A Revised Oceanic Plate Cooling Model

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Key Points:

- Global subsidence (n = 10,863) and heatflow (n = 3,573) measurements used to reassess oceanic plate cooling model.
- Optimal model yields equilibrium plate thickness of 105±10 km with basal temperature of 1326 ± 50°C.
- Thickness of seismogenic layer determined by depth to 700° C isothermal surface.

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Abstract

Global age-depth and heatflow observations provide constraints for cooling and subsidence of oceanic plates. Numerous studies have addressed this problem which has a bearing upon the calibration of shear-wave tomographic models and upon lithospheric rheology. The robustness of these results depends upon the quality and spatial distribution of both age-depth and heatflow measurements. Here, we revisit plate cooling for two reasons. First, a significantly revised and augmented database of 10,863 age-depth measurements that are evenly distributed throughout the oceanic realm has been constructed. This database is combined with a modern heatflow database of 3,573 measurements. Secondly, we wish to explore a range of analytical and numerical plate models that incorporate temperature- and pressure-dependence of conductivity, of expansivity, and of specific heat capacity. Our goal is to identify plate models that jointly fit observational constraints, whilst honoring laboratory-based constraints for the values of key thermal parameters. The optimal model yields an equilibrium plate thickness of 105±10 km with a basal temperature of 1326±50°C and a zero-age depth of 2.60±0.3 km. The recovered basal temperature is consistent with independent petrologic constraints. Spatial analysis of our database of age-depth measurements demonstrates that previously invoked transient shallowing is a sampling artefact. The absence of shallowing implies that the Rayleigh-Taylor instability, which stabilizes equilibrium plate thickness, probably acts on a smaller length scale than implied by previously reported transient shallowing. Finally, our revised plate model is used to track lithospheric thermal structure with time and to calculate residual depth anomalies.

Plain Language Summary

The Earths outermost layer consists of interlocking tectonic plates. Throughout the oceans, seafloor is continuously created at mid-oceanic ridges, which are associated with high temperatures. When new seafloor moves away from the mid-oceanic ridge, it gradually cools and sinks over many millions of years. Although this interesting problem has been studied by many scientists, it is always insightful to revisit it using modern measurements. Here, we use a newly assembled set of global depth and temperature measurements to revise the plate cooling model. Our results show that the model fits the measurements really well. We show that the way in which the oceanic plates cool requires some modification.

1 Introduction

It is generally accepted that oceanic lithosphere cools and subsides as a function of plate age (McKenzie, 1967; Turcotte & Oxburgh, 1967). This phenomenon is manifest by a combination of both subsidence and heatflow observations from the oceanic realm. Since the late 1960s, two principal classes of models have been developed to account for the evolution of oceanic lithosphere. Early studies suggested that a half-space cooling model accounts for age-depth observations (Takeuchi & Sakata, 1970; Davis, 1974). In this model, subsidence and heat flow are functions of the square root of crustal age. The half-space cooling model is based upon observations from relatively young (i.e. <80 Ma) oceanic crust and it implicitly assumes that cooling and subsidence continue indefinitely with time. The observational database expanded when studies of oceanic magnetic anomalies enabled ages of subsidence anomalies on older oceanic crust to be dated (Vogt et al., 1971; Larson & Hilde, 1975; Hilde et al., 2013). Subsequently, Parsons and Sclater (1977) analyzed measurements from oceanic crust as old as 163 Ma and observed that the depth of seafloor that is older than $\sim$ 80 Ma is shallower than that predicted by the half-space cooling model. They proposed that the plate model of McKenzie (1967), where temperature along a basal boundary of the cooling lithosphere is fixed to simulate resupply of heat, yields a better fit to a combination of subsidence and heatflow measurements. Stein

and Stein (1992) built upon this pioneering analysis by jointly inverting an extensive database of subsidence and heatflow measurements from the North Pacific Ocean and from the northwest Atlantic Ocean. It is now generally accepted that the plate model, rather than the half-space cooling model, provides a superior description of the subsidence and heatflow characteristics of oceanic lithosphere. A significant drawback of the half-space cooling model is its inability to predict a basal temperature that is consistent with independent petrologic and geochemical constraints (Hoggard et al., 2017).

In recent years, there has been considerable interest in developing and applying more physically realistic plate models that include, for example, temperature- and pressure-dependence of thermal parameters. McKenzie et al. (2005) showed that the thermal structure of the lithosphere can be constrained by using experimentally determined values of thermal conductivity, k, of thermal expansivity, α, and of specific heat capacity, C_p. Subsequent models have incorporated more sophisticated parameterizations, which include lithostatic pressure gradients, the radiative component of thermal conductivity, mineral phase transitions, hydrothermal circulation as well as an insulating crustal layer (Afonso et al., 2007; Grose & Afonso, 2013; Korenaga & Korenaga, 2016; Richards et al., 2018). Notwithstanding these developments, the success of a lithospheric thermal model still depends upon the ability to match a combination of subsidence and heatflow measurements. When jointly inverting these measurements, the robustness of an optimal model is predicated upon their quality and spatial distribution. A revised and augmented global database of subsidence observations assembled by Holdt et al. (2022) presents an opportunity to reassess the range of analytical and numerical plate models.

