheses. Sensitivity analyses are implicit in our variation of thermal and burial histories. We recognize that the values we used for variables such as specific storage, matrix and fluid densities, etc., are not constant, but variables which are dependent upon factors such as lithology, temperature, and pressure. However, in most cases the uncertainties inherent in these variables is considerably less than an order of magnitude. As the number of variables becomes very large an exhaustive set of sensitivity analyses would lead to incoherent ambiguity. Pragmatically, we interpret the effect of these uncertainties as implying that the permeabilities inferred from our scale analyses and numerical modeling have implicit uncertainties of an order of magnitude. For example, if mean activation energies (E) are varied by plus-or-minus one standard deviation (Table 2), inferred permeabilities change by a factor of 2 to 3.

9. Discussion

9.1 Permeability Structure

Our primary results can be summarized in terms of the maximum average permeabilities required to lead to overpressuring in the Anadarko Basin. If overpressures originate from compaction disequilibrium, the average basin permeability must be lower than 10^{-23} m^2. If overpressures originate from hydrocarbon (mostly gas) generation, the average basin permeability can be as high as 10^{-21} m^2 depending upon the basin's thermal history. For a given thermal and burial history, the uncertainty inherent in these numbers is about a factor of 10.

The results of our modeling must be interpreted with an understanding of the importance of physical scaling. A basin consisting of interlayered and mixed units of highly contrasting permeability will respond differently than a thick, homogeneous layer (Bethke, 1988). Although our simple, one-dimensional model provides insight into the physics of pressure generation and dissipation, the assumed homogeneous permeability structure is unrealistic. The permeability of sedimentary rocks in the Anadarko Basin likely varies from maximum values near 10^{-12} m^2 for certain hydrocarbon reservoirs and other rocks, to values as low as 10^{-23} m^2 for the Woodford Shale (Figure 7).

The nominal thickness (9.6 km) used in most of our model simulations makes it easier to confine high fluid pressures with higher permeabilities. However, in nature low-permeability confining layers must constitute some fraction of the entire sedimentary section. Therefore, the permeabilities of the confining layers in the Anadarko Basin must be substantially lower than the average values deduced from our modeling exercises. The permeabilities of these confining layers can be estimated by noting that for diffusion problems the ratio of thickness squared (z^2) to permeability (k) must be constant (equation 10). If confining rocks constitute 10% of the sedimentary section in the Anadarko Basin (about 1 km thickness), application of equation 10 implies that permeabilities of 10^{-25} m^2 or lower are necessary to validate the compaction disequilibrium hypothesis. Even a scenario with energetic gas generation necessitates permeabilities in the range of 10^{-23} to 10^{-25} m^2. More realistically, if confining layers have thicknesses in the neighborhood of 100 m or less, permeabilities in the range of 10^{-25} to 10^{-27} m^2 are necessary to explain the occurrence of present day overpressures. These permeabilities are uncomfortably low. Neuzil (1994) reviewed permeability data for clays and shales, and found that permeability on the laboratory scale (about 0.1 m) ranged from 10^{-15} to 10^{-23} m^2 for a number of argillaceous media. Laboratory measurements are usually interpreted as minimum estimates of *in situ* permeabilities as permeability tends to increase with scale. It is thus difficult to reasonably conclude that any sedimentary rock could act as a pressure seal unless it were altered diagenetically to a layer of virtually zero permeability. Furthermore, some type of active sealing mechanism would probably be necessary, as it seems unlikely that any rock layer could escape fracturing or faulting for hundreds of millions of years. Our measurements on the Woodford Shale found permeabilities as low as 10^{-23} m^2. However, the Woodford is an atypical shale; deeply buried, dense, and highly indurated. Scaling our results down to the thickness of the Woodford Shale (100 m) would imply that a permeability lower than 10^{-25} m^2 would be necessary even if the Anadarko Basin has undergone a minimal amount of Cenozoic cooling.

In Cenozoic basins undergoing active sedimentation (e.g., the Gulf Coast Basin of the southeastern US), overpressuring is readily attributable to the rapid accumulation of several kilometers of shale with a permeability in the range of 10^{-16} to 10^{-20} m^2 (Bethke, 1986). The Anadarko Basin is very much different from this setting; it contains less than 50% shale (Figure 2) and lacks any discernible geologic process which can apparently account for pressure generation. The primary problem is that the AFT and vitrinite reflectance data indicate unambiguously that the basin has been subject to Cenozoic erosion greater than a kilometer, with cooling of at least about 20 °C. This cooling effectively dampens the most promising mechanism for overpressuring, hydrocarbon generation. As a result, we are led inexorably to conclude that our understanding of why and how the Anadarko Basin is overpressured is incomplete.

