Geo-Mar Lett (2016) 36:25–33
DOI 10.1007/s00367-015-0425-3

ORIGINAL

Conductive heat flow and nonlinear geothermal gradients in marine sediments—observations from Ocean Drilling Program boreholes

Christian Stranne[1] **· Matt O'Regan**[1]

Received: 15 July 2015 / Accepted: 28 September 2015 / Published online: 5 October 2015
© Springer-Verlag Berlin Heidelberg 2015

Abstract A basic premise in marine heat flow studies is that the temperature gradient varies with depth as a function of the bulk thermal conductivity of the sediments. As sediments become more deeply buried, compaction reduces the porosity and causes an increase in the bulk thermal conductivity. Therefore, while the heat flow may remain constant with depth, the thermal gradient is not necessarily linear. However, it has been argued that measurements showing increased sediment thermal conductivity with burial depth may be caused by a horizontal measurement bias generated by increasing anisotropy in sediments during consolidation. This study reanalyses a synthesis of Ocean Drilling Program data from 186 boreholes, and investigates the occurrence of nonlinear geothermal gradients in marine sediments. The aim is to identify whether observed downhole changes in thermal conductivity influence the measured temperature gradient, and to investigate potential errors in the prediction of in-situ temperatures derived from the extrapolation of near-surface thermal gradients. The results indicate that the measured thermal conductivity does influence the geothermal gradient. Furthermore, comparisons between shallow measurements (<10 m) from surface heat flow surveys and the deeply constrained temperature data from 98 ODP boreholes indicate that the shallow gradients are consistently higher by on average 19 °C km^{-1}. This is consistent with higher porosity and generally lower thermal conductivity in near-seafloor sediments, and highlights the need to develop robust porosity–thermal conductivity models to accurately predict temperatures at depth from shallow heat flow surveys.

Electronic supplementary material The online version of this article (doi:10.1007/s00367-015-0425-3) contains supplementary material, which is available to authorized users.

✉ Christian Stranne
christian.stranne@geo.su.se

[1] Department of Geological Sciences and Bolin Centre for Climate Research, Stockholm University, 10691 Stockholm, Sweden

Introduction

Geothermal heat flow is a critical parameter for studying the evolution of continental and oceanic lithosphere (Chapman 1986; Artemieva and Mooney 2001; Grose 2012; Hasterok 2013; Davies 2013), investigating fluid flow patterns (Kukkonen 1988; Stein and Abbott 1991), diagenetic alteration in sediments (Moore and Saffer 2001), petroleum generation (Tissot et al. 1987), and gas hydrate dynamics (Grevemeyer and Villinger 2001; Wallmann et al. 2012; Villar-Muñoz et al. 2013). Insights into lithospheric and crustal processes are often obtained by assessing crustal heat flow patterns, while for many sedimentary processes it is the geothermal gradient that is of paramount interest. In this paper two key questions related to marine heat flow, and how in-situ temperature estimates are derived from shallow penetrating heat flow surveys, are investigated: (1) are geothermal gradients in consolidating marine sediments linear or nonlinear? (2) How do linear extrapolations from shallow measurements compare to measured temperatures at depth?

;Under purely conductive heat transfer, the three basic parameters governing the geothermal regime are the heat flux from the underlying crust, the temperature at the sediment–water interface, and the thermal conductivity of the sediments. Under one-dimensional steady-state conductive heat flow (with no significant internal heat sources or sinks), the heat flow (q, W m^{-2}) remains constant with depth, and is given by a version of Fourier's law,

$$q = \lambda(z)\frac{\mathrm{d}T(z)}{\mathrm{d}z} \quad (1)$$

where λ is the bulk thermal conductivity (W m^{-1} K^{-1}), z the depth below seafloor (m) and T the temperature (K).

Large variations in λ with depth can occur across major lithologic boundaries, unconformities and, more gradually, through the reduction in porosity with burial depth (compaction). To account for these variations, heat flow is estimated using the Bullard method (Bullard 1939; Powell et al. 1988), whereby downhole variations in λ are accounted for by defining the cumulative thermal resistance (Ω, (m^2 K) W^{-1})) with depth,

$$\Omega(z) = \int_0^z \mathrm{d}z/\lambda(z) \quad (2)$$

so that

$$q = \frac{\mathrm{d}T(z)}{\mathrm{d}\Omega(z)} \quad (3)$$

The Bullard method predicts that, in a purely conductive system, the thermal gradient is a linear function of the thermal resistance, and not necessarily of depth below the seafloor. Nonlinearity in the geothermal gradient is expected when there are variations in the thermal conductivity profile within the sediments. If a reliable thermal conductivity model for the sediments is available, then the Bullard method provides a way to use the measured surface heat flow to predict the in-situ temperatures at greater depths (Chapman et al. 1984; Pribnow et al. 2000; O'Regan and Moran 2010).

Grevemeyer and Villinger (2001) argue that measured downhole increases in sediment thermal conductivity may be an artefact caused by a horizontal bias in the measurement of increasingly anisotropic sediments during consolidation. They conclude that a linear extrapolation of near-surface temperature gradients provides a reasonable and fairly robust method for estimating temperatures at depths between 100 and 500 meters below the seafloor (mbsf). This presents a major conceptual and numerical problem when modelling in-situ temperatures using shallow heat flow measurements (Wallmann et al. 2012), as empirical formulations relating thermal conductivity with depth are grounded in the widely observed reduction in porosity (ϕ) that occurs during sediment burial (Athy 1930; Mondol et al. 2007). Most compaction models define porosity loss as either an exponential function of burial depth (e.g. Athy 1930) or a logarithmic function of effective stress (Terzaghi 1943; Gibson 1958). These models both imply a strong reduction in porosity in the upper few hundred meters below the seafloor.

