Thermal effects of Zechstein salt and the Early to Middle Jurassic hydrothermal event in the central Polish Basin

Gary W. Zielinski, Pawel Poprawa, Jan Szewczyk, Izabella Grotek, Hubert Kiersnowski, and Robyn L. B. Zielinski

ABSTRACT

Deep gas potential in the Polish Basin may factor significantly in European geopolitics, and thermal effects can influence that outcome there and elsewhere. Deep (>3 km [9843 ft]) well data from the Kujawy area of the central Polish Basin reveal average geothermal gradient (36°C/km), thermal conductivity of Mesozoic strata (k = 2.29 W/m K), and present-day heat flow (Q = 82.4 mW/m^2) that is 3% less than that obtained using the entire borehole. The extrapolated surface temperature (–6.2°C) is in good agreement with temperatures during the Weichselian glaciation. The thermal conductivity of the Upper Permian Zechstein (4.89 W/m K) is in good agreement with values from the North Sea and northern Germany. Steady-state heat-flow theory (one-dimensional [1-D]) predicts present-day temperature (199°C) at the base of Zechstein cap rock at 6-km (19,685-ft) depth in Kujawy. This is reduced just more than 10°C by low Zechstein thermal gradients (16.8°C/km). Because of thermal refraction, two-dimensional and three-dimensional models of Zechstein salt pillows can significantly negate this cooling effect; however, such effects appear absent in the Kujawy wells studied.

A widespread Early to Middle Jurassic (~195–175 Ma) hydrothermal event appears to have reached maximum in the Kujawy area. A 455°C paleotemperature at 7-km (22,966-ft) depth (Carboniferous) is predicted by 1-D conductive heat transfer; however, geologic evidence does not support this result. The discrepancy is reconciled by convective heat transfer

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with upward fluid flow (3.3×10^{-10} m/s [10.8×10^{-10} ft/s]), resulting in a maximum paleotemperature of 273°C at 7-km (22,966-ft) depth, despite a paleoheat flow of 142 mW/m^2. The trend of intensity of the hydrothermal event correlates with the present-day heat-flow trend. Hydrothermal event sites are subparallel to the major northwest-southeast structural and regional heat-flow trend, whereas other sites as close as 14 km (45,932 ft) and without hydrothermal event are not. The decay of the hydrothermal event is consistent with localized cylindrical plumes (10-km [32,808-ft] radius) that cool by conduction. Results suggest a long-term (~185 m.y.) structural control on heat flow. Linear regression to vitrinite paleotemperatures yields a 185-Ma Jurassic surface temperature of approximately 21.3°C that is approximately 13°C higher than the present-day temperature for Warsaw, Poland. The duration of maximum reservoir and source rock paleotemperature (<50 m.y.) is contrary to the kinetics of nitrogen and CO_2-producing wells. Equilibrium thermodynamics predicts approximately 60% methane for present-day Kujawy reservoirs, with considerable uncertainty that should be removed by anticipated new deep drilling.

INTRODUCTION

Temperature is important in the formation of hydrocarbons (Tissot and Espitalie, 1975; Tissot and Welte, 1978). Heat flow, the flux of heat at the surface of Earth, $Q = -k\frac{dT}{dz}$ (where k is thermal conductivity, T is temperature, and z is depth), provides an important boundary condition for calculating temperatures at depth and for comparing the thermal regimes of different regions. As of the 1990s, the deepest petroleum exploration well (9583 m [31,440 ft]) of the world was drilled in the Anadarko Basin, United States. Particularly at such depths, temperature and other factors affect the presence and relative abundances of hydrocarbons, including methane (Hunt, 1979; Sassen and Moore, 1988; Barker and Takach, 1992). This also applies to nitrogen and carbon dioxide, which can occupy deep reservoirs in unwelcome abundances (Littke et al., 1995; Schoell and Cathles, 2000).

The purpose of this study is to place limits on maximum temperatures at and below 6-km (19,685-ft) depth and to quantify the thermal effects of Zechstein salt and the Early to Middle Jurassic hydrothermal event (Kozlowska and Poprawa, 2004; Poprawa and Grotek, 2004) on methane preservation in the Kujawy area of the Polish trough, located in the center of our study area (Figure 1). Relevance of results to other regions is anticipated. We make use of published heat-flow data (e.g., Majorowicz and Plewa, 1979), newly revised heat-flow values (Figure 1, from Szewczyk and Gientka, 2009), and well data provided by the Polish Geological Institute (PGI), Warsaw. Paleotemperatures are based on vitrinite reflectance data from core samples (Kozlowska and Poprawa, 2004; Poprawa and Grotek, 2004). The locations of wells used in this study and their respective heat-flow values are shown in Figure 2. The wells were drilled and cored in the 1970s and 1980s. Based on the calculated temperatures at depth, we can estimate the gas composition likely to exist in potential Rotliegende (Lower Permian) reservoirs (e.g., Hunt, 1979; Sassen and Moore, 1988; Barker and Takach, 1992). The results of this work will be put to the test by new deep (>6 km [19,685 ft]) drilling slated for the Kujawy region.

GEOLOGIC SETTING OF THE KUJAWY REGION

Our study area lies at the western edge of a vast Carboniferous–Rotliegende petroleum system (Gautier, 2003). It is defined by source rocks consisting mostly of Carboniferous coals and reservoirs consisting of Rotliegende (Lower Permian) sandstones sealed by Zechstein (Upper Permian) evaporites.

The Kujawy region (e.g., Poprawa and Grotek, 2004) occupies the central part of the Polish Basin, which is a part of a large system of Permian–Mesozoic basins of central and northwestern Europe (e.g., Ziegler, 1990). In the southern part of this system, referred to as the southern Permian Basin (Van Wees et al., 2000), a large Rotliegende gas province developed, which extends as far as western Poland and, possibly, into the Kujawy region.

The Kujawy region is characterized in Figure 3 by the continuous deposition from the Late Permian to the early Paleocene (Dadlez, 1989). Several hundred meters of sediments accumulated during that period, with Rotliegende at the base of the section. Facies development of the Rotliegende in Kujawy is not constrained by borehole data. However, based on regional models, it is represented either by playa or by eolian to fluvial sediments (Kiersnowski, 1997). In the second case, Rotliegende could be potential reservoirs for hydrocarbons.

During the Late Cretaceous and/or Paleocene, the Polish Basin was inverted and the previous northwest-southeast–trending depocenter was uplifted and partly eroded. Inversion resulted in the development of the northwest-southeast–trending Polish swell, where bases of Cenozoic, Jurassic, or Triassic sediments subcrop (Dadlez, 1997). The quantity of the section removed remains controversial.

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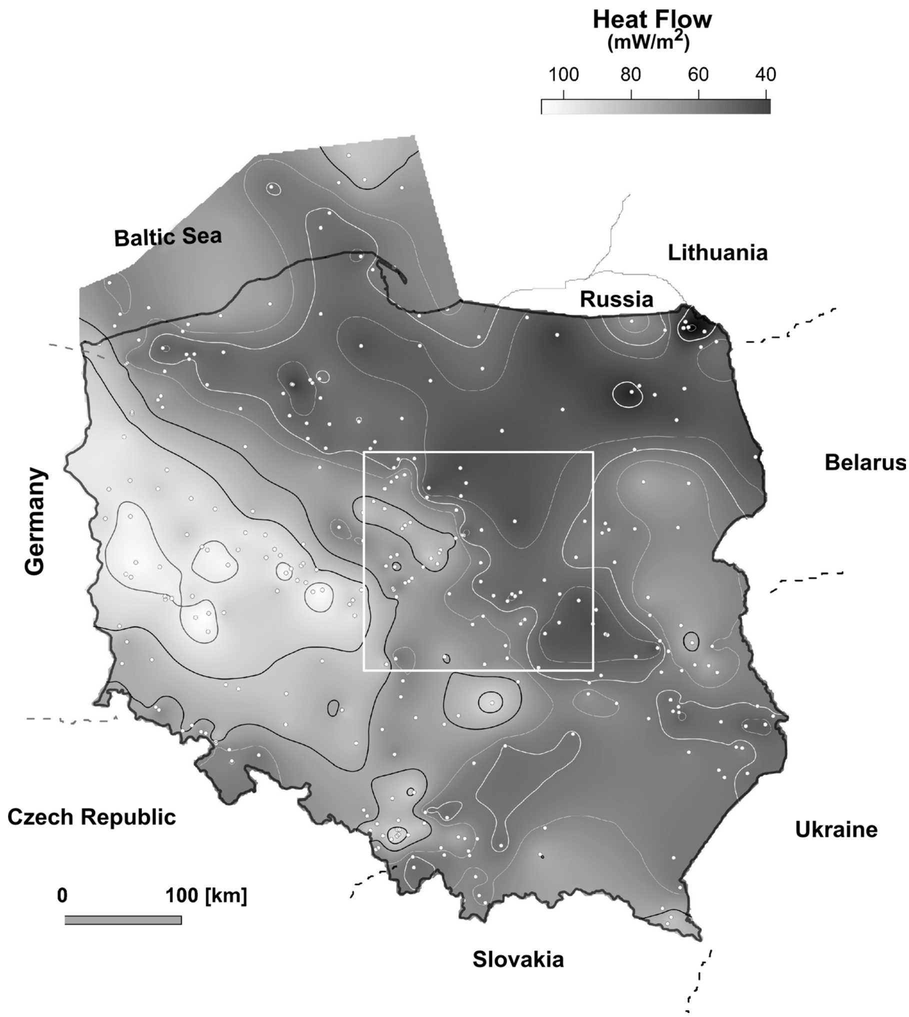

Figure 1. Study area shown in context of revised heat-flow values trending northwest-southeast (modified from Szewczyk and Gientka, 2009).