Several possibilities suggest themselves. It may be that there exists some unknown process which allows the Anadarko Basin to remain overpressured. If such a process exists, it would probably not be one of the many mechanisms commonly discussed in the geologic literature. For example, processes such as aquathermal pressuring and clay alteration are driven by temperature, just as hydrocarbon generation is. Whatever mechanism is operating must not be substantially influenced by Cenozoic cooling.

Another complication which we have not considered or described is compartmentalization. Although our level of analysis does not allow us to discern small-scale pressure compartments, Al-Shaieb et al. (1992, 1994a, b) describe overpressures in the Anadarko Basin as being compartmentalized. Instead of one monolithic overpressured area, they describe a broad area of overpressured compartments existing next to hydrostatically-pressured units. Empirical evidence of compartmentalization is often encountered during drilling in the deep Anadarko Basin (Kennedy, 1982, p. 77). It is common to encounter "kicks", which are sudden increases in pressure due to the rush of gas into a borehole which penetrates a zone of relatively high permeability. If the invaded zone were not highly permeable, the "kick" could not occur. And if the permeable zone were not surrounded by less permeable boundaries, high pressures could not exist. Thus a compartment is required. The presence of compartmentalization makes it difficult to ascribe pressure confinement in the Anadarko Basin to a few low-permeability units such as the Woodford Shale or Permian evaporites. Rather, there must exist a multitude of relatively thin pressure seals. The results of our modeling imply that these pressure seals must have permeabilities so low as to be effectively zero. Furthermore, the seals must exist in all three dimensions. The inferred presence of these pressure seals is in contradiction to established hydrodynamic paradigms, and appears to partially validate the static viewpoint. We are not predisposed to hydrodynamic theories, but find it difficult to escape from a static interpretation in this case.

9.2 Gas Capillary Seals

Larry Cathles and his coworkers at Cornell have recently proposed that abnormal pressures can be preserved for long periods of time in sedimentary basins through capillary seals induced by the presence of gas (Revil et al., 1998). In brief, the idea is that capillary forces prevent gas from being displaced from relatively coarse-grained rocks into relatively fine-grained lithologies. A gas capillary-seal is a layer of zero permeability to fluid phases unless fluid pressures become high enough to overcome capillary forces and displace pore gasses. Formations that consist of alternating layers of relatively coarse- and fine-grained sediment may develop several barriers whose effect is cumulative.

Although we are not necessarily inclined to either favor or disfavor this hypothesis, it must be conceded that it appears to have several advantages. The gas-generation hypothesis is parsimonious in that it simultaneously provides a mechanism for both generating and maintaining anomalous fluid pressures. Capillary sealing provides a mechanism for achieving zero permeabilities without contradicting established dynamic paradigms which maintain that aquicludes are rare to nonexistent. Capillary sealing also explains how pressure compartments can be sealed in all three dimensions. Gas follows fluids along the path of least resistance until all escape routes are plugged. Finally, sealing by capillary forces can produce the type of compartmentalization apparently observed in basins such as the Anadarko. The existence of levels of pressure compartmentalization is difficult to explain unless some type of physical or chemical sealing-mechanism is invoked.

More generally, it is worth noting that the presence of gas in sedimentary basins is commonly associated with anomalous pressures, both overpressures and underpressures. Subnormal pressures are associated with gas accumulations in the Western Canadian Sedimentary Basin in Alberta and the San Juan Basin in New Mexico (Hunt, 1990; Davis, 1984). Overpressures are found in association with gas in the Red Desert and Green River Basins of Wyoming (Davis, 1984). Davis (1984, p. 189) stated that the gas-saturated sections of North American sedimentary basins were "never" normally pressured. The strong confluence of gas with anomalous subsurface pressures suggests there may be a causal relationship between the presence of gas and reduced permeability.