Plate models typically recover values of temperature, T, lithospheric thickness, z_p, and ridge depth, z_r. For analytical models, the T is defined as the temperature along the basal boundary of the lithosphere, T_b. In contrast, numerical models have a more flexible parameterization, which can be used to design a more physically realistic temperature structures at and away from boundaries. In these schemes, T is the average potential temperature of the basal boundary, T_p (Mckenzie & Bickle, 1988). z_p is the base of oceanic lithosphere which is defined to consist of the mechanical boundary layer and part of the thermal boundary in accordance with White and McKenzie (1988). It can also be regarded as an isothermal surface (see Priestley and McKenzie (2013)). Finally, z_r is defined as the global average depth of the mid-oceanic ridge axes that is isostatically corrected for sedimentary load and variable crustal thickness. Since independent constraints for these three parameters exist, values recovered for the optimal plate model are testable.

The aims of this contribution are threefold. First, we test a variety of plate models by jointly inverting global databases of subsidence and heatflow measurements constructed by Holdt et al. (2022) and by Hasterok et al. (2011), respectively (Figure 1). These plate models include two analytical plate models and a more sophisticated temperature- and pressure-dependent parameterization. Our goal is to identify a parsimonious model that yields the best fit to observational constraints. Secondly, we demonstrate that transient shallowing for plate ages of 80–130 Ma that was observed by Crosby et al. (2006) and by some older studies is an artefact of spatial sampling. This demonstration highlights the importance of exploiting global subsidence and heatflow databases which have uniform global coverage. Finally, we use our results to discuss the nature of the Rayleigh-Taylor instability, which is inferred to stabilize plate thickness, and to revise the rheological properties of oceanic lithosphere.

2 Observational Constraints

2.1 Subsidence Measurements

Age-depth observations are the primary constraint for any plate model since they determine how oceanic lithosphere subsides as a function of time as a result of cooling and thickening. Here, we exploit a revised and augmented global database of water-loaded measurements of the depth to oceanic basement that was compiled by Holdt et al. (2022). This database contains 10,863 measurements and represents the largest compilation of accurately determined water-loaded estimates obtained from a combination of modern/legacy seismic reflection profiles and wide-angle experiments (Figure 1a). Both academic and commercial datasets were exploited.

Water-loaded depth to oceanic basement, w, is calculated as accurately as possible by carrying out two very well-known corrections. The sedimentary correction, C_s, converts the variable thickness and density of the sedimentary load into an equivalent water load using the scheme described in detail by Hoggard et al. (2017) and subsequently revised by Holdt et al. (2022). The crustal correction, C_c, converts the variable thickness and density of oceanic crust into a standard crustal reference column. The corrected depth to oceanic basement is given by $w = z_{sb} + C_s + C_c$ where z_{sb} is bathymetry. It is important to emphasize that C_s is generally more significant than C_c. From a global perspective, C_s is accurately determinable since $\sim 70\%$ of locations have sedimentary columns that are ≤ 0.5 km. Where crustal thickness and density are unknown, the value of C_c is gauged from regional crustal information, which yields either an upper or a lower bound for w. Table 1 lists the symbols and values of all parameters used within this study.

The database constructed by Holdt et al. (2022) represents a significant improvement in terms of both the quantity and the spatial distribution of accurate age-depth measurements. Previously constructed databases are often of only regional extent (see, e.g., Parsons and Sclater (1977); Stein and Stein (1992)), rely upon gridded values whose accuracy is difficult to gauge (see, e.g., Crosby et al. (2006); Crosby and McKenzie (2009)), or are spatially biased toward continental margins (see, e.g., Richards et al. (2018)). The revised and augmented database builds upon the detailed approach taken by Hoggard et al. (2017), primarily by interpreting substantial quantities of modern multi-channel and legacy single-channel seismic reflection profiles that criss-cross oceanic lithosphere of different ages. The inclusion of additional profiles dramatically improves spatial coverage throughout the oceanic realm. Subsidence measurements are always carried out on *bona fide* oceanic crust located away from anomalous tectonic features (e.g. seamounts, fracture zones, flexural moats). To aid spatial clarity, detailed measurements along individual profiles are binned every 1°. They are also assigned ages using a modern magnetic age grid (Seton et al. (2020); Figure 1c).