At the bottom of the Permian–Mesozoic, beneath an angular unconformity, Carboniferous sediments are documented in a few deep wells in this region, for example, Budziszewice IG-1 and Byczyna-1 (Figure 2). The Kutno-1 well (Figure 2) reached nearly 6-km (19,685-ft) depth, the last section (>500 m [1640 ft]) consisting of Zechstein salts with clay intercalations, but did not reach the Rotliegende. There, the Carboniferous is most probably represented by distal turbidites (Kulm facies) (Leszczyñski, 2008). This section contains packages of mature shale with elevated total organic carbon contents and is regarded as a potential source rock for natural gas. Regionally, Carboniferous deposits of the Variscan foreland are widely regarded as potential source rocks (Narkiewicz, 2007; Mazur et al., 2010). In Kujawy, shallow (7000–7500 m [22,966–24,606 ft]) Carboniferous source rocks are likely to be Namurian and, possibly, Westphalian (Waksmundzka, 2010), spanning approximately 290 to 320 Ma in age. An uplifted Carboniferous block is assumed to be overlain by Lower Permian Rotliegende deposits, the most probable reservoir rock for gas accumulation in the top of the structure (Kiersnowski et al., 2010). These are capped by Upper Permian Zechstein salts, likely forming an impermeable seal (Wagner, 1994). In the Kujawy area, such uplifted blocks can be capped by salt pillows that can be approximately 2 km (6562 ft) thick and approximately 12 km (39,370 ft) across (Dadlez et al., 1998). These structures are likely to have developed in several stages, mainly Late Triassic and Early to Middle Jurassic, with the final shape achieved during a Late Cretaceous–Paleocene inversion (Dadlez et al., 1998).

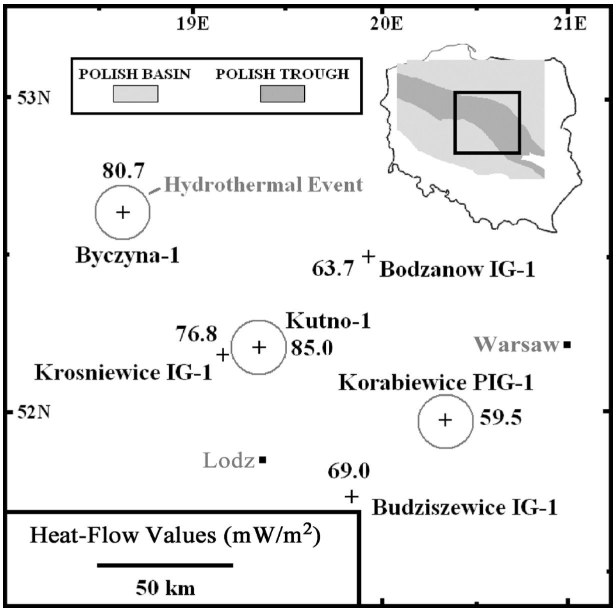

Figure 2. Locations and heat-flow values of deep (>3 km [9843 ft]) wells representing the Kujawy area of the Polish trough.

PRESENT TEMPERATURES

The heat-flow database for Poland (Szewczyk and Gientka, 2009) was recently revised (Figure 1), mainly in recognition of the large effect of the last (Weichselian) glaciation on ground surface temperatures (Safanda et al., 2004; Szewczyk, 2005; Szewczyk and Gientka, 2009) and its propagation to depths as much as 2 km (6562 ft). The new data analyses (Szewczyk and Gientka, 2009) follow the procedures described by Szewczyk (2001), resulting in thermal conductivity profiles derived from lithologic data and geophysical well logs (Brigaud et al., 1990) and equilibrium temperature profiles for each well. Figure 4 shows revised temperature data for depths greater than 3 km (9843 ft) for the Kutno-1, Bodzanow IG-1, and Byczyna-1 wells, located in Figure 2. The degree of linearity in the temperature profile for the Kutno well is highly significant ($R^2 = 0.9996$). This indicates relatively uniform thermal conductivity from 3000 to 5276 m (9842–17,310 ft). Deeper extrapolation of temperature using the regression equation, $T = 36z - 6.2$, where T is predicted temperature (°C) and z is depth (km), does not account for any difference in the thermal conductivity of the Zechstein, not reached by logging at Kutno.

The Kutno well was drilled from September 12, 1979, until December 27, 1983 (t_1 = 4.3 yr). Afterward, no circulation of fluids until April 26, 1985 (t = 1.3 yr) was observed. Lachenbruch and Brewer (1959) show in a practical case that

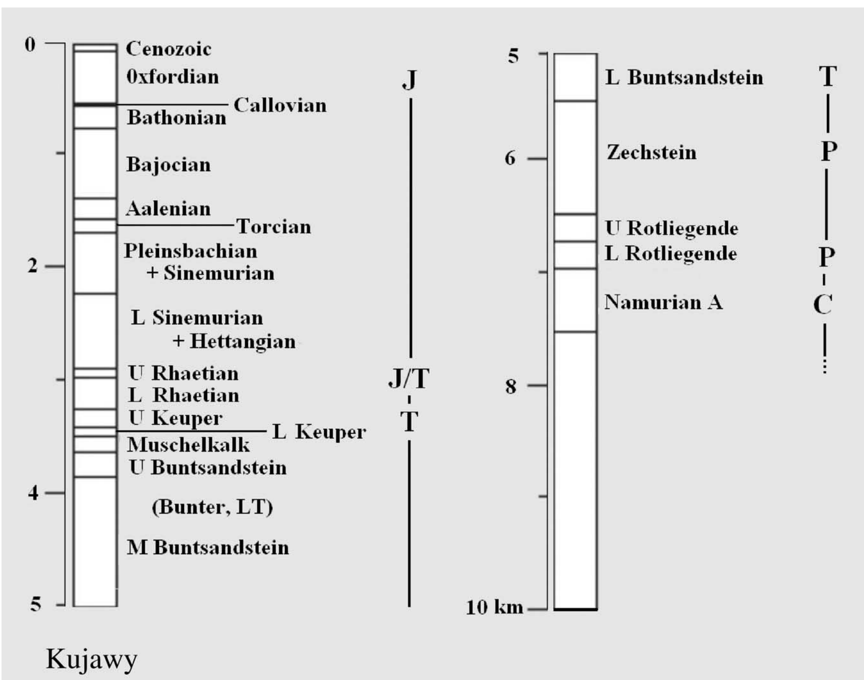

Figure 3. Stratigraphy of Kujawy. U = upper; M = middle; L = lower; J = Jurassic; T = Triassic; P = Permian; C = Carboniferous; Buntsandstein = Bunter, LT.

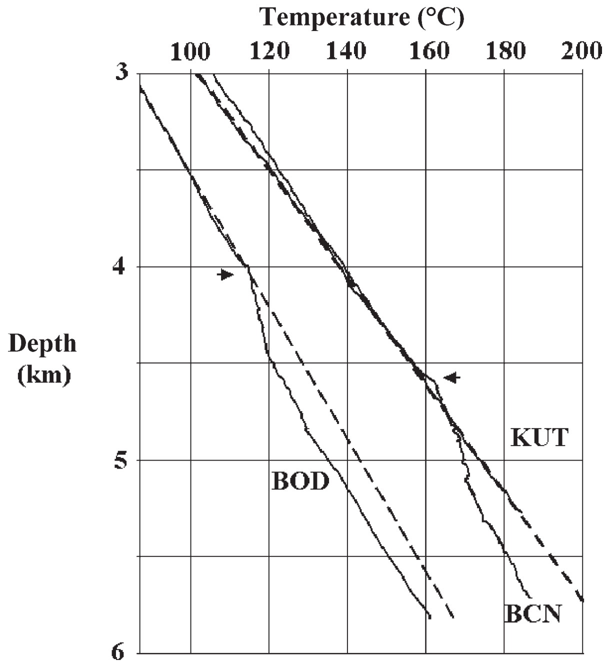

Figure 4. Equilibrium temperature logs (>3 km [9843 ft]) for three deep wells in the Kujawy area. See Figure 2 for locations. The dashed lines represent extrapolations of Bodzanow IG-1 (BOD) and Kutno-1 (KUT) data. The arrows show the top of Zechstein for BOD and for Byczyna-1 (BCN) and corresponding change in thermal gradient.

temperatures are within 0.05°C of equilibrium when $t = 3t_1$. Assuming that circulation was continuous for 4.3 yr at a constant drilling rate, only the bottom 600 m (1969 ft) meets this criterion. However, the linearity of temperatures from 3000 to 5276 m (9842–17,310 ft) (KUT, Figure 4) strongly supports thermal equilibrium over that entire interval. Fifteen days before logging temperatures on May 10, 1985, circulation was resumed for 2 hr, yielding $t = 180t_1$ and implying that no corresponding disturbance from the latter circulation remained at the time of logging.