Conductivity–depth models are commonly based on either an arithmetic mixing law,

$$\lambda = \lambda_\mathrm{w}\phi + \lambda_\mathrm{m}(1-\phi) \quad (4)$$

or a geometric mixing law,

$$\lambda = \lambda_\mathrm{w}^{\phi}\lambda_\mathrm{m}^{1-\phi} \quad (5)$$

where λ_w is the conductivity of the pore fluid and λ_m is the conductivity of the matrix material in W m^{-1} K^{-1} (Beardsmore and Cull 2001). Since porosity loss can be on the order of tens of percent within the upper few hundred meters of the sediment column, it implies pronounced variations in thermal conductivity and, under conductive heat flow, would theoretically generate a nonlinear geothermal gradient (Fig. 1). In an attempt to further investigate the degree to which observed thermal conductivity variations influence the geothermal gradient in marine sediments, the present study revisits a large dataset of in-situ temperature and thermal conductivity measurements from 205 Ocean Drilling Program (ODP) boreholes drilled by the *Joides Resolution* during Legs 101–180 (Pribnow et al. 2000; Fig. 2).

Background and methods

In a comprehensive report by Pribnow et al. (2000), heat flow calculations using the Bullard method were provided from ODP borehole data collected on Legs 101–180. They identified 205 holes where reliable in-situ temperature and thermal conductivity measurements exist. The average number of temperature measurements was 4.8 per hole, with the majority taken between 20 and 250 mbsf, and the deepest measurement from 550 mbsf. Of the 205 holes reported in the summary data sheet of Pribnow et al. (2000), only 186 holes were selected for analysis in the present study because certain holes were excluded due to insufficient temperature and/or conductivity observations (cf. Fig. 2 for geographic locations). The data are reported in Table 1 of the electronic supplementary material available online for this article.

The Pribnow et al. (2000) analysis of heat flow data adopted a method where a continuous thermal conductivity profile was modelled (λ_model) from discrete measurements performed on recovered sediments. This method ensured that thermal conductivity data were available over the full range of in-situ temperature measurements. The downhole thermal resistance profiles were generated by Pribnow et al. (2000) in three ways, depending on how thermal conductivity evolved with depth: type a: at 58 holes (31%), no discernible downhole trend existed in the thermal conductivity data, and a harmonic mean was adopted; type b: at 84 holes (45%), downhole increases (71 holes) or decreases (13 holes) were approximated

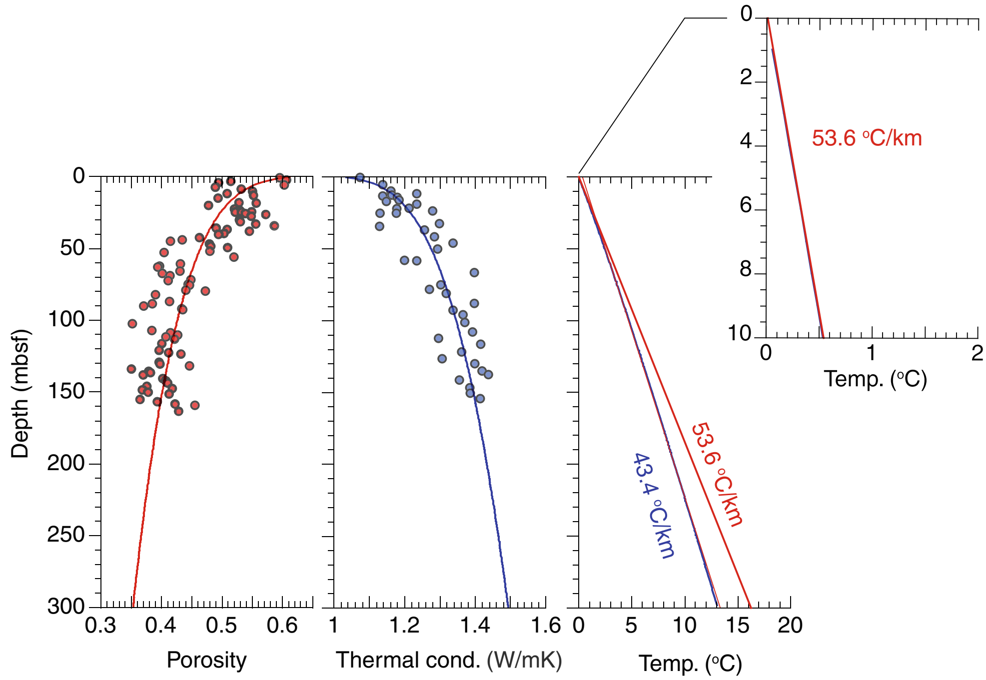

Fig. 1 Conceptual illustration of the relationships between porosity, thermal conductivity and in-situ temperature. Porosity and thermal conductivity data are from the Integrated Ocean Drilling Program's Expedition 302 to the Lomonosov Ridge in the central Arctic Ocean, for the upper 165 m of Neogene glaciomarine clays (O'Regan and Moran 2010; O'Regan et al. 2010). An exponential decrease in porosity is mirrored by an increase in thermal conductivity. Assuming a heat flow of 60 mW m^{-2}, the calculated in-situ temperature profile (*blue line*) can be approximated as a linear gradient of 43.4 °C km^{-1}, but a slight curvature is revealed by a linear gradient of 53.6 °C km^{-1} in the upper 10 mbsf

using a linear regression with depth; type c: at 44 holes (24%), the conductivity data were fitted with an exponential function (Fig. 2). This latter case represents the predicted profile of a relatively homogenous lithology undergoing compaction-driven porosity loss with depth (Fig. 1). Of the 186 holes, 115 (or 62%) show increasing conductivity with depth.