Figure 2a presents 10,863 age-depth measurements as a function of oceanic plate age. Note that more uncertain age-depth measurements from oceanic crust that is older than 200 Ma are omitted. This cloud of measurements has two significant characteristics. First, a mean relationship is evident whereby subsidence increases exponentially from ~ 3 km to ~ 6 km as a function of plate age. Secondly, there is considerable but relatively uniform scatter about this mean relationship. The envelope of scatter is consistently ± 1 km, regardless of plate age. Since the pioneering work of Menard (1973), it is generally accepted that age-depth measurements that lie above or below the mean relationship represent positive and negative residual depth anomalies, respectively. These anomalies are interpreted as maximum signals of dynamic topography generated by mantle convective processes. For comparative purposes, Figure 2b shows these age-depth measurements binned every 2.5 Ma.

Previous plate model studies have often attempted to minimize the influence of dynamic topography by excising regions dominated by dynamic topographic anomalies. For

example, Crosby et al. (2006) exploited subsidence observations from regions which featured near-zero (or at least minimal) gravity anomalies in order to avoid significant dynamic upwelling and downwelling. Goutorbe and Hillier (2013) removed subsidence measurements within the vicinity of major mantle plumes. Subsequently, Hoggard et al. (2016) used a global database of age-depth measurements which demonstrated that residual depth anomalies are ubiquitous down to wavelengths as short as 1000 km. Here, we have not applied any such spatial filters. Instead, we assume that our database of 10,863 measurements represents a uniformly dense sampling of the oceanic realm so that dynamic upwellings and downwellings are potentially equally included over a complete range of plate ages.

2.2 Heatflow Measurements

Heatflow observations, which enable conductive heat loss through the surface of the lithospheric plate to be determined as a function of time, also help to constrain plate cooling. In the oceanic realm, standard heatflow measurements are acquired using a probe that is inserted into the sedimentary pile which measures temperature gradient and the *in situ* thermal conductivity (Sclater et al., 2014). Heat flow, H, is given by $H = k(\Delta T/\Delta z)$, where k is thermal conductivity, and $\Delta T/\Delta z$ is the thermal gradient between two points. A comprehensive global compilation of 23,428 heat flow observations was carefully compiled by Hasterok et al. (2011). This substantial undertaking provides an excellent complementary constraint for plate model considerations (Figure 1b).

Following the detailed methodology outlined by Hasterok et al. (2011) and by Richards et al. (2018), heatflow measurements are filtered and corrected. First, observations are edited by removing points which do not lie upon *bona fide* oceanic crust, by deleting negative values, and by binning the data at intervals of 0.1° to reduce spatial bias. Secondly, measurements are excised if they lie between 0 and 65 Ma and have a sedimentary thickness of less than 0.4 km, or if they are within 60 km of seamounts or igneous plateaux. These filtering strategies isolate and remove any measurements that could be significantly perturbed by hydrothermal circulation, which primarily affects young oceanic crust (Hasterok, 2013; Stein & Stein, 1992). For this purpose, sedimentary thicknesses are taken from the GlobSed global grid (Straume et al., 2019). Locations of seamounts and large igneous provinces are taken from Wessel et al. (2010) and from Coffin and Eldholm (1994), respectively. Heatflow measurements from oceanic crust that is older than 200 Ma are omitted on two grounds. First, there are only ~ 10 measurements on oceanic crust beyond 200 Ma. Secondly, their values lie below the average value of 0.05 W m^{-2} at old plate ages. Finally, a sedimentation correction is applied in order to account for depression of the geothermal profile caused by rapid deposition. Reduced heat flow occurs since the initial temperature of uncompacted sediment is equal to that of bottom water. Assuming that the rate of sedimentation and the value of thermal diffusivity are constant as a function of time, Von Herzen and Uyeda (1963) calculated the fractional perturbation of the geothermal profile at the seabed, F, in the absence of internal heat generation. A simplified version of the analytical solution is given by

$$F = 1 + 2Y^2\mathrm{erfc}(Y) - \mathrm{erf}(Y) - \frac{2Y}{\sqrt{\pi}}\exp(-Y^2), \tag{1}$$

where $Y = \frac{1}{2}Ut_s^{\frac{1}{2}}\kappa_s^{-\frac{1}{2}}$, U is the rate of sedimentation, t_s is the duration of sedimentation, and κ_s is thermal diffusivity of the sedimentary pile. Following Hasterok et al. (2011) and Richards et al. (2018), U is estimated by dividing sedimentary thickness by t_s, which is assumed to be equal to plate age. The impact of the sedimentation correction is relatively small. For example, 76% of measurements are corrected by $\leq$10%, when κ_s is assumed to be 0.25 mm^2 s^{-1}. Following Richards et al. (2018), measurements from certain anomalous regions (e.g. Caspian Sea, Gulf of Mexico, Black Sea) are omit-

ted because U greatly increased in these regions during Miocene times, which violates the assumption of sedimentation rate (Knapp et al., 2000; Galloway et al., 2011; Sydorenko et al., 2017).