Heat-flow values (Figure 2) for Kutno and Byczyna, 70 km (43 mi) to northwest, following the general northwest-southeast trend of the Polish trough, differ by only 5%. Their temperatures below 4-km (2-mi) depth are in good agreement down to the change in Byczyna thermal gradient (Figure 4) that correlates with an increase in thermal conductivity at the top of the Zechstein (arrow). This increase is mostly caused by the presence of approximately 5 W/m K thermal conductivity Zechstein salt, anhydrite, and dolomite from approximately 4.6-km (15,092-ft) to approximately 5.1-km (16,732-ft) depth at Byczyna. This higher conductivity, in turn, mostly accounts for the temperature at the bottom of Byczyna being reduced by nearly 15°C compared to the extrapolated temperatures (dashed) for Kutno (Figure 4). Whereas geophysical logging at Kutno did not include the Zechstein, it was confirmed by drilling to exist from nearly 5.5 km (18,045 ft) to nearly 6 km (19,685 ft).

Kutno thermal conductivity (k) from revised values obtained at 1-m (3-ft) intervals is computed as the harmonic mean of all values at depths greater than 3 km (9843 ft), where $\frac{1}{k} = \frac{1}{n}\sum_i \frac{1}{k_i}$, yielding $k =$ 2.29 W/m K for Mesozoic strata. The harmonic mean of all thermal conductivity values ($z > 0$) for Kutno is 2.27 W/m K. Using the thermal gradient (36°C/km) obtained from linear regression ($z > 3$ km [9843 ft]), we compute heat flow at 82.4 mW/m^2. This simple method of deriving heat flow is analogous to that traditionally used to measure heat flow in equilibrated small diameter boreholes (Beck, 1965). Accordingly, a tool that measures thermal gradient is lowered to a deep zone of uniform lithology and well-determined thermal conductivity, where the gradient is precisely measured and the heat flow through that interval is taken to be the value for the location. This is in contrast to using data from the entire borehole, where shallow depths are influenced by climate change. This value is only 3% less than the latest (Szewczyk and Gientka, 2009) value of 85.0 mW/m^2, and it differs from the heat flow for Byczyna (80.7 mW/m^2) by only 2%. The ground surface temperature predicted by the linear regression for Kutno is –6.2°C, which is close to the mean surface temperature (–7°C) for the Polish Basin during the last (Weichselian) glaciation (Szewczyk and Gientka, 2009), using data from the entire boreholes ($z > 0$).

Zechstein thermal conductivity values are also computed as harmonic means of values at 1-m (3-ft)–depth spacing for Bodzanow, yielding 4.89 W/m K (4063–4483 m [13,330–14,708 ft]). For Byczyna, the deepest Zechstein (4987–5053 m [16,362–16,578 ft]) yields a harmonic mean thermal conductivity of 4.94 W/m K, with a value of 4.23 W/m K for the entire section designated Zechstein

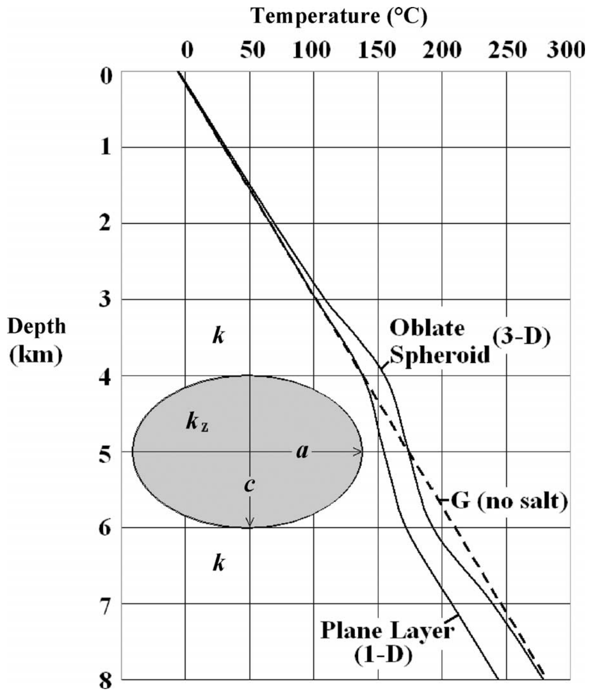

Figure 5. Temperature effect ($x = y = 0$) of an ellipsoid (oblate spheroid) 2 km (6562 ft) thick by 12 km (39,370 ft) wide, with a thermal conductivity contrast of 2.14, simulating a Kujawy area salt pillow, compared with a 2-km (6562-ft)-thick plane layer and normal geotherm (G) (dashed line). The solid geotherms illustrate the effect of the pillow on isotherms above and below it. k_z is the Zechstein thermal conductivity and k that of the surrounding strata.

(4581–5053 m [15,030–16,578 ft]). These values are consistent with measurements from the North Sea (Andrews-Speed et al., 1984) and the northeast German Basin (Norden and Förster, 2006). For steady-state heat flow, the thermal gradient within the Zechstein is given by $G_z = (k / k_z)G$. Using the Bodzanow thermal conductivity value (k_z = 4.89 W/m K) for Kutno Zechstein, Kutno values k = 2.29 W/m K and G = 36°C/km, we obtain G_z = 16.8°C/km. Then, using the observed Kutno temperatures (Figure 4) extrapolated to the top of the Zechstein at 5457 m (17,903 ft) (T = 190.2°C), we compute a temperature of approximately 199°C at 6-km (19,685-ft) depth for Kutno. This is opposed to approximately 210°C obtained by simple extrapolation that does not account for the high thermal conductivity of the Zechstein. The reduction in temperature (>10°C) is consistent with observed temperature differences between extrapolated temperatures (dashed) and data (solid) in Figure 4. Were the Zechstein to continue to 7-km (22,966-ft) depth at Kutno, the reduction in temperature would be nearly 30°C for a value of approximately 216°C. Szewczyk and Gientka (2009) report even higher values of Zechstein thermal conductivity approaching 6 W/m K. An upper limit to the temperature-lowering effect of the Zechstein can be seen by assuming infinite thermal conductivity. In that case, the Zechstein is isothermal with depth, and the temperature at its base is approximately 190°C instead of 199°C. In general, the steady-state temperature lowering effect of the Zechstein is given by $\Delta T_z = -Q\ (k^{-1} - k_z^{-1})\ \Delta z$, where Q is heat flow and Δz is thickness or depth from top into the Zechstein in meters. For Kujawy, we have $\Delta T_z = -19.1\ \Delta z$. The effect is seen to be directly proportional to heat flow (Q); hence, the PGI database (e.g., Figure 1) can be used to estimate this effect elsewhere in Poland.

Thus far, we have considered the effect of the high thermal conductivity Zechstein to reduce temperatures assuming one-dimensional (1-D) steady-state heat flow. However, in the Kujawy area, the Zechstein can cap structure in the form of a two-dimensional (2-D) or three-dimensional (3-D) pillow or lens. We model this (Figure 5) as an oblate spheroid with the center at a depth of 5 km (16,404 ft), 2 km (6562 ft) thick, 12 km (39,370 ft) across, and with thermal conductivity contrast $\varepsilon = k_z/k = 2.14$. The latter is the ratio of Zechstein thermal conductivity from Bodzanow (k_z = 4.89 W/m K) and Mesozoic rocks from Kutno (k = 2.29 W/m K). In steady state $(dT / dt) = 0$ (T is temperature and t is time), the thermal regime is governed by the Laplace equation ($\nabla^2 T = 0$) and exact analytical solutions for ellipsoids $(x^2 / a^2) + (y^2 / b^2) + (z^2 / c^2) = 1$ in infinite media are well known in potential theory and earth science (e.g., MacMillan, 1930; Carslaw and Jaeger, 1959; Von Herzen and Uyeda, 1963; Lachenbruch and Marshall, 1966; Lee and Henyey, 1974; Zielinski et al., 2007).

We make use of the solution for oblate spheroids ($a = b > c$) in Carslaw and Jaeger (1959, p. 427–428) to compute the geotherm from the surface to 8-km (26,247-ft) depth through the center ($x = y = 0$) of an ellipsoidal salt pillow (Figure 5). Outside the ellipsoid (Figure 5), the normal geothermal gradient, assumed to be that observed at the Kutno well (G = 36°C/km), is altered by the presence of the salt pillow by the factor F_o, where

$$F_o = (\varepsilon - 1)C_\lambda/(1 + C_0[\varepsilon - 1]) \qquad (1)$$

$$\lambda \text{ is the positive root of } z^2/(c^2 + \lambda) = 1 \qquad (2)$$

$$C_0 = C_\lambda \text{ for } \lambda = 0 \qquad (3)$$

$$C_\lambda = ([1 - e^2]^{1/2}/e^3)(v^{-1} - \cot^{-1} v] \qquad (4)$$

and

$$v = ([c^2 + \lambda]/[a^2 - c^2])^{1/2} \qquad (5)$$

Inside the ellipsoid (Figure 5), the normal geothermal gradient G is altered by the factor F_i, where

$$F_i = 1/(1 + C_0\ [\varepsilon - 1]) \qquad (6)$$

to the value $0.52G$ = 18.8°C/km and heat flow of approximately 92 mW/m^2. By comparison, the 1-D steady-state gradient within a plane layer of Zechstein is 16.8°C/km, as noted previously.