Thermal conductivity measurements at ODP boreholes are carried out on recovered sediments under laboratory

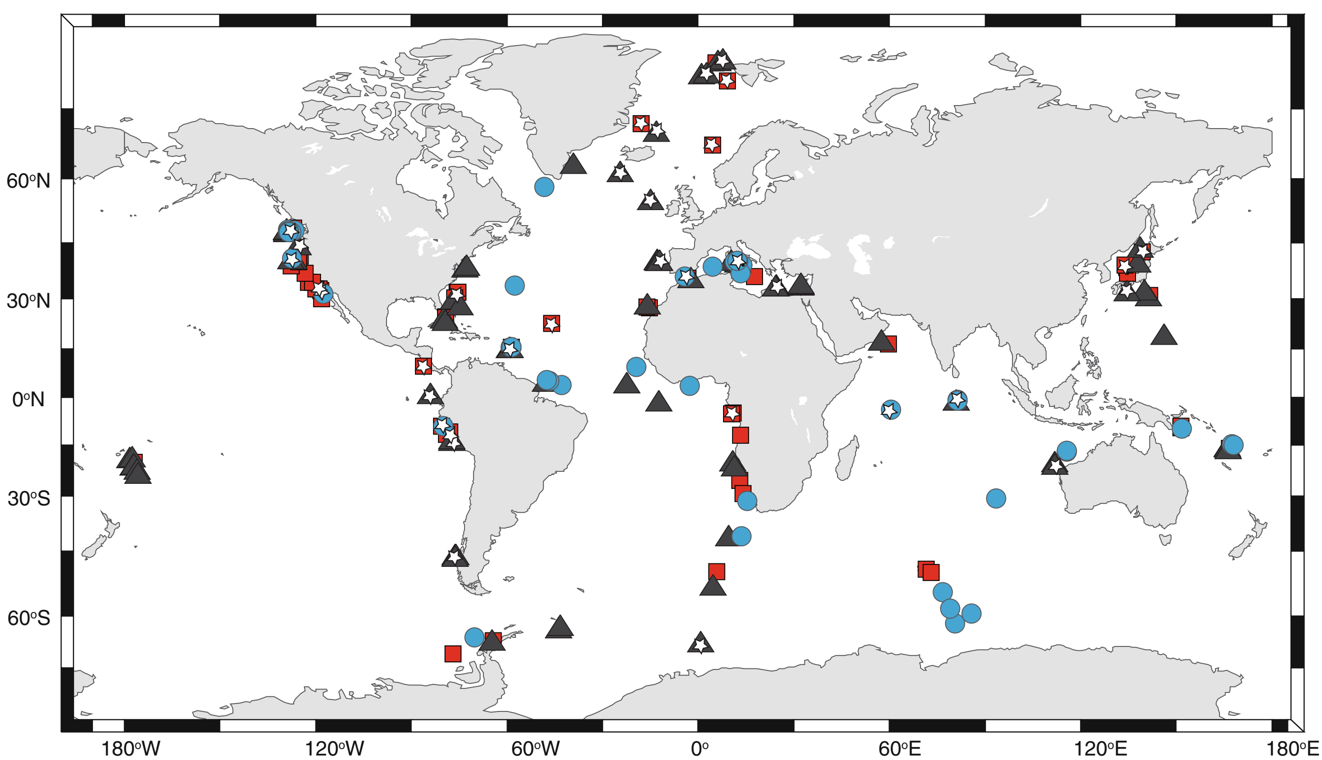

Fig. 2 Map of 186 ODP holes analysed by Pribnow et al. (2000) and revisited in this study. Downhole thermal conductivity data either remain constant (*red squares*), exhibit a linearly increasing/decreasing gradient (*grey triangles*), or increase exponentially with depth (*blue circles*). *White stars* Holes where nearby shallow lister-probe type measurements of surface heat flow exist in the Global Heat Flow Database (Pollack et al. 1993; Gosnold and Panda 2002)

temperature and pressure conditions. The in-situ thermal conductivity of marine sediments increases as both temperature and pressure increase (Ratcliffe 1960; Hyndman et al. 1974; Beardsmore and Cull 2001). An empirically derived formula presented by Hyndman et al. (1974) served to convert laboratory measurements to in-situ values:

$$\lambda_{\mathrm{P,T}}(z) = \lambda_{\mathrm{lab}}\left(1 + \frac{z_{\mathrm{w}} + \rho \cdot z}{1829 \cdot 100} + \frac{T_{\mathrm{z}} - T_{\mathrm{lab}}}{4 \cdot 100}\right) \tag{6}$$

where $\lambda_{\mathrm{P,T}}(z)$ is the in-situ thermal conductivity at depth z (meters below seafloor), λ_{lab} the laboratory measurement, z_{w} the water depth (m), ρ the average bulk density (g cm^{-3}) of the sediments, T_{z} the in-situ temperature, and T_{lab} the sample temperature during the λ_{lab} measurement. The empirical equation is only applicable for temperature ranges between 5 and 25 °C. At some holes where geothermal gradients were particularly high, the correction resulted in seemingly unrealistic in-situ thermal conductivity values. These 12 holes were excluded from this analysis.

Heat flow was calculated by performing linear regression analyses on the temperature data and the thermal resistance profiles. In reviewing the compiled heat flow calculations provided by Pribnow et al. (2000), a subtle error was recognised in the final regression analysis. Instead of performing a least squares regression using temperature as the dependent variable and thermal resistance as the independent variable, the published analysis derives heat flow as the reciprocal slope of a least squares regression using temperature as the independent variable.