Application of this filtering and editing process yields a set of 3,573 heatflow measurements. Figure 2c presents corrected heatflow measurements as a function of oceanic plate age. Although there is considerable and variable scatter, a clear average relationship is visible. The average heat flow at mid-oceanic ridges is ~ 0.25 W m^{-2} and decreases exponentially to a value of ~ 0.05 W m^{-2} for the oldest oceanic lithosphere. For comparative purposes, these measurements have also been binned every 2.5 Ma (Figure 2d).

3 Modeling Strategy

The first detailed application of a plate model to age-depth and heatflow observations was carried out by Parsons and Sclater (1977) who derived appropriate partial differential equations, developed analytical solutions, and matched theory with observation. Their primary focus concerned fitting of age-depth observations from the North Pacific and North Atlantic oceans. They separately showed that their optimal plate model also matched limited heatflow observations. Subsequently, Stein and Stein (1992) employed a very similar analytical approach to jointly invert age-depth and heatflow observations. In both publications, the primary focus is recovery of basal temperature and equilibrium plate thickness. The critical underpinning thermal parameters are thermal conductivity, k, thermal expansivity, α, and isobaric specific heat capacity, C_p. Analytical solutions can be derived if these three parameters are assumed to be constant.

McKenzie et al. (2005) developed a more physically realistic model, which accounts for the temperature- and pressure-dependence of k, α and C_p. This elaboration was motivated by the results of laboratory experiments, which clearly demonstrate that the values of these thermal parameters vary significantly over the temperature and pressure ranges observed within an olivine-dominated lithospheric mantle (Berman & Aranovich, 1997; Bouhifd et al., 1996; Hofmeister, 2010). Subsequent plate models have become increasingly sophisticated and, in general, are numerically solved in the absence of easily derived analytical solutions (Grose & Afonso, 2013; Korenaga & Korenaga, 2016; Richards et al., 2018).

Thermal evolution of oceanic lithosphere is calculated by solving the one-dimensional heat equation in a reference frame that translates horizontally as the lithosphere moves away from the ridge axis. The general form of this equation is given by

$$\rho(T,P,X), C_p(T,X)\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left[k(T,P,X)\frac{\partial T}{\partial z}\right], \tag{2}$$

where t is time, z is depth and ρ is density. T, P, and X are temperature, pressure and composition, respectively. In Equation (2), k and ρ depend upon T, P, and X. C_p primarily depends upon T and X since its pressure dependence is negligible over the range of pressures encountered within the lithosphere (Hofmeister, 2007).

If k, α and C_p are assumed to be constant, analytical solutions for subsidence and heat flow can be derived using standard Fourier expansion (Carslaw & Jaeger, 1950). Subsidence as a function of time, $w(t)$, is given by

$$w(t) = z_r + \frac{\rho_m \alpha (T - T_0) z_p}{2(\rho_m - \rho_w)}\left[1 - \frac{8}{\pi^2}\sum_{i=0}^{N}\frac{1}{(1+2i)^2}\exp - \frac{\kappa(1+2i)^2\pi^2 t)}{z_p^2}\right], \tag{3}$$

where $\kappa = k/\rho_m C_p$, ρ_m is density of mantle at 0°C, ρ_w is density of the water column, T is temperature of the basal boundary at the ridge axis, $T_0 = 0$°C is surface temperature, and i is an integer whose maximum value is chosen to achieve convergence. Heat flow as a function of time, $H(t)$, is given by

$$H(t) = \frac{k(T - T_0)}{z_p}\left[1 + 2\sum_{i=1}^{N} \exp\frac{-\kappa i^2 \pi^2 t}{z_p^2}\right]. \tag{4}$$

If k, α and C_p vary as a function of pressure and temperature, $w(t)$ is calculated by numerically solving

$$w(t) = z_r + \frac{\rho_b}{\rho_m - \rho_w(t)}\int_0^{z_p}\left[1 - \frac{\rho(0, z')}{\rho(t, z')}\right]\mathrm{d}z', \tag{5}$$

where z' is the Lagrangian depth co-ordinate that contracts as a result of lithospheric cooling, ρ_b is density at the depth of compensation which is defined as the depth down to which conductive cooling occurs (i.e. the shallowest depth where $\rho(t,\, z) = \rho(0,\, z)$), and $\rho_w(t) = 1.028 + 0.00424w(t)$ Mg m^{-3} to account for compressibility of the water column (Grose & Afonso, 2013). ρ_w increases as a function of time since the seabed deepens due to lithospheric cooling, which slightly amplifies the thickness of the water column, $w(t)$.