The higher gradient inside the ellipsoid is the result of 3-D refraction of heat by the ellipsoid, similar to the well-known effect of salt domes on heat flow (e.g., Kappelmeyer and Haenel, 1974). Thermal gradient and heat flow along with thermal conductivity are constant within the ellipsoid. Above the ellipsoid, isotherms (Figure 5) bow upward (concave down), causing higher temperatures at shallower depths; the effect seen in the 3-D geotherm diminishes to zero, 1 km (3281 ft) above the ellipsoid. Below the ellipsoid, isotherms bow downward (reduced temperatures), the effect also diminishing to zero in 1-km (3281-ft) distance from the bottom of the ellipsoid. In Figure 5, the 3-D effect, maximum for the geotherm passing through the center of the ellipsoid ($x = y = 0$), can be seen in comparison with the normal geotherm without salt and the 1-D steady-state model (plane layer). The top and bottom of the Zechstein are at 4 and 6 km (13,123 and 19,685 ft, respectively) in both salt cases. The temperature at 5-km (16,404-ft) depth ($z = 0$) inside the ellipsoid, where theory predicts no change in temperature caused by the ellipsoid, is taken to be the same as for the normal (no salt) geotherm (174.0°C). Compared with the 1-D plane-layer model, where the temperature at 5 km (16,404 ft) is 154.8°C, it is evident that most of the effect of the salt in reducing temperatures at depth is lost because of 3-D heat transfer.

The maximum temperature lowering effect of the ellipsoid occurs at approximately 6 km (19,685 ft) (base of Zechstein), where the temperature is 193°C, which is 17°C lower than that for the normal geotherm (210°C) but 21°C higher than that for the plane (1-D) layer (172°C). At best, the temperature lowering effect is seen less than half of that for the 1-D plane-layer case. By 8 km (26,247 ft), little difference in the temperature between the ellipsoid and the no-salt geotherm G exists (Figure 5), whereas the plane-layer temperature remains a constant 38°C less.

The sensitivity of results is conveniently checked using solutions for gradients within a comparable (radius = 1 km [0.6 mi]) sphere and a semiinfinite cylinder (Carslaw and Jaeger, 1959, p. 426) for a 2-1 (inside-outside) thermal conductivity contrast (ε = 2). The magnitudes of thermal focusing on gradients inside are sphere = $0.75G$, cylinder = $0.67G$, oblate spheroid (ellipsoid) = $0.56G$, and plane layer = $0.50G$. Values higher than $0.5G$ indicate proportionally higher heat flow and increase in temperature with respect to the plane layer. It appears reasonable that the flattened ellipsoid result is closest but not above that for the plane layer. It also appears reasonable that the 2-D cylinder result is less than for the 3-D sphere. This has an implication to the Kujawy salt pillows that are elongate with length 2 to 3 times width. We therefore expect a 2-D simulated salt-pillow effect to be greater than that for the plane layer, falling between 0.56 and $0.50G$, rendering the Figure 5 ellipsoid (3-D) effect an upper limit. This is confirmed by the Lachenbruch and Marshall (1966) solution for a 2-D semielliptic cylinder ($a / c = 6$; ε = 2.14), where an 8% increase in heat flow, $\varepsilon\ ([\{c / a\} + 1] / [\{c / a\} + \varepsilon]) = 1.08$, inside the cylinder is predicted compared to the 11% increase found for the oblate spheroid ($0.52\varepsilon = 1.11$) in Figure 5.

Whereas heat flow is constant everywhere within the ellipsoid, a discontinuous edge effect results in a drop in heat flow at the contact just

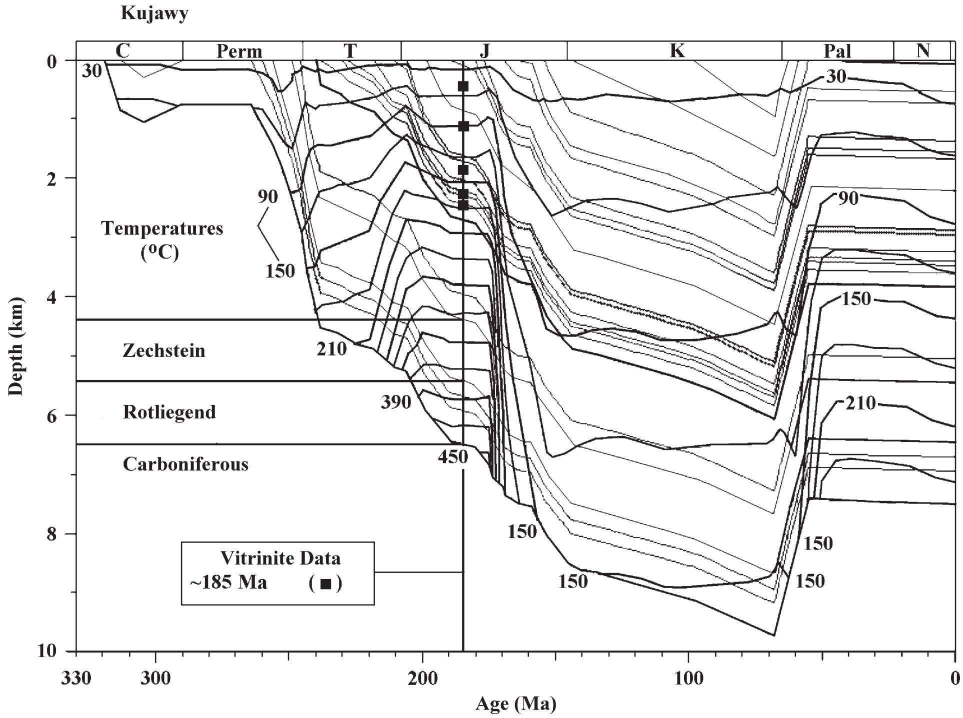

Figure 6. One-dimensional basin modeling (BasinMod™) for the Kujawy stratigraphy (Figure 3) showing vitrinite data control at approximately 185 Ma during the hydrothermal event (Poprawa and Grotek, 2004). Zechstein and the top of the Carboniferous are indicated along with temperatures of isotherms (bold lines). C = Carboniferous; Perm = Permian; T = Triassic; J = Jurassic; K = Cretaceous; Pal = Paleogene; N = Neogene.

outside the ellipsoid ($x > a$). The upper plots of Figure 5 of Lee and Henyey (1974) illustrate this for two parallel half cylinders. The magnitude of this decrease in heat flow for a semielliptic cylinder with $a / c = 6$ and $\varepsilon = 2.14$ is $(1 - \varepsilon) / ([c / a] + \varepsilon) = 0.49$, nearly 50% (Lachenbruch and Marshall, 1966).

As the salt pillow (oblate spheroid, Figure 5) causes curvature in the geotherm starting 1 km (3281 ft) above its top, the linearity of the deep geotherms studied (Figure 4), particularly for the Kutno-1 well, renders 2-D and 3-D thermal effects insignificant at those locations. There, reduction in temperature associated with the Zechstein (Figures 4, 5) of approximately 10°C appears typical for approximately 500-m (1640-ft) thickness of Zechstein. The magnitude of this effect translates to 50% reduction in reaction rates according to the Lopatin (1971) and Waples (1980) time-temperature index (TTI) model, where reaction rates double with each 10°C rise in temperature.

PALEOTEMPERATURES

Poprawa et al. (2002), Kozlowska and Poprawa (2004), and Poprawa and Grotek (2004) use 1-D modeling (BasinMod™; e.g., Figure 6) constrained by vitrinite reflectance data to postulate an Early to Middle Jurassic hydrothermal event, reflected in two wells (circled) in our study area (Figure 2), Kutno-1 and Korabiewice PIG-1. The largest, for Kutno, resulted in paleoheat flow near 130 mW/m^2 lasting 30 m.y. and then abruptly decaying to half this value in approximately 10 m.y. (Poprawa and Grotek, 2004). For the same duration at Korabiewice, the maximum paleoheat flow was approximately 80 mW/m^2 (Kozlowska and Poprawa, 2004). However, no evidence for the event was seen in Krosniewice IG-1, 14 km (45,932 ft) from Kutno-1, or Budziszewice IG-1, approximately 60 km (196,850 ft) south-southeast (Figure 2). One possible explanation is that Kutno and Korabiewice trend subparallel to the major northwest-southeast structural trend also expressed in the heat-flow contours (Figure 1), whereas Krosniewice, despite its proximity to Kutno, and Budziszewice do not. Unfortunately, we have no vitrinite data results for Byczyna, which also falls along that trend.