The problem is exemplified in Fig. 3 where the heat flow reported by Pribnow et al. (2000) for hole 748 is 68 mW m^{-2} and where it is 52 mW m^{-2} from the corrected regression. Therefore, using the temperature and thermal resistance data published by Pribnow et al. (2000), the heat flows for the 186 holes were recalculated in the present study. The error in the calculated heat flow tends to be small but reaches up to several hundred percent at some boreholes. A histogram showing the relative difference (%) between the published and recalculated heat flows (q_{Pribnow}–$q_{\mathrm{This\_study}}$)/q_{Pribnow}) illustrates the overall impact of this recalculation (Fig. 4).

The corrected data served to investigate the suspected existence of nonlinear thermal gradients in marine sediments, and the implications for forward modelling of temperature from shallow heat flow surveys. Four aspects were evaluated: (1) the agreement between heat flow calculations that rely on λ_{model} and those based on $\lambda_{\mathrm{P,T}}$, reflecting the sensitivity of thermal resistance profiles in the Bullard method to these different approaches, (2) linear versus nonlinear geothermal gradients, by comparing linear regression statistics (R^2) performed for depth vs. temperature and thermal resistance vs. depth at each of the holes, (3) potential errors that would arise in forward modelling of in-situ temperatures if shallow (<10 mbsf) thermal conductivity measurements are assumed to be representative of the entire sedimentary column, and (4) interrelationships between linear temperature gradients approximated from in-situ measurements at ODP holes and temperature gradients from shallow penetrating probes deployed within a 10 km radius of the boreholes.

Results

Modelled vs. observed thermal conductivity profiles

To investigate the accuracy of the thermal conductivity modelling method, the recalculated heat flow using λ_{model} is compared with estimates that simply integrate the available thermal conductivity measurements ($\lambda_{\mathrm{P,T}}$) to produce a thermal resistance profile (Ω). Conductivity observations <0.6 W m^{-1} K^{-1} are excluded but an upper cut-off for λ is not applied. In order to obtain the seafloor temperature from this method, the Ω profile needs to intersect the seafloor. Therefore, a nearest neighbour extrapolation was applied to obtain the seafloor $\lambda_{\mathrm{P,T}}$.

Comparing the heat flow estimates from the λ_{model} and $\lambda_{\mathrm{P,T}}$ methods (Fig. 5), the average absolute difference between the two approaches is 4.1% but reaches up to 60%. In some instances, large discrepancies arise because temperature observations are sometimes excluded in the $\lambda_{\mathrm{P,T}}$ method, since only temperature observations within the depth range over which

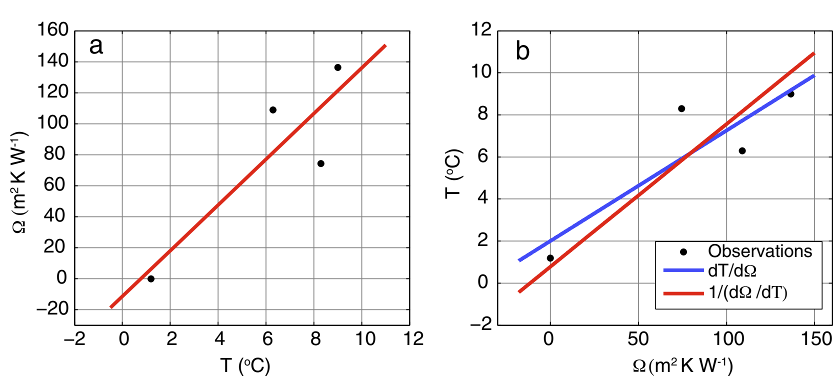

Fig. 3 Regression approaches for temperature and thermal resistance. **a** Observations of in-situ temperature and cumulative thermal resistance at ODP hole 748. The reciprocal of the linear regression provides a heat flow estimate of 68 mW m^{-2} (*red line*). **b** When thermal resistance is used as the independent variable, the same data produce a slope of 52 mW m^{-2} (*blue line*)

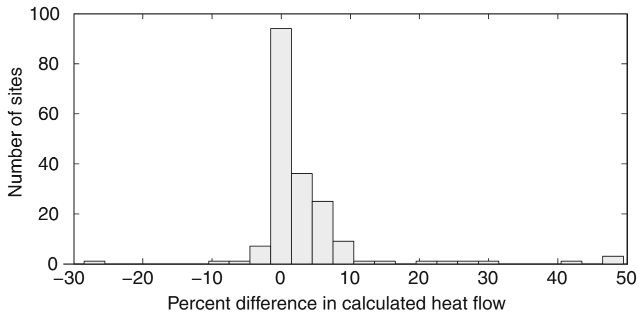

Fig. 4 Relative difference between the published heat flow (Pribnow et al. 2000) and recalculated heat flow (this study) for 186 ODP boreholes

conductivity measurements exist are taken into account. This leads to a less constrained geothermal gradient and results in a potentially larger heat flow error. Another possible reason for discrepancies between the methods are local differences between $\lambda_{P,T}$ and λ_{model}. This occurs when temperature measurements are confined to a narrow depth range or when well-defined lithology changes are present that are not captured in λ_{model}. In general, a carefully constructed λ_{model} profile that accounts for substantial changes in the downhole lithology produces results similar to a simple integration of $\lambda_{P,T}$, but allows for data gaps in either thermal conductivity or temperature to be more easily overcome. The λ_{model} profiles provided by Pribnow et al. (2000) are used in further data comparisons.

A further comparison is made between heat flow estimates using the Bullard method and λ_{model} and using Fourier's law (Eq. 1) with a constant thermal conductivity represented by the harmonic average (λ_{ave}) of $\lambda_{P,T}$. In this case, a linear geothermal gradient (T_{grad}) is calculated from the available temperature measurements. The different heat flow calculations are reported in Table 1 of the electronic supplementary material available online for this article. Differences in the calculated heat flow are usually less than 20% and are evenly distributed around zero (Fig. 5). This result suggests that using the harmonic average of thermal conductivity measurements does not generate systematic biases in the heat flow calculation if the thermal gradient and conductivity profile are constrained by deep measurements (i.e. many 10s to 100s of meters beneath the seafloor).