Heat flow is calculated using a finite-difference scheme which solves

$$H(n\Delta t) = \frac{k_0^n(T_1^n - T_0^n)}{\Delta z_0^n}, \tag{6}$$

where n is an integer, Δt is the time interval between nodes, Δz_0^n is the depth interval between nodes, k_0^n is surface conductivity, T_0^n is surface temperature, and T_1^n is temperature at the next deepest node.

The temperature and pressure dependence of k, α and C_p have been determined using a range of different laboratory techniques which are summarized in detail by Richards et al. (2020). For k, we use the parameterization of Pertermann and Hofmeister (2006) for dunite; for α, we use the parameterization of Bouhifd et al. (1996) presented in their Figure 3 (i.e. not the values from their Table 2); and for C_p, we use the parameterization of Berman (1988), which assumes an olivine composition of 89% forsterite and 11% fayalite.

Finally, the insulating properties of oceanic crust can be included. Following Grose and Afonso (2013) and Richards et al. (2018), a 7 km thick insulating layer at the top of the cooling lithospheric plate is simulated by assuming that plagioclase is the dominant crystalline phase. Since plagioclase has a thermal conductivity which is $\sim$ 25% that of olivine, oceanic crust can have a potentially significant insulating effect upon the underlying lithospheric mantle, provided that heat is tranferred by conduction alone. A geometric mean is used to estimate the conductivity of an aggregate which comprises plagioclase, diopside and olivine (Grose & Afonso, 2013). This assumption yields an average crustal conductivity of 2.65 W m^{-1} °C^{-1} which is broadly consistent with deep-sea drilling results (e.g. 2 W m^{-1} °C^{-1}; Kelemen et al. (2004)).

4 Model Assessment and Benchmarking

Sets of subsidence and heatflow calculations are generated by systematically varying three key parameters: mantle potential temperature, T_p; lithospheric thickness, z_p; and ridge depth, z_r. Thus T_p is varied between 1100 and 1600°C at increments of 25°C,

z_p is varied between 60 and 160 km at increments of 5 km, and z_r is varied between 2 and 3 km at increments of 0.05 km. In order to identify those models that yield optimal fits to subsidence and heatflow observations, the root mean squared (rms) misfit is calculated by carrying out parameter sweeps which interrogate the three-dimensional misfit space. The formal misfit between observed and calculated subsidence, χ_s, is given by

$$\chi_s = \sqrt{\frac{1}{N_s}\sum_{i=1}^{N_s}\left(\frac{w_i^o - w_i^c}{\sigma_i}\right)^2}, \tag{7}$$

where w_i^o and w_i^c are observed and calculated values of water-loaded subsidence, N_s is the number of age bins, and σ_i is the interquartile range of each bin. The misfit between observed and calculated heat flow, χ_h, is given by

$$\chi_h = \sqrt{\frac{1}{N_h}\sum_{i=1}^{N_h}\left(\frac{H_i^o - H_i^c}{\sigma_i^*}\right)^2}, \tag{8}$$

where H_i^o and H_i^c are observed and calculated values of heat flow, N_h is the number of age bins, and σ_i^* is the interquartile range of each bin. Figure 2a shows that there is a noticeable reduction in the number of age-depth measurements older than 168 Ma, which contributes to a greater degree of uncertainty (Figure 2b). Following Richards et al. (2018), age-depth and heatflow measurements from oceanic crust that is older than 168 Ma are excluded to avoid spatial bias. Furthermore, measurements younger than 5 Ma are also excluded omitted to mitigate the potential influence of hydrothermal circulation at the ridge axis (Grose & Afonso, 2013).

Following Stein and Stein (1992), we also interrogate the joint misfit between observed and calculated values of subsidence and heatflow, χ_t. For simplicity, equal weight is given to each set of observations so that

$$\chi_t = \sqrt{\left(\frac{\chi_s^2 + \chi_h^2}{2}\right)}. \tag{9}$$

We wish to revisit a range of both analytical and numerical plate models by using a combination of subsidence modeling, heatflow modeling and joint modeling. The purpose of investigating these different forms of modeling is to probe differences between proposed models in the context of the self-consistency of subsidence and heatflow measurements. In each case, we are interested in locating a global minimum which identifies the optimal values of T_p, z_p and z_r. We are also interested in exploring any obvious trade-offs between these three parameters. Recovered values of these plate parameters are benchmarked against independent geologic constraints. For example, mantle potential temperature has been estimated from petrologic and geochemical studies at mid-oceanic ridges (e.g. 1250–1350°C: Katsura et al. (2004); 1280–1400°C: Herzberg et al. (2007); 1350±50°C: Lee et al. (2009); 1314–1464°C: Dalton et al. (2014); 1318^{+44}_{-32} °C: Matthews et al. (2016)). The midpoint and range which honors the results of these studies is 1333± 19°C