A simple thermal model treats the Kutno and Korabiewice anomalies as a vertically buried cylinder (Mundry, 1968; Kappelmeyer and Haenel, 1974) that decays to half its initial value when

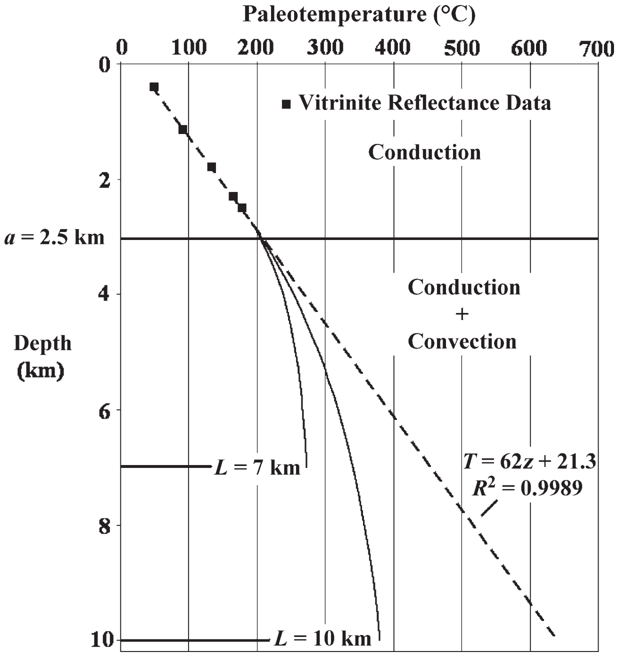

Figure 7. Hydrothermal event temperature model with conductive layer (depth $z = 0$–2.5 km [0–8202 ft]) and convection plus conduction ($z = a = 2.5$ km [8202 ft] to $z = L$). Fluid flow v_z originates from $z = L$. The temperature-depth data (squares) are taken directly from the Figure 6 BasinMod™ results. The dashed regression line to the vitrinite data (squares) approximates the Figure 6 conduction-only BasinMod™ temperatures (Poprawa and Grotek, 2004), with $T > 400$°C for $z > 6$ km (13,123 ft). Solid curves are for $L = 7$, 8, and 10 km (13,123, 16,404, and 19,685 ft, respectively). L = depth of origin of flow; T = temperature; R^2 = coefficient of determination.

$\exp(-\frac{\kappa t}{4r^2}) = \frac{1}{2}$, where κ is thermal diffusivity (10^{-6} m^2/s) and t is the observed (Kozlowska and Poprawa, 2004; Poprawa and Grotek, 2004) 10-m.y. decay time for the thermal event (Figure 6). This results in a value for the radius, $r = 10$ km (6 mi) (the radius of circles in Figure 2), which is consistent with the absence of the anomaly at Krosniewice if the anomaly is centered at Kutno. The model supports a thermal event that is localized at Kutno and Korabiewice, perhaps along preexisting northwest-southeast trending faults, because broader events would take longer to decay.

At the height of the postulated Early to Middle Jurassic anomaly, the deepest vitrinite samples, now at 3.6-km (11,811-ft) depth, indicate a paleotemperature near 180°C at 2.5 km (8202 ft). Below this depth, temperatures are not constrained by vitrinite data and all predictions are model dependent. The linearity of the present-day Kutno geotherm (Figure 4) gives low probability to significant vertical fluid flow. However, to place realistic limits on the temperature effects of the postulated hydrothermal event (Poprawa and Grotek, 2004), we consider heat transport by convection resulting from vertical fluid flow from depth.

Our model (Figure 7) consists of a conductive layer constrained by the vitrinite data of Poprawa and Grotek (2004) and extends to depth $a = 2.5$ km (8202 ft). Within the conductive layer, the thermal gradient $G = 62$°C/km and the temperature T at 2.5 km (8202 ft) ($T_a = 177$°C) are both well constrained by the vitrinite reflectance data results from BasinMod™ (squares, Figure 7), which are based on Sweeney and Burnham (1990). The resulting geotherm is indicated by the regression line (dashed), where $T = 62z + 21.3$ ($R^2 = 0.9989$). For the Kutno thermal conductivity ($k = 2.29$ W/m K), this yields a heat flow of 142 mW/m^2, nearly 10% more than the 130 mW/m^2 reported by Poprawa and Grotek (2004).

The regression to the vitrinite data also yields a surface temperature $T_o = 21.3$°C for the Jurassic (~185 Ma). This is 27.5°C higher than the –6.2°C temperature derived previously for the Weichselian glaciation from the present-day thermal data for the Kutno-1 well (dashed line, Figure 4) for depth = 0. The result appears to be in good agreement with evidence from reef-building corals found as far as 60°N paleolatitude, which suggest a minimum water temperature of approximately 20°C, and evidence from abundant Jurassic ferns, whose present-day counterparts cannot tolerate frost (Hallam, 1982). It is also consistent with the distribution of evaporites, carbonates, and coals (Hallam, 1982). Present-day mean temperatures approximately 21.3°C (Kelsey, 1984) are found in such places as La Palmas, Canary Island (20°C); Touggourt and Tamanrasset, Algeria (21.6°C and 21.1°C, respectively); Entebbe, Uganda (21.1°C); Peshawar, Pakistan (21.6°C); Calcutta, India (20°C); and Port-Au-Prince, Haiti (21.6°C). Kelsey (1984) cites the present-day mean temperature for Warsaw, Poland at 7.8°C.

Below the conductive layer of our model (Figure 7), we add to our model heat transport by convection from a hydrothermal event originating at depth $z = L$ and continuing to the base of the conductive layer ($z = a$). The geotherm in this zone is governed by Bredehoeft and Papadopulos (1965), where

$$\frac{T - T_a}{T_L - T_a} = \frac{e^{\beta\frac{(z-a)}{L-a}} - 1}{e^{\beta} - 1} (T_a = 177°\text{C at } z = a = 2.5 \text{ km}),\ \beta = \frac{\rho_w c_w v_z (L - a)}{k} \quad (7)$$

and ρ_w, c_w, v_z, and k are the density, heat capacity, vertical flow velocity of water, and thermal conductivity, respectively (T_a and T_L are the temperatures at depths a and L; k = 2.29 W/m K; $\rho_w c_w$ = 4.186 × 10^6 J/m^3 K; 1W = 1J/s). Steady-state heat flow for our composite model (Figure 7) requires continuity of heat flux at $z = a$, where $\frac{T_L - T_a}{L-a}\frac{\beta}{e^{\beta}-1} = G$ (62°C/km). We solve these equations for L = 7 km (22,966 ft) and L = 10 km (32,808 ft). For T_L at each depth, we assume the present-day thermal gradient (G = 36°C/km) from the Kutno data (Figure 2), but with Jurassic surface temperature T_o = 21.3°C from the vitrinite data, extrapolated to 7-km (22,966-ft) and 10-km (32,808-ft) depths (273°C and 381°C, respectively). These represent normal Jurassic temperatures upon which the subject hydrothermal event was superimposed. The corresponding upward flow velocities required to maintain steady-state conditions (3.3 × 10^{-10} and 1.4 × 10^{-10} m/s [10.8 × 10^{-10} and 4.6 × 33^{-10} ft/s, respectively]) decrease with increasing depth (L) and are consistent with values obtained by studies of fluid flow in sedimentary basins (Majorowicz and Jessop, 1981a, b; Zielinski and Bruchhausen, 1983; Davis et al., 1990; Ritter et al., 2004; Zielinski et al., 2007) as are values of L. The temperature results for the zone of convection plus conduction are given in Figure 7 (solid curves). With increasing L and decreasing flow velocity (v_z), the solid curves are seen to asymptotically approach the purely conductive geotherm (dashed line) and the temperature reduction via convection becomes negligible much below L = 10 km.

It is immediately evident (Figure 7) that temperatures with convection (solid curves) are considerably less than those predicted by simple extrapolation of temperature from the conductive zone (dashed). The latter abruptly results in temperatures near 400°C at relatively shallow depths (6 km [19,685 ft]), consistent with the BasinMod™ results (Figure 6) and for which no geologic evidence exists anywhere in our study area. Flow originating from 10-km (32,808-ft) depth is extreme compared with other areas (Majorowicz and Jessop, 1981a, b; Zielinski and Bruchhausen, 1983; Davis et al., 1990; Ritter et al., 2004; Zielinski et al., 2007), probably because of the lack of water sources at greater depths. Flow from more likely shallower depths (<7 km [22,966 ft]) results in temperatures well below 300°C (Figure 7). Zielinski et al. (2007) present evidence for an offshore hydrocarbon seep with heat flow greater than 600 mW/m^2 believed to be caused by fluid advection from 6-km (19,685-ft) depth. Our model essentially replaces 1-D conductive heat flow with a 1-D two-layer model (Figure 7) that accounts for convection using Bredehoeft and Papadopulos (1965). The latter provides a relatively simple next level of complexity when fluid flow must be considered and purely conductive models prove inadequate. It provides a first-order parameterization of fluid flow in terms of vertical velocity (v_z) and depth of its origin (L) even where patterns of flow and associated mathematics are, in reality, more complex and/or poorly constrained.