9.3 Effect of Pressure on Hydrocarbon Generation

Although it is well known that hydrocarbon generation from kerogen has an exponential dependence upon temperature, the dependence on pressure has received less attention (Price et al., 1983; Domine, 1991; Price and Wenger, 1992; Domine and Enguehard, 1992; Fang et al., 1995; Jackson et al, 1995). Recent experimental work shows that increasing pressure can increase the activation energy of hydrocarbon generation, thus retarding hydrocarbon generation (Fang et al., 1995). If this effect were significant, it might explain how overpressures could be maintained in sedimentary basins for long periods of time through a negative feedback mechanism.

However, the effect of ambient pressure on hydrocarbon generation seems to be most pronounced at moderate pressures, i.e., below 20 MPa (Ungerer, 1993). For example, an experimental result from hexane pyrolysis showed that rates of hydrocarbon generation did not change drastically between 20 and 200 MPa (Domine and Enguehard, 1992). A pressure of 20 MPa corresponds approximately to the hydrostatic fluid pressure at 2 km depth (Figure 3). In the Anadarko Basin, overpressuring begins at about 2.5 km. The range of fluid pressures encountered in the overpressured zone is about 30 to 150 MPa (Figure 3). Thus, we tentatively conclude that high fluid pressures in the Anadarko Basin have not significantly retarded hydrocarbon generation. However, the magnitude of pressure dependence is still uncertain and more experiments for a wide range of conditions are needed (Jackson et al., 1995).

10. Conclusions

1. Although old and tectonically quiescent, the Anadarko Basin in southwestern Oklahoma is extensively overpressured. Well-head shut-in pressures and mud

weights collected in the vicinity of a north-south cross section through the basin show that overpressures are first encountered at a depth of about 2.5 km. Normalized fluid pressures P_n range from about 0.4 to 0.8, where $P_n = 0$ is hydrostatic pressure, and P_n =1 is lithostatic pressure.

2. The average thermal gradient along a north-south cross section through the Anadarko Basin is 21 °C/km.

3. Although it is difficult to reconstruct eroded sections, the Anadarko Basin appears to have experienced Mesozoic sedimentation followed by 1 to 3 km of Cenozoic erosion starting about 40 to 50 Ma.

4. The average thermal gradient in the Anadarko Basin as estimated from vitrinite reflectance data was 21 to 25 °C/km at the time of maximum burial, about 40 to 50 Ma.

5. To explain overpressuring in the Anadarko Basin as resulting from the lingering effect of Paleozoic compaction disequilibrium requires that the average permeability of the 10-km-thick basin be in the neighborhood of 10^{-23} m^2. More realistically, if pressures have been preserved through low permeability layers whose thickness is 100 m or less, permeabilities lower than 10^{-27} m^2 are required. These inferred permeabilities are uncertain by a factor of 10.

6. To explain overpressuring in the Anadarko Basin as resulting from hydrocarbon generation requires that the average permeability of the 10-km-thick basin be in the range of 10^{-21} to 10^{-23} m^2. Lower permeabilities are required if the basin was hotter in the past, because a hotter thermal history has exhausted the generation potential of existing source rocks. More realistically, if pressures have been preserved through low permeability layers whose thickness is of the order of 100 m or less, permeabilities in the range of 10^{-25} to 10^{-27} m^2 or lower are required. These inferred permeabilities are uncertain by a factor of 10, but are as low or lower than the lowest values ever recorded for sedimentary rocks.

7. Due primarily to the apparently well-constrained geologic fact that the Anadarko Basin has experienced 1 to 3 km of uplift, erosion, and cooling over the last tens of millions of years, it is difficult to reconcile the existence of overpressuring with any known geologic mechanism. Permeabilities which are at or below the extreme end of measured values appear to be necessary. The pressure regime in the Anadarko Basin is apparently compartmentalized by the existence of numerous and relatively thin layers of near-zero permeability.

8. The presence of anomalous pressures in older sedimentary basins does not appear to be analogous to compaction disequilibrium in Cenozoic basins undergoing relatively rapid sedimentation. The Anadarko appears to belong to a group of older basins in which anomalous pressures are associated with the presence of gas.

9. The presence of overpressures in the Anadarko Basin appears to be better explained by static paradigms which invoke pressure seals than by classic hydrodynamic theories which rely upon the slow diffusion of pressure transients through aquitards. At the present time, the occurrence and preservation of geopressures in the Anadarko Basin seems to be best explained by gas generation and gas capillary seals.

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