Equilibrium lithospheric thickness is more difficult to independently and accurately determine. Several studies have estimated z_p by calibrating shear wave tomographic models against a plate cooling model which is, to some extent, a circular argument. Steinberger and Becker (2018) determine lithospheric thickness from the results of five earthquake tomographic models by identifying a critical isothermal surface. They obtain a mean thickness of equilibrium oceanic lithosphere of 109 ± 22 km.

Although the zero-age depth at mid-oceanic ridges is determined directly from subsidence observations, its average value is difficult to pinpoint since depths of mid-oceanic ridges vary by ±1 km or more, principally as a consequence of mantle dynamic topography. We include z_r as a quasi-independent parameter because the value obtained by interrogating the joint misfit between observed and calculated subsidence and heat flow should not differ significantly from that obtained by modeling subsidence observations alone. Note that the mean global ridge depth between 0 and 2.5 Ma is 2.95 ± 0.3 km.

5 Results

5.1 Original Analytical Model (HWR-1)

Figure 3 presents the results of an analytical plate model, which adopts the exact parameterization assumed by Parsons and Sclater (1977). In their scheme, the value of mean thermal conductivity, $k = 3.138$ W m^{-1} K^{-1}, is based upon the laboratory experiments of Schatz and Simmons (1972). The value of mean heat capacity, $C_p = 1171.52$ J°C^{-1}, is based upon measurements of typical minerals over a temperature range that is appropriate for oceanic lithosphere (Goranson, 1942). Note that Parsons and Sclater (1977) varied thermal expansivity to minimize the misfit between observed and calculated subsidence from the North Pacific and North Atlantic oceans. They obtain a value of $\alpha = 3.2 \times 10^{-5}$ K^{-1}.

Our modeling strategy is implemented in three stages. First, subsidence observations alone were fitted by systematically varying T, z_p and z_r. There is a weak global minimum at $\chi_s < 0.2$ with a clear negative trade-off between T and z_p such that a thinner, hotter plate or a cooler thicker plate fits the observations equally well (Figure 3c). At the global minimum, $T = 1300$°C, $z_p = 110$ km, and $z_r = 2.6$ km. In comparison, Parsons and Sclater (1977) obtained $T = 1350$°C, $z_p = 125$ km and $z_r = 2.50$ km. Minor differences partly reflect the fact that Parsons and Sclater (1977) exploit 377 age-depth measurements from the North Pacific and North Atlantic oceans whereas our global database contains 10,863 measurements. Comparison of Figure 3a and f shows that the average slope of their 377 measurements is slightly steeper. Although this difference is small, it acts to displace our estimates of T and z_p to lower and higher values, respectively. Secondly, heatflow observations are fitted. In this case, we obtain $T = 1375$°C and $z_p = 81$ km (i.e. a thinner, hotter plate; Figure 3d). Finally, we jointly fit subsidence and heatflow observation which yields $T = 1503$°C, $z_p = 98$ km and $z_r = 2.35$ km at the global minimum of $\chi_t = 0.24$. Stein and Stein (1992) obtained very similar values of $T = 1450$°C, $z_p = 95$ km and $z_r = 2.5$ km with a thermal expansion coefficient of $\alpha = 3.1 \times 10^{-5}$ K^{-1}. Although they exploited different global databases of subsidence and heatflow measurements, the principal reason for this close similarity is the fact that, in both cases, joint fitting of subsidence and heatflow observations has been carried out, which results in a thinner, hotter plate.

Figure 3f and g show the original subsidence and heatflow measurements of Parsons and Sclater (1977). These observations include 377 age-depth measurements and 29 heatflow measurements. Figure 3h demonstrates that the global minimum obtained by fitting subsidence measurements occurs at $T = 1333$°C, $z_p = 125$ km and $z_r = 2.50$ km (i.e. identical to the values recovered by Parsons and Sclater (1977)). However, 3i demonstrates that the global minimum obtained by fitting heatflow observations occurs at $T = 1465$°C and $z_p = 100$ km.

Note that, in this case, misfit was only calculated using heatflow measurements >60 Ma in order to negate the influence of hydrothermal circulation which affects young oceanic crust. Parsons and Sclater (1977) did not carry out this essential filtering step. Similarly, the joint inverse modeling yields $T = 1490$°C and $z_p = 105$ km. This result highlights the fact that combined modeling of subsidence and heatflow measurements tends to gen-

erate an optimal model with an excessively high temperature (see also Stein and Stein (1992). This result is a direct consequence of formally fitting heatflow measurements which Parsons and Sclater (1977) did not do.