Linear regressions of temperature, thermal resistance and depth

In instances where thermal conductivity changes downhole, the Bullard method predicts a nonlinear temperature profile. This is the basis for introducing the concept of the thermal resistance profile, which produces a linear relationship with temperature in a system dominated by conductive heat flow (Bullard 1939).

A simple test of whether this is observed in the ODP boreholes is made by comparing the coefficients of determination (R^2) for linear regressions performed on temperature (T) vs. depth (z), and on T vs. Ω at each hole (Fig. 6). A perfect linear fit would produce an R^2 value of 1. Theoretically, the linear fit should be better between T and Ω than between T and z. The R^2 from both methods are generally high (between 0.47 and 1.00 with a mean of 0.96), implying that any nonlinearity of the temperature profiles is subtle at the holes in general. However, the T vs. Ω fit is equal to or better than the T vs. z fit at 60% of the investigated holes (Fig. 6). There is no difference in the methods when the thermal conductivity profile remains constant with depth, but a clear bias towards higher coefficients of determination exists when the thermal conductivity profile increases exponentially with depth (Fig. 6).

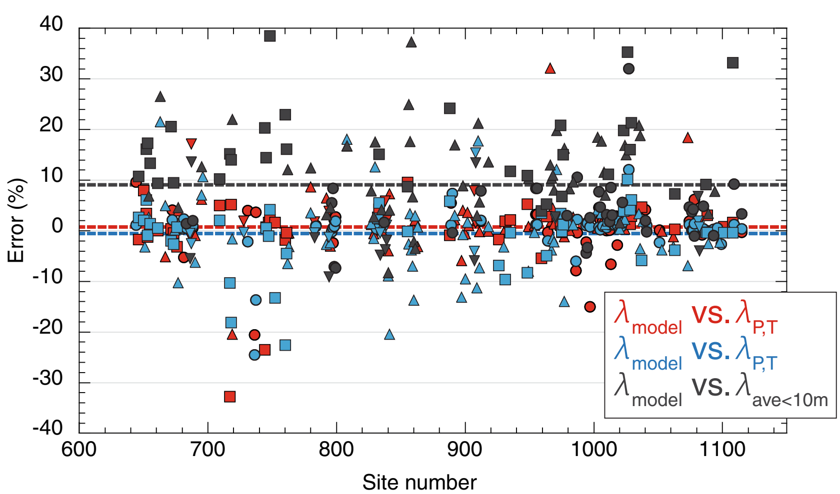

Fig. 5 Percent error between the heat flow calculated using hole-specific thermal conductivity models (λ_{model}) and (1) the integrated thermal resistance profile from $\lambda_{P,T}$, (2) the average thermal conductivity from all measurements at the hole λ_{ave}, and (3) the average thermal conductivity from the upper 10 m of the hole ($\lambda_{ave>10\ m}$). Thermal conductivity models either remain constant (*squares*), exhibit a linearly increasing/decreasing gradient (*triangles*), or increase exponentially with depth (*circles*)

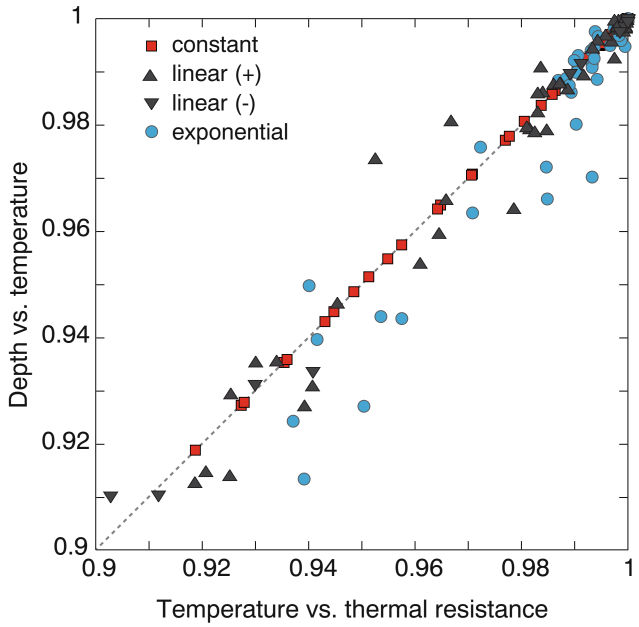

Fig. 6 Comparison of the coefficient of determination R^2 for depth vs. temperature and temperature vs. thermal resistance. The downhole thermal conductivity data either remain constant (*red squares*), exhibit a linearly increasing/decreasing gradient (*grey triangles*), or increase exponentially with depth (*blue circles*)

Extrapolating shallow temperature gradients to depth

To simulate a case where only shallow thermal conductivity measurements are available, the heat flow at each hole is recalculated using the harmonic mean of the thermal conductivity measurements from the upper 10 m ($\lambda_{ave<10\ m}$). Excluded are holes where the number of available λ measurements shallower than 10 mbsf is less than 4 (i.e. 54 holes). The results show that using shallow conductivity measurements in the heat flow calculations introduces a 10% bias towards underestimating the heat flow when the geothermal gradient is constrained by deeper measurements (Fig. 5). The linear approximations of heat flow using the harmonic mean of the thermal conductivity measurements (from either the full depth range λ_{ave} or the upper 10 m $\lambda_{ave<10\ m}$) result in increasingly large errors for the predicted in-situ temperature as depth increases (Fig. 7). The overall tendency is for both these methods to overestimate the temperature at depth.