Some evidence in support of the lower temperatures implied by the Figure 7 convective model are shown in Figure 8, where the vitrinite reflectance (R_o) data for Kutno-1 and Krosniewice IG-1, with and without thermal events, respectively (Poprawa and Grotek, 2004), are plotted along with the data from Byczyna-1, approximately 70-km (229,659-ft) northwest (Figure 2). Like Kutno, the latter exhibit a marked increase in R_o below 3-km (9843-ft) depth. The top of the Jurassic at Byczyna is at 1528 m (949 ft) and at 1554 m (966 ft) for Kutno, whereas the Jurassic–Triassic boundary is approximately 500 m (1640 ft) deeper at Kutno. Present-day heat flow is 5% greater at Kutno. The combined data from Kutno (squares) and Byczyna (diamonds) appear in Figure 8 as if from a single well. Within margins of error, the hydrothermal events appear approximately simultaneous and of similar magnitude (~130–150 mW/m^2).

For Korabiewice PIG-1 (Figure 2) and southwest of our study area, fluid inclusion data (from

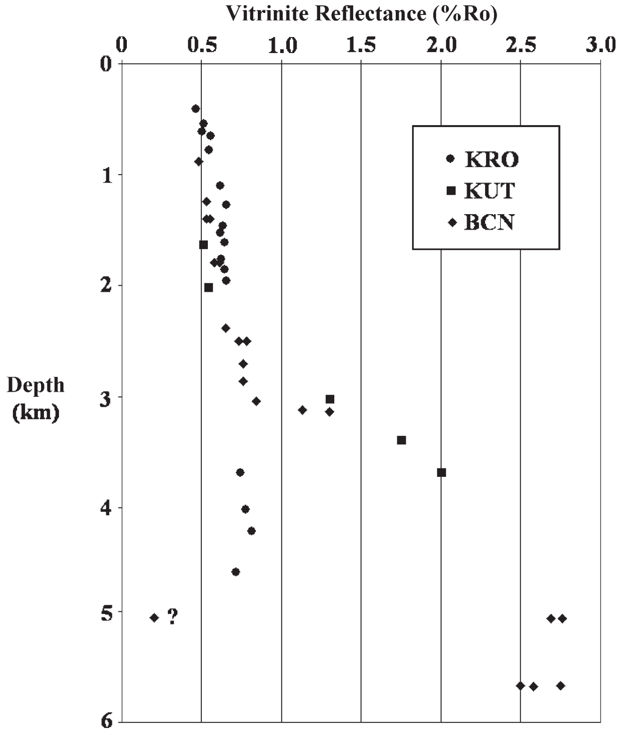

Figure 8. Vitrinite reflectance data for Kutno-1 (KUT) and Krosniewice IG-1 (KRO), with and without hydrothermal events (Poprawa and Grotek, 2004), with data from Byczyna-1 (BCN) that also exhibit an event.

quartz rims and Fe-dolomite and ankerite cements) and K/Ar dating (Kozlowska and Poprawa, 2004) yield maximum paleotemperatures mostly less than 200°C corroborating the vitrinite data. Fluid inclusion data (from quartz and diagenetic carbonates) yield similar results for the Lublin Basin of southeastern Poland (Poprawa and Zywiecki, 2005) except for paleotemperatures associated with igneous intrusion, which are still less than 400°C based on computed formation-water salinities. The impact of hot fluids associated with Permian volcanics on Carboniferous sediment maturation (R_o) in the Variscan orogen (southwestern Poland) is seen in fluid-inclusion homogenization temperatures for calcite and anhydrite that are also less than 400°C (Poprawa et al., 2005) based on derived formation-water salinities. Furthermore, even for heating rates appropriate to geothermal systems, Sweeney and Burnham (1990) predict maximum paleotemperatures of less than 300°C, based on their own kinetic calculations and those of Barker (1989), for the maximum R_o = 2.75% (Figure 8) observed at Byczyna. This and the fluid-inclusion thermometry are in good agreement with paleotemperatures predicted by our Figure 7 convective model as opposed to the Figure 6 conductive model results from BasinMod™.

The Figure 7 hydrothermal event model accounts for an increase in heat flow from present day (82.4 mW/m^2) to 142 mW/m^2 in the Jurassic based on vitrinite data for Kutno. The Jurassic paleoheat flow (~80 mW/m^2) caused by the corresponding thermal event at Korabiewice (Kozlowska and Poprawa, 2004) is significantly less than for Kutno. However, Korabiewice resides in a zone of significantly lower (59.5 mW/m^2) present-day heat flow (Figure 1). In both cases, the increases in heat flow (Q), $\frac{Q_J}{Q} = \frac{\beta}{e^{\beta}-1}$, results in β (Kutno) / β (Korabiewice) ≅ 1.9. The harmonic mean thermal conductivity for Korabiewice (k = 2.89 W/m K) is 26% higher than for Kutno (2.29 W/m K). From the equation for β above (7), this implies Q (Kutno) /Q (Korabiewice) ≅ 1.5, where $Q = v_z (L - a)$. Assuming the same depth of origin ($L - a$) for both, v_z (Kutno) must be 50% greater than for Korabiewice, or for equal flow velocity, the depth of origin of flow ($L - a$) for Kutno must be 50% greater than for Korabiewice, or a combination of both may exist. Using the present-day heat flow from the entire well (85.0 mW/m^2), that percentage reduces to 40%.

ON METHANE PRESERVATION

Hunt (1979) cites the phase out of gas in source rocks recognized from its disappearance in cuttings and from pyrolysis (e.g., Littke et al., 1995), whereas methane preservation in reservoir rocks is more uncertain and may be more related to physical properties like porosity because methane is thermally indestructible at sedimentary rock temperatures. However, thermodynamic modeling results of Barker and Takach (1992) indicate a wide range of gas composition possible in sandstone reservoirs depending upon chemistry as well as thermal history.

Barker and Takach (1992) can be used to predict the gas composition for 6-km (19,685-ft) depth for the Kujawy area for an upper limit extrapolated temperature of 210°C (Figure 4). This temperature is approximated at 7620-m (25,000-ft) depth, for the Barker and Takach (1992) model, where methane concentrations near 60% are predicted for most systems capable of emulating the observed 81% methane in considerably shallower (~3 km [9843 ft]) Rotliegende reservoirs of the Groningen gas field, Netherlands (Stäuble and Millius, 1970). Pressure has only minor influence; however, the strong dependency of gas composition on the mineral components of reservoirs is evident.

Even for systems involving calcite and sulfur, the presence of small amounts of residual graphite (C), as from cracking of crude oil, assures the presence of methane. If sulfur plus graphite are involved, methane reduces to 25%. Low-methane, high-CO_2 Rotliegende reservoirs in northern German basins can result from cross-formation fluid flow associated with extensive block faulting (Gaup et al., 1993), creating calcite and sulfur systems and thermochemical sulfate reduction (Sassen and Moore, 1988; Steinhoff et al., 1997; Steinhoff and Strohmenger, 1998) associated with hydrocarbon migration through Zechstein carbonates.

High heat flow and temperatures greater than 300°C (Schoell and Cathles, 2000; Schoell et al., 2000) result in reservoirs containing low methane and high CO_2, some resulting from inorganic high-temperature mineral reactions. Furthermore, extensive Early Permian volcanism in the northeast German Basin (Kossow et al., 1999) may have subjected Carboniferous source rocks to very high temperatures. Temperatures greater than 300°C are associated with nitrogen-rich German Rotliegende reservoirs (Littke et al., 1995), which may also be associated with Permian volcanism. They report nitrogen-rich Rotliegende reservoirs in the North Sea and 15% nitrogen in the Rotliegende at Groningen (Netherlands). For Kujawy, even the extrapolated Kutno-1 geotherm (dashed line, Figure 4) predicts present-day temperatures less than 300°C at 8 km (26,247 ft) (Figure 5) with potential Carboniferous source rocks as shallow as 7 km (22,966 ft) (Figure 3).

The nitrogen found in deep Rotliegende reservoirs in northern Germany (Littke et al., 1995) is most likely generated during the Tertiary from highly mature Carboniferous (Westphalian) coals. These have most likely been subjected to temperatures greater than 300°C for isothermal conditions lasting on the order of 300 m.y., as opposed to the approximately 185-Ma Jurassic thermal event lasting less than 50 m.y. in Kujawy. These coals are widely overlain by up to 600 m (1969 ft) of Lower Permian volcanoclastic rocks, the result of intense but relatively short-lived volcanism, where rocks may have been exposed to temperatures greater than 500°C.

Hence, the duration of maximum reservoir and source rock paleotemperature (<50 m.y., 273°C) inferred from Figures 6 and 7 is contrary to the kinetics of nitrogen and CO_2-producing wells, that is, temperatures greater than 300°C for isothermal conditions lasting on the order of 300 m.y.

CONCLUSIONS

An extremely linear present-day thermal profile (R^2 = 0.9996) is observed from 3000- to 5276-m (9843–17,310-ft) depth in the Kujawy area of the central Polish Basin. The temperatures represent equilibrium steady-state conditions, not influenced by transient surface temperature or drilling history. The linearity also precludes significant active vertical fluid flow within the interval.