In summary, there are two obvious problems that are highlighted by our three-stage analysis (Figure 3). First, parameter values recovered by separately fitting subsidence and heatflow observations are strikingly different— subsidence observations predict a thicker cooler plate and heatflow observations predict a thinner, hotter plate. Secondly, joint fitting of subsidence and heatflow measurements does not yield parameter values that are consistent with independent constraints, particularly with respect to the recovered basal temperature which is nearly 200°C hotter than expected. This exceptionally hot basal temperature is strongly influenced by the fitting of heatflow measurements, which yields a very thin plate (<100 km) for all basal temperatures.

5.2 Revised Analytical Model (HWR-2)

The mismatch between parameter values recovered from both separate and joint modeling of subsidence and heatflow measurements indicates that the original parameterization of Parsons and Sclater (1977) requires modification. Here, we suggest that this mismatch arises because subsidence is sensitive to the average value of thermal conductivity that is obtained by integration over the total thickness of the cooling lithosphere whereas heat flow is sensitive to the value of thermal conductivity at the top of the lithosphere (compare Equations 3 and 4).

The value of k assumed by Parsons and Sclater (1977) is 3.138 W m^{-1} K^{-1}. It is a mean value for the upper 120 km of the mantle (Schatz & Simmons, 1972) which is therefore relevant to subsidence observations. In order to identify an appropriate value of k at the top of the plate that is relevant to heatflow observations, its value is varied between 3.20 and 4.10 W m^{-1} K^{-1} at intervals of 0.02 W m^{-1} K^{-1}. For k = 3.90 W m^{-1} K^{-1}, a global minimum of $\chi_t = 0.22$, obtained by jointly fitting subsidence and heatflow observations, is located at T = 1326°C, z_p = 105 km and z_r = 2.60 km (Figure 4). In contrast to the HWR-1 plate model, the recovered value of T is now in much closer agreement with that determined from independent petrologic and geochemical studies (i.e. T_p=1333±19°C; Figure 4e). The recovered value of lithospheric thickness, z_p = 105 km, is also consistent with independent estimates of equilibrium lithospheric thickness (z_p = 109±22 km; Steinberger and Becker (2018)). The recovered ridge depth, z_r = 2.60 km, is marginally shallower than the mean observed ridge depth (2.95 ± 0.3 km).

Our justification for choosing a higher value of k for the top of the lithospheric plate is that modern seismic reflection images of oceanic crust and the uppermost mantle directly beneath the Moho show that faulting and fracturing occur which facilitates fluid circulation within the top ten km. Therefore heat transport through the top of oceanic lithosphere is unlikely to be purely conductive. A significant (but unknown) component of advective heat transport can be simulated by using a higher value of k.

In order to sidestep any possible effects of efficient hydrothermal circulation in proximity to the ridge axis, we followed the approach of Richards et al. (2018) who excluded subsidence and heatflow observations from seafloor younger than 5 Ma. In order to test the effects that exclusion of measurements from the youngest lithosphere has upon our results, we have also individually and jointly modeled the original (i.e. unedited) subsidence and heatflow observations. Our results are almost exactly the same— joint modeling yields a global minimum located at T = 1325°C, z_p = 105 km and z_r = 2.60 km which differs by one degree (compare, for example, panels e and h of Figure 4). We confidently conclude that removal of measurements from seafloor that is younger than 5 Ma, for whatever reason, has little discernible influence upon the recovered plate cooling model.

5.3 Temperature- and Pressure-dependent Model (HWR-3)

Notwithstanding the fact that the HWR-2 analytical model yields an excellent fit between observed and calculated values of subsidence and heat flow, we acknowledge that there is considerable interest in developing more complex (i.e. physically realistic) plate models. The starting point concerns the results of laboratory experiments, which clearly demonstrate that α, k and C_p vary as a function of temperature and pressure (Figure 5; McKenzie et al. (2005); Richards et al. (2018)). Here, we apply with minor changes the temperature- and pressure-dependent parameterization implemented Richards et al. (2018) which is a revised version of earlier parameterizations developed by McKenzie et al. (2005) and by Grose and Afonso (2013). Our minor changes are as follows— the temperature-dependence of k is based on the lattice conductivity and excludes the radiative component; the temperature-dependence of α is taken from Figure 3 of Bouhifd et al. (1996) rather than from their Table 2 (values quoted in table and shown in figure are mutually inconsistent); and the parameterization of C_p is chosen so that olivine consists of 89% forsterite and 11% fayalite (Berman, 1988). Figure 5 is a schematic summary of the chosen parameterization, the details of which are described in considerable detail by Richards et al. (2018). Oceanic crust is assumed to act as an insulating layer (Grose & Afonso, 2013).