Shallow vs. deep temperature gradients

Linear approximations of geothermal gradients derived from shallow heat flow surveys (<10 mbsf) should be higher than those from deep boreholes if downhole increases in thermal conductivity exist. To test this, geothermal gradients from the ODP boreholes are compared with nearby shallow heat flow measurements extracted from the World Heat Flow databank (Pollack et al. 1993; Gosnold and Panda 2002).

Using only the holes from Pribnow et al. (2000) with a geothermal gradient <250 °C km^{-1} (180 out of 205 holes), 98 holes were found to have shallow geothermal gradient measurements performed within 10 km of the borehole (Fig. 1). In some cases there were more than one observation, so the average was calculated. Cross-plotting these data, it is clear that shallow measurements consistently produce a higher geothermal gradient than the linear approximation from the ODP boreholes (Fig. 8). The average difference between shallow and deep observations is 19 °C km^{-1}.

Discussion

The overwhelming majority of heat flow surveys are conducted using shallow (<10 m) in-situ measurements of thermal conductivity and sediment temperature. The surface heat flow is derived from these measurements, and used to predict the temperature at greater burial depths. It is largely recognised that this requires an understanding of how thermal conductivity changes with burial depth (Beardsmore and Cull 2001). However, Grevemeyer and Villinger (2001) suggest that extrapolating the near-surface temperature gradient is sufficient to model temperatures at depth, and that observed downhole increases in thermal conductivity arise from measurement artefacts. They argue that the majority of thermal conductivity measurements are biased by a large horizontal component influenced by sediment fabric development during consolidation, and that the vertical thermal conductivity likely remains constant with depth, exerting the primary control on the geothermal gradient.

In ODP boreholes, the large depth range over which in-situ temperature and thermal conductivity data are collected allows the influence of thermal conductivity to be directly tested. Overall, the thermal conductivity of sediments at the 186 holes investigated here does increase with burial depth (Fig. 9). Using the complete dataset, the harmonic mean for λ is 1.15 W m^{-1} K^{-1}, whereas in the upper 10 m it is only 1.00 W m^{-1} K^{-1}. This increasing trend exists in the uncorrected and corrected datasets, and is not generated by the in-situ temperature and pressure correction (Hyndman et al. 1974; Fig. 9). Because both the temperature and the thermal conductivity of sediments are constrained in ODP boreholes over a large depth range, on average the heat flow calculations using the Bullard method and those derived by Fourier's law (using a linear geothermal gradient and average thermal conductivity) provide similar results (Fig. 5). However, this is not the case when the average thermal conductivity from only the upper 10 mbsf is used (Fig. 5). This implies that estimating the heat flow from Eq. 1 using deep in-situ

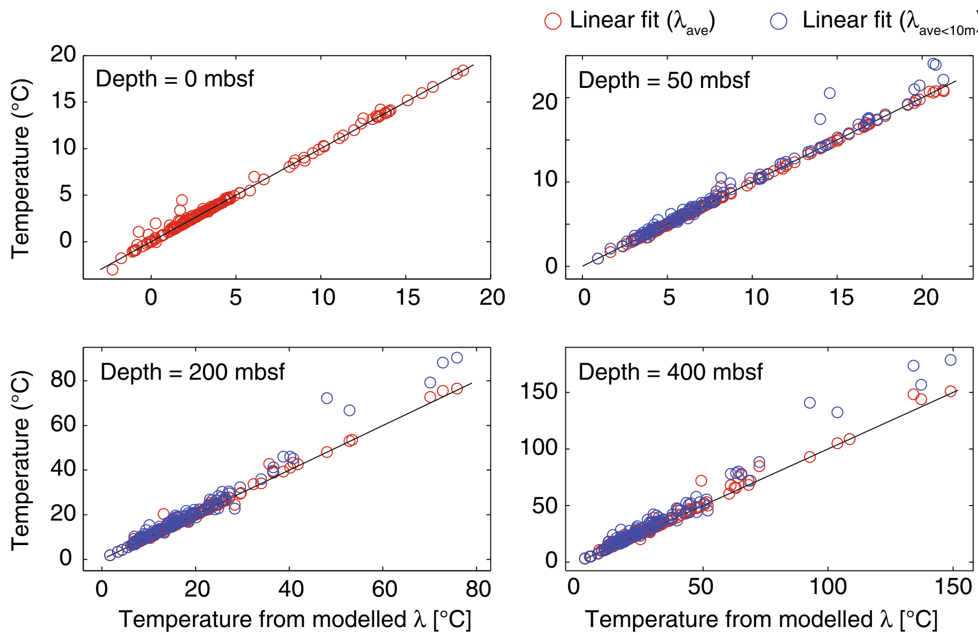

Fig. 7 Differences in predicted in-situ temperature when heat flow is calculated using λ_{ave} and $\lambda_{ave<10\ m}$

temperature estimates (e.g. calculated from seismic observations on the depth to the base of the gas hydrate stability zone), and with thermal conductivity derived only from shallow measurements, would lead to underestimates on the order of 10% (Fig. 5).

The role that thermal conductivity has on regulating the in-situ temperature profile is perhaps best exemplified in Fig. 6. In the majority of cases where the thermal conductivity changes downhole, the temperature vs. thermal resistance regression (the Bullard method) produces a more linear fit. Together, these observations strongly argue against an apparent increase in thermal conductivity related to anisotropy and fabric development, and point towards a clear interplay between the measured thermal conductivity of sediments and changes in the in-situ thermal gradient.