Thermal data from deep wells (Figure 4) predict the presence of methane-dominated (~60%) gas in sandstone systems (Barker and Takach, 1992) that approximate potential Rotliegende reservoirs. Uncertainty in this prediction is likely to be resolved by new deep drilling, slated for the area. The high thermal conductivity and thickness of Zechstein cap rock reduce temperatures in Rotliegende reservoirs and Carboniferous source rocks. This is similar to the effect reported for salt bodies in the northwest German Basin (Schwarzer and Littke, 2007). Models (2-D and 3-D) predict that salt pillows in the area may negate this effect; however, evidence for this is absent in the deep wells studied, where temperature changes (ΔT_z = ~–10°C) appear typical for the approximately 500-m (1640-ft) thickness of Zechstein. The magnitude of this effect translates to 50% reduction in reaction rates according to the Lopatin (1971) and Waples (1980) TTI model, effectively protecting underlying deep methane reservoirs from thermal cracking. A simple relation $\Delta T_z = -Q\ (k^{-1} - k_z^{-1})\ \Delta z$, where Q is heat flow, k_z and Δz are the thermal conductivity and thickness (meters) of Zechstein, and k is the thermal conductivity of overlying rocks, allows estimation of the effect elsewhere from heat flow and lithostratigrapy obtained from deep wells such as in the PGI database (e.g., Figure 1; Szewczyk and Gientka, 2009). For Kujawy, $\Delta T_z = 19.1\ \Delta z$.

Because of lower geothermal gradients associated with convection, the Early to Middle Jurassic hydrothermal thermal event (Poprawa and Grotek, 2004) is not expected to significantly alter the present-day gas composition in potential Kujawy area reservoirs. Contrary to purely conductive models that predict 455°C paleotemperature at 7-km (22,966-ft) depth (Carboniferous), the convective model predicts 273°C despite a paleoheat flow of 142 mW/m^2, with an upward fluid flow of 3.3×10^{-10} m/s (10.8×10^{-10} ft/s).

Furthermore, nitrogen kinetics inferred from wells in northern Germany support generation at temperatures greater than 300°C over long (~300 m.y.) periods, instead of during short-term (~50 m.y.) thermal events, unless magma bodies are involved. Similar kinetics are likely to apply to CO_2 generation. The duration of maximum reservoir and source rock paleotemperature (<50 m.y.) is contrary to the kinetics of nitrogen- and CO_2-producing wells (i.e., temperatures >300°C for isothermal conditions lasting on the order of 300 m.y.).

The approximately 185-Ma hydrothermal event is seen in three widespread sites trending subparallel to major northwest-southeast–trending structure, faulting, and regional heat flow (Figures 1, 2). Other sites as close as 14 km (9 mi) to hydrothermal event sites are without hydrothermal events. This, along with the 10-m.y. decay of the anomalous heat flow of the event (based on vitrinite data), suggests localized fluid movement along structural elements in existence since the Late Triassic, with thermal effects extending no more than approximately 10 km (6 mi) laterally from their axes. This is consistent with localized cylindrical plumes of 10-km (6-mi) radius that cool by conduction. One such site (Kutno-1; Figure 2) exhibits more than 80 mW/m^2 present-day heat flow and experienced more than 140 mW/m^2 paleoheat flow. The Kutno hydrothermal event is characterized by a 40 to 50% higher fluid-flow velocity or 40 to 50% greater depth of origin of flow (or a combination of the two) than at Korabiewice (Figure 2), where present-day heat flow just less than 60 mW/m^2 with 80 mW/m^2 paleoheat flow. This suggests possible structural control on heat flow persisting 185 m.y., where the trend of intensity of the 185-Ma hydrothermal event correlates with the trend of increasing present-day heat flow (Figure 1). Kujawy-area deep-well data provide reliable estimates of surface temperatures during the last (Weichselian) glaciation (–6.2°C) and in the approximately 185-Ma Jurassic (21.3°C).

REFERENCES CITED

Andrews-Speed, C. P., E. R. Oxburgh, and B. A. Cooper, 1984, Temperatures and depth-dependent heat flow in the North Sea: AAPG Bulletin, v. 68, p. 1764–1781.

Barker, C., and N. E. Takach, 1992, Prediction of natural gas composition in ultradeep sandstone reservoirs: AAPG Bulletin, v. 76, p. 1859–1873.

Barker, C. E., 1989, Temperature and time in the thermal maturation of organic matter, *in* N. D. Naeser and T. H. McCulloh, eds., Thermal history of sedimentary basins: Methods and case histories: New York, Springer-Verlag, p. 217–238.

Beck, A. E., 1965, Techniques of measuring heat flow on land, *in* W. Lee, ed., Terrestrial heat flow: American Geophysical Union Monograph Series 8, p. 24–57.

Bredehoeft, J. D., and I. S. Papadopulos, 1965, Rates of vertical groundwater movement estimated from the earth's thermal profile: Water Resources Research, v. 1, p. 325–328, doi:10.1029/WR001i002p00325.

Brigaud, E. C., D. S. Chapman, and S. LeDouran, 1990, Estimating thermal conductivity in sedimentary basins using lithological data and geophysical well logs: AAPG Bulletin, v. 74, p. 1459–1477.

Carslaw, H. S., and J. C. Jaeger, 1959, Conduction of heat in solids: Oxford, United Kingdom, Clarendon Press, 510 p.

Dadlez, R., 1989, Epicontinental Permian and Mesozoic basins in Poland: Geological Quarterly, v. 33, p. 175–198.

Dadlez, R., 1997, Epicontinental basins in Poland: Devonian to Cretaceous: Relationship between the crystalline basement and sedimentary infill: Geological Quarterly, v. 41, p. 419–432.

Dadlez, R., A. Iwanow, K. Leszczyñski, and S. Marek, 1998, Mapa tektoniczna kompleksu cechsztyñsko-mezozoicznego na Ni¿u Polskim: Warsaw, Poland, Polish Geological Institute, scale 1:500,000, 1 sheet.

Davis, E. E., R. D. Hyndman, and H. Villinger, 1990, Rates of fluid expulsion across the northern Cascadia accretionary prism: Constraints from new heat flow and multichannel seismic data: Journal of Geophysical Research, v. 95, p. 8869–8889, doi:10.1029/JB095iB06p08869.

Gaup, R., A. Matter, J. Platt, K. Ramseyer, and J. Walzebuck, 1993, Diagenesis and fluid evolution of deeply buried Permian (Rotliegende) gas reservoirs, northwest Germany: AAPG Bulletin, v. 77, p. 1111–1128.

Gautier, D. L., 2003, Carboniferous-Rotliegend total petroleum system description and assessment results summary: U.S. Geological Survey Bulletin 2211, 29 p.

Hallam, A., 1982, The Jurassic climate: Climate in Earth history—Studies in geophysics: Washington, DC, National Academies Press, p. 159–163.

Hunt, J. M., 1979, Petroleum geochemistry and geology: San Francisco, W. H. Freeman and Company, 617 p.

Kappelmeyer, O., and R. Haenel, 1974, Geothermics with special reference to application: Berlin, Gebrüder Borntraeger, Geoexploration Monographs Series 1, 238 p.

Kelsey, M.R., 1984, Climbers and hikers guide to the world's mountains, 2d ed.: Springville, Utah, Kelsey Publishing Company, p. 782–795.

Kiersnowski, H., 1997, Depositional development of the Polish upper Rotliegend Basin and evolution of its sediment source areas: Geological Quarterly, v. 41, p. 433–456.

Kiersnowski, H., A. Buniak, M. Kuberska, and A. Srokowska-Okoñska, 2010, Tight-gas accumulations in Rotliegend sandstones of Poland (in Polish with English summary): Przeglad Geologiczny, v. 58, p. 335–346.

Kossow, D., C. M Krawczyk, T. McCann, J. F. W. Negendank, and M. Strecker, 1999, The Permo-Triassic evolution of the northern margin of the NE German basin: Constraints from cross section balancing (abs.): AAPG Annual Meeting, San Antonio, Texas, April 11–14, 1999, http://www.searchanddiscovery.com/abstracts/html/1999/annual/abstracts/312.htm?q=%2BtextStrip%3Akossow (accessed July 2012).

Kozlowska, A., and P. Poprawa, 2004, Diagenesis of the Carboniferous clastic sediments of the Mazowsze region and the northern Lublin region related to their burial and thermal history (in Polish with English summary and figure captions): Przeglad Geologiczny, v. 52, p. 491–500.

Lachenbruch, A. H., and M. C. Brewer, 1959, Dissipation of the temperature effect in drilling a well in Arctic Alaska: U.S. Geological Survey Bulletin 1083-C, p. 73–109.

Lachenbruch, A. H., and B. V. Marshall, 1966, Heat flow through the Arctic Ocean floor: The Canada Basin–Alpha Rise boundary: Journal of Geophysical Research, v. 71, p. 1223–1248, doi:10.1029/JZ071i004p01223.

Lee, T.-C., and T. L. Henyey, 1974, Heat-flow refraction across dissimilar media: Geophysical Journal of the Royal Astronomical Society, v. 39, p. 319–333, doi:10.1111/j.1365-246X.1974.tb05457.x.

Leszczyñski, K., 2008, Budziszewice IG-1: Polish Geological Institute deep wells profiles (in Polish with English descriptions): Warsaw, Poland, Polish Geological Institute, v. 127, 149 p.