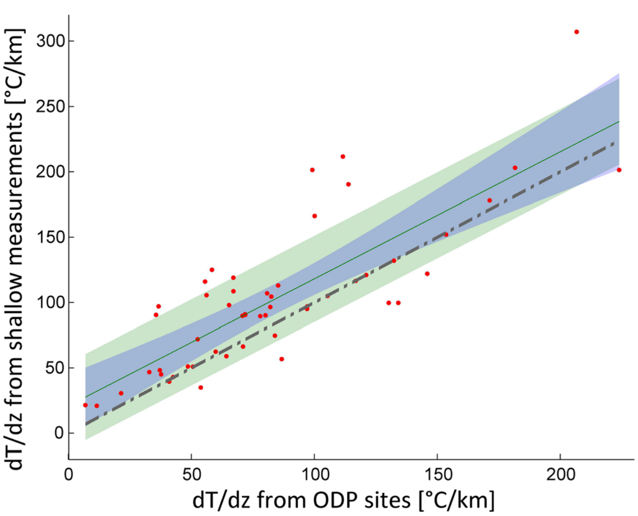

Fig. 8 Comparison of linear geothermal gradients from ODP boreholes, and those from shallow in-situ measurements at stations located within a 10 km radius of the drilling holes. *Dashed line* 1-to-1 line, *green line* linear regression, *green shading* one standard deviation from the linear regression, *blue shading* 95% confidence interval for the linear regression. The mean difference is 19 °C km^{-1}

The comparison between the linearly approximated geothermal gradients derived from the ODP boreholes and those taken from shallow penetrating measurements located within a radius of 10 km (Fig. 8) further highlights the role that thermal conductivity has on in-situ temperatures. Shallow measurements generally provide a substantially higher geothermal gradient than the one derived from the deeper borehole. This is consistent with theoretical predictions, where porosity loss during burial increases the thermal conductivity and lowers the geothermal gradient (Fig. 1).

Downhole changes in lithology and sediment composition will also influence porosity and thermal conductivity profiles. As such, an exponential decrease in porosity and increase in thermal conductivity (cf. Fig. 1) is not a default scenario for marine sediments. In this study, it only represented 24% of the investigated holes. However, irrespective of how the thermal conductivity changes during burial, it clearly influences the geothermal gradient. Due to the clear influence that thermal conductivity has on the downhole geothermal gradient, a more detailed look at lithologic controls and geographic variability (cf. Stein and Abbott 1991) in porosity and thermal conductivity relationships is warranted.

Conclusions

Using data from the Ocean Drilling Program compiled by Pribnow et al. (2000), the present study investigated the

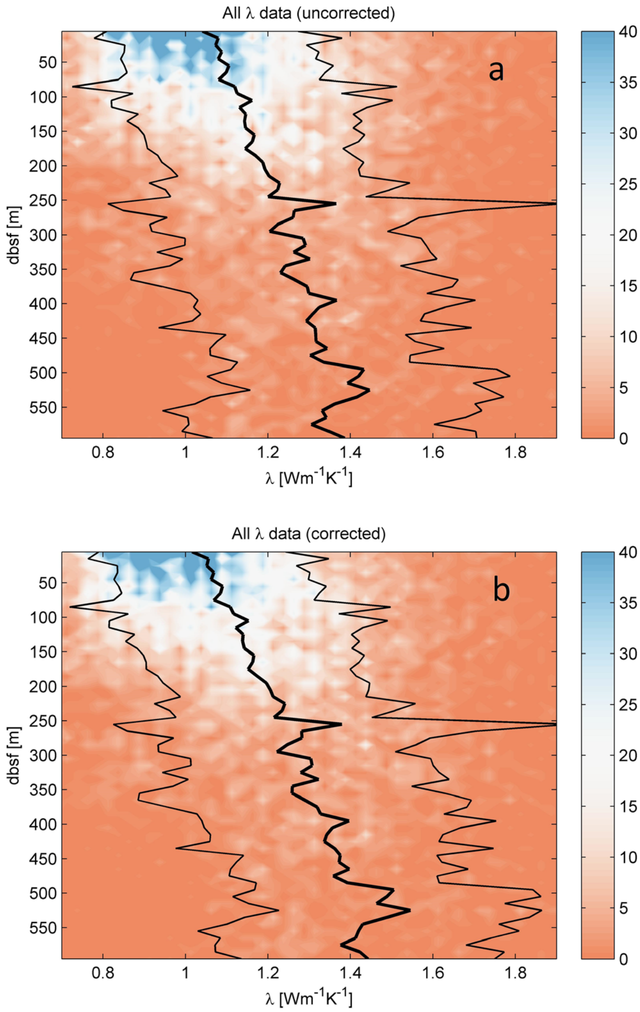

Fig. 9 Thermal conductivity of sediments from the 186 ODP holes based on a given number of measurements in each depth range. *Thick black line* Averages for each 10 m depth interval, *thin lines* ± 1 standard deviation. **a** Data from all boreholes, uncorrected for in-situ temperature and pressure. **b** Corresponding corrected data

occurrence of nonlinear geothermal gradients generated by changes in sediment thermal conductivity. It has been shown that knowledge concerning how the thermal conductivity of sediments changes with burial is needed to accurately predict in-situ temperatures if the surface heat flow is known. The observed downhole increase of sediment thermal conductivity measured in ODP boreholes does impact the geothermal gradient, and is not simply a horizontal measurement bias in sediments that display increasingly anisotropic behaviour during consolidation (Grevemeyer and Villinger 2001). Comparisons between shallow measurements (<10 m) from surface heat flow surveys and the deeply constrained temperature data from 98 ODP boreholes indicate that the shallow gradients are consistently higher by on average 19 °C km^{-1}. The results provide empirical support for the need to develop robust porosity–thermal conductivity models to accurately predict temperature at depth from shallow heat flow surveys.

Acknowledgements Funding for M. O'Regan and C. Stranne was provided by the Swedish research Council (VR). This research used data from the Ocean Drilling Program (ODP). ODP was sponsored by the U.S. National Science Foundation (NSF) and participating countries under management of Joint Oceanographic Institutions (JOI), Inc. We acknowledge constructive comments from an anonymous reviewer and the journal editors.

Compliance with ethical standards

Conflict of interest The authors declare that there is no conflict of interest with third parties.

References

Artemieva IM, Mooney WD (2001) Thermal thickness and evolution of Precambrian lithosphere: a global study. J Geophys Res 106:16387. doi:10.1029/2000JB900439

Athy LF (1930) Density, porosity, and compaction of sedimentary rocks. AAPG Bull 14:1–24

Beardsmore GR, Cull JP (2001) Crustal heat flow: a guide to measurement and modelling. Cambridge University Press, Cambridge

Bullard EC (1939) Heat flow in South Africa. Proc R Soc Lond Ser Math Phys Sci 173:474–502

Chapman DS (1986) Thermal gradients in the continental crust. Geol Soc Lond Spec Publ 24:63–70. doi:10.1144/GSL.SP.1986.024.01.07

Chapman D, Keho T, Bauer M, Picard M (1984) Heat flow in the Uinta Basin determined from bottom hole temperature (BHT) data. Geophysics 49:453–466. doi:10.1190/1.1441680

Davies JH (2013) Global map of solid Earth surface heat flow. Geochem Geophys Geosyst 14:4608–4622. doi:10.1002/ggge.20271

Gibson RE (1958) The progress of consolidation in a clay layer increasing in thickness with time. Geotechnique 8:171–182. doi:10.1680/geot.1958.8.4.171

Gosnold WD, Panda B (2002) The global heat flow database of the international heat flow commission. http://www.und.edu/org/ihfc/index2.html

Grevemeyer I, Villinger H (2001) Gas hydrate stability and the assessment of heat flow through continental margins. Geophys J Int 145: 647–660. doi:10.1046/j.0956-540x.2001.01404.x

Grose CJ (2012) Properties of oceanic lithosphere: revised plate cooling model predictions. Earth Planet Sci Lett 333–334: 250–264. doi:10.1016/j.epsl.2012.03.037

Hasterok D (2013) A heat flow based cooling model for tectonic plates. Earth Planet Sci Lett 361:34–43. doi:10.1016/j.epsl.2012.10.036

Hyndman RD, Erickson AJ, Von Herzen RP (1974) Geothermal measurements on DSDP Leg 26. In: Davies TA, Luyendyk BP et al., Init Repts DSDP 26:451–463

Kukkonen IT (1988) Terrestrial heat flow and groundwater circulation in the bedrock in the central Baltic Shield. Tectonophysics 156:59–74. doi:10.1016/0040-1951(88)90283-1

Mondol NH, Bjørlykke K, Jahren J, Høeg K (2007) Experimental mechanical compaction of clay mineral aggregates—changes in physical properties of mudstones during burial. Mar Pet Geol 24:289–311. doi:10.1016/j.marpetgeo.2007.03.006

Moore JC, Saffer D (2001) Updip limit of the seismogenic zone beneath the accretionary prism of southwest Japan: an effect of diagenetic to low-grade metamorphic processes and increasing effective stress. Geology 29:183–186. doi:10.1130/0091-7613(2001)029<0183:ULOTSZ>2.0.CO;2

O'Regan M, Moran K (2010) Deep water methane hydrates in the Arctic Ocean: reassessing the significance of a shallow BSR on the Lomonosov Ridge. J Geophys Res Solid Earth 115, B05102. doi: 10.1029/2009JB006820

O'Regan M, Moran K, Baxter CDP, Cartwright J, Vogt C, Kölling M (2010) Towards ground truthing exploration in the central Arctic Ocean: a Cenozoic compaction history from the Lomonosov Ridge. Basin Res 22:215–235. doi:10.1111/j.1365-2117.2009.00403.x

Pollack HN, Hurter SJ, Johnson JR (1993) Heat flow from the Earth's interior: analysis of the global data set. Rev Geophys 31:267–280. doi:10.1029/93RG01249

Powell WG, Chapman DS, Balling N, Beck AE (1988) Continental heat-flow density. In: Haenel R, Rybach L, Stegena L (eds) Handbook of terrestrial heat-flow density determination. Springer, Dordrecht, pp 167–222

Pribnow D, Kinoshita M, Stein C (2000) Thermal data collection and heat flow recalculations for Ocean Drilling Program Legs 101–180. ODP Heat Flow Rep, pp 1–25

Ratcliffe EH (1960) The thermal conductivities of ocean sediments. J Geophys Res 65:1535–1541

Stein CA, Abbott DH (1991) Implications of estimated and measured thermal conductivity for oceanic heat flow studies. Mar Geophys Res 13:311–329. doi:10.1007/BF00366281

Terzaghi K (1943) Theoretical soil mechanics. Wiley, New York

Tissot BP, Pelet R, Ungerer P (1987) Thermal history of sedimentary basins, maturation indices, and kinetics of oil and gas generation. AAPG Bull 71:1445–1466

Villar-Muñoz L, Behrmann JH, Diaz-Naveas J, Klaeschen D, Karstens J (2013) Heat flow in the southern Chile forearc controlled by large-scale tectonic processes. Geo-Mar Lett 34:185–198. doi:10.1007/s00367-013-0353-z

Wallmann K, Pinero E, Burwicz E, Haeckel M, Hensen C, Dale A, Ruepke L (2012) The global inventory of methane hydrate in marine sediments: a theoretical approach. Energies 5:2449–2498. doi:10.3390/en5072449