Littke, R., B. Krooss, E. Idiz, and J. Frielingsdorf, 1995, Molecular nitrogen in natural gas accumulations: Generation from sedimentary organic matter at high temperatures: AAPG Bulletin, v. 79, p. 410–430.

Lopatin, N. V., 1971, Temperature and geologic time as factors in coalification (in Russian): Izvestiya Akademii Nauk SSSR, Seriya Geologicheskaya, no. 3, p. 95–106.

MacMillan, W. D., 1930, Theoretical mechanics: The theory of the potential: McGraw-Hill Book Company, reprinted 1958, New York, Dover Publications, 469 p.

Majorowicz, J. A., and A. M. Jessop, 1981a, Regional heat-flow patterns in the Western Canadian sedimentary basin: Tectonophysics, v. 74, p. 209–238.

Majorowicz, J. A., and A. M. Jessop, 1981b, Present heat flow and a preliminary paleogeothermal history of the central Prairies Basin, Canada: Geothermics, v. 10, p. 81–93, doi:10.1016/0375-6505(81)90014-6.

Majorowicz, J., and S. Plewa, 1979, Study of heat flow in Poland with special regard to tectonophysical problems, *in* V. Cermak and L. Rybach, eds., Terrestrial heat flow in Europe: Inter-Union Commission of Geodynamics Scientific Report 58, p. 240–252.

Mazur, S., P. Aleksandrowski, K. Turniak, L. Krzeminski, K. Mastalerz, A. Gorecka-Nowak, L. Kurowski, P. Krzywiec, A. Zelazniewicz, and M. C. Fanning, 2010, Uplift and late orogenic deformation of the central European Variscan belt as revealed by sediment provenance and structural record in the Carboniferous foreland basin of western Poland: International Journal of Earth Sciences (Geologische Rundschau), v. 99, p. 47–64, doi:10.1007/s00531-008-0367-3.

Mundry, E., 1968, Über die abkülung magmatische körper: Geologisches Jahrbuch, v. 85, p. 755–766.

Narkiewicz, M., 2007, Development and inversion of Devonian and Carboniferous basins in the eastern part of the Variscan foreland (Poland): Geological Quarterly, v. 51, no. 3, p. 231–256.

Norden, B., and A. Förster, 2006, Thermal conductivity and radiogenic heat production of sedimentary and magmatic rocks in the northeast German Basin: AAPG Bulletin, v. 90, p. 939–962, doi:10.1306/01250605100.

Poprawa, P., and I. Grotek, 2004, Thermal evolution of the Permian–Mesozoic Polish Basin: Model predictions confronted with analytical data: Bollettino di Geofisica, Teorica ed Applicata, v. 45, p. 259–261.

Poprawa, P., and M. M. Zywiecki, 2005, Heat transfer during development of the Lublin Basin (SE Poland): Preliminary results of maturity modeling and fluid inclusion analysis: Mineralogical Society of Poland—Special Papers, v. 26, p. 255–261.

Poprawa, P., I. Grotek, M. Wagner, and H Matyja, 2002, Phanerozoic thermal history of the Trans-European suture zone in Poland: Recent stage of research in the project PAP (in Polish with English summary and figure captions): Przeglad Geologiczny, v. 50, p. 1219–1220 (in Polish).

Poprawa, P., I. Grotek, and M. M. Zywiecki, 2005, Impact of the Permian magmatic activity on the thermal maturation of the Carboniferous sediments in the outer Variscan orogen (SW Poland): Mineralogical Society of Poland—Special Papers, v. 26, p. 241–250.

Ritter, U., G. W. Zielinski, H. M. Weiss, R. L. B. Zielinski, and J. Saettem, 2004, Heat flow in the Vøring Basin, mid-Norwegian Shelf: Petroleum Geoscience, v. 10, p. 353–365, doi:10.1144/1354-079303-616.

Safanda, J., J. Szewczyk, and J. Majorowicz, 2004, Geothermal evidence for very low glacial temperatures on a rim of the Fennoscandian ice sheet: Geophysical Research Letters, v. 31, p. 1–4, doi:10.1029/2004GL019547.

Sassen, R., and C. H. Moore, 1988, Framework of hydrocarbon generation and destruction in eastern Smackover trend: AAPG Bulletin, v. 72, p. 649–663.

Schoell, M., and L. M. Cathles, 2000, Formation of CO_2 in sedimentary basins and assessment of CO_2 risk in gas prospects (abs.): AAPG International Conference and Exhibition, Bali, Indonesia, October 15–18, 2000: http://www.searchanddiscovery.com/abstracts/html/2000/intl/abstracts/362.htm (accessed May 2012).

Schoell, M., Y. Tang, R. J. Hwang, D. K. Baskin, N. B. Schoellkopf, and L. M. Cathles, 2000, Formation and occurrence of hydrocarbons and nonhydrocarbon gases in the Yinggehai and Qiongdongnan basins/South China Sea (abs.): AAPG International Conference and Exhibition, Bali, Indonesia, October 15–18, 2000: http://www.searchanddiscovery.com/abstracts/html/2000/intl/abstracts/362.htm (accessed May 2012).

Schwarzer, D., and R. Littke, 2007, Petroleum generation and migration in the "tight gas" area of the German Rotliegende natural gas play: A basin modeling study: Petroleum Geoscience, v. 13, p. 37–62, doi:10.1144/1354-079306-703.

Stäuble, A. J., and G. Millius, 1970, Geology of Groningen gas field, Netherlands, *in* M. T. Halbouty ed., Geology of giant petroleum fields: AAPG Memoir 14, p. 359–369.

Steinhoff, I., and C. Strohmenger, 1998, Facies analysis and sequence stratigraphy of the Werra anhydrite (Upper Permian of Germany) (abs.): AAPG Annual Convention, Salt Lake City, Utah, May 17–28, 1998, 4 p, http://www.searchanddiscovery.com/abstracts/html/1998/annual/abstracts/633.htm (accessed May 2012).

Steinhoff, I., N. E. Pingitore Jr., C. Strohmenger, and K. Rockenbauch, 1997, Sour-gas generation during hydrocarbon migration through Zechstein evaporites (Upper Permian of Germany) (abs.): AAPG Annual Meeting, Dallas, Texas, April 6–9,1997, 1 p

Sweeney, J. J., and A. K. Burnham, 1990, Evaluation of a simple model of vitrinite reflectance based on chemical kinetics: AAPG Bulletin, v. 74, p. 1559–1570.

Szewczyk, J., 2001, Estimation of the heat-flow density using thermal parameter modeling (in Polish with English summary and figure captions): Przeglad Geologiczny, v. 49, p. 1083–1088.

Szewczyk, J., 2005, Climate changes and their influence on subsurface temperature of the Earth (in Polish with English summary and figure captions): Przeglad Geologiczny, v. 53, p. 77–86.

Szewczyk, J., and D. Gientka, 2009, Terrestrial heat-flow density in Poland: A new approach: Geological Quarterly, v. 53, no. 1, p. 125–140.

Tissot, B., and J. Espitalie, 1975, L'evolution thermique de la matierer organique des sediments: Applications d'une simulation mathematique: Revues l'Institut Francais du Petrole, v. 30, p. 743–777.

Tissot, B., and D. Welte, 1978, Petroleum formation and occurrence: New York, Springer-Verlag, 538 p.

van Wees, J.-D., R. A. Stephenson, P. A. Ziegler, U. Bayer, T. McCann, R. Dadlez, R. Gaupp, M. Narkiewicz, F. Bitzer, and M. Scheck, 2000, On the origin of the southern Permian Basin, central Europe: Marine and Petroleum Geology, v. 17, no. 1, p. 43–59, doi:10.1016/S0264-8172(99)00052-5.

von Herzen, R. P., and S. Uyeda, 1963, Heat flow through the Pacific Ocean floor: Journal of Geophysical Research, v. 68, p. 4219–4250.

Wagner, R., 1994, Stratigraphy and evolution of the Zechstein Basin in the Polish Lowlands (in Polish with English summary): Prace Państwowego Instytutu Geologicznego, v. 146, p. 1–71.

Waksmundzka, M. I., 2010, Pre-Permian subcrop map of Poland, *in* H. Doornenbal and A. Stevenson, eds., Petroleum geological atlas of the southern Permian Basin area: Houten, Netherlands, European Association of Geoscientists and Engineers Publications, 1 p.

Waples, D. W., 1980, Time and temperature in petroleum exploration: AAPG Bulletin, v. 64, p. 916–926.

Ziegler, P., 1990, Geological atlas of western and central Europe, 2d ed.: The Hague, Netherlands, Shell Internationale Petroleum Maatschappij BV, 239 p.

Zielinski, G. W., and P. M. Bruchhausen, 1983, Shallow temperatures and thermal regime in the hydrocarbon province of Tierra del Fuego: AAPG Bulletin, v. 67, p. 166–177.

Zielinski, G. W., M. Bjorøy, R. L. B. Zielinski, and I. L. Ferriday, 2007, Heat flow and surface hydrocarbons on the Brunei continental margin: AAPG Bulletin, v. 91, p. 1053–1080, doi:10.1306/01150706117.