Our results are presented in Figure 6. The optimal model obtained by jointly fitting subsidence and heatflow observations is located at $T = 1326°$C, $z_p = 96$ km and $z_r = 2.80$ km where $\chi_t = 0.19$. This result is in close agreement with that obtained using the HWR-2 analytical model. Note that the decrease in the value of χ_t statistically insignificant, that the decrease in equilibrium plate thickness from 105 km to 96 km is minor, and that there is no change in the value of T. The recovered value of z_r is within one standard deviation of the observed mean ridge depth of 2.95 ± 0.3 km.

We conclude that the revised and augmented databases of subsidence and heatflow measurements are equally and optimally fitted by either the HWR-2 model or the HWR-3 model. Our preference is for the HWR-2 model on three significant grounds. First, this analytical model is the simplest and most parsimonious one that yields the best fit to both subsidence and heatflow observations. Secondly, even though α, k and C_p vary as a function of temperature and to a lesser extent pressure, both plate subsidence and heat flow principally depend upon the integrated averages of these parameters over plate thickness. Thus the observations are insensitive to the functional variation of each parameter, the only exception being the value of k at the top of the plate which controls surface heat flow. Finally, the results of many different laboratory experiments are often problematic— many of these experiments have been carried out upon single crystals or aggregates which are not necessarily representative of lithospheric mantle and there are considerable differences between the results of different laboratories as well significant uncertainties (see Figure 7 in Richards et al. (2018) for a detailed summary of different temperature- and pressure- dependent laboratory experiments carried out upon α, k and C_p). One notable problem concerns the variation of k as a function of T (Figure 5a). Results obtained using either the contact method or laser flash analysis both show that k significantly increases at low temperatures (i.e. 5.5–7.0 W m^{-1} K^{-1} at room temperature). This increase dramatically perturbs the recovered value of basal temperature in particular, pushing it outside of the independently accepted range and requiring the inclusion of a low conductivity crust (see Figure 7a of Richards et al. (2018) for further details). One way of resolving these experimentally related issues is to carry out improved laboratory analyses on more realistic aggregates.

6 Discussion

Reference models for cooling and subsidence of oceanic lithosphere have necessarily evolved as observational constraints improve and methodologies become more sophis-

ticated. In Table 2, the databases exploited by key reference models together with values of assumed and recovered parameters are listed. Since the seminal analysis of Parsons and Sclater (1977), it is evident that increasing amounts of subsidence and heatflow measurements have been used to calculate optimal plate cooling models. Figure 7a summarizes basal temperature estimates and equilibrium lithospheric thicknesses predicted by the reference models listed in Table 2. Both the HWR-1 and Stein and Stein (1992) model yield basal temperatures that significantly exceed independent petrologic estimates (Figure 7b). We contend that this discrepancy is a direct consequence of joint inverse modeling where the value of k is fixed. Richards et al. (2018) obtained an equilibrium plate thickness of 136 km with a basal temperature of 1302°C. We now believe that this plate thickness is too thick, which is the direct consequence of two factors. First, the isostatic sedimentary correction employed by Hoggard et al. (2017) gives rise to a modest overprediction of subsidence for old plate ages. A revised sedimentary correction, which assumes a smaller (i.e. siltier) value of solid grain velocity, has been proposed and implemented by Holdt et al. (2022). This minor adjustment slightly decreases the gradient of water-loaded subsidence for plate ages > 100 Ma which in turn yields a thinner equilibrium plate thickness. This relatively minor adjustment highlights the importance of accurate age-depth measurements. Secondly, Richards et al. (2018) include heatflow measurements for plate ages of 0–25 Ma. However, many of these measurements are from the Gulf of Aden where heat flow is anomalously elevated due to the presence of a mantle plume. By fitting these anomalous heatflow values, the optimal plate model must include a steep gradient to match older heatflow measurements which necessarily yields a much thicker plate. Here, we exclude heatflow measurements from the Gulf of Aden. Figure 7 shows that several plate models yield acceptable values of basal temperature. Of these, we favor the HWR-2 model since it is the simplest model which best fits the most substantial observational databases.

Two additional points of discussion arise from our reanalysis of global plate cooling. The first point concerns the existence of transient age-depth shallowing. The second point concerns the relationship between the thermal structure of oceanic lithosphere and the depth distribution of intraplate earthquakes. The reference model of Crosby et al. (2006) predicts transient shallowing of ∼250 m for plate ages of 80–130 Ma (Figure 8a). This reference model is generated by using an empirical fit to water-loaded age-depth observations from the Pacific Ocean, although it is important to note that they report similar shallowing in four other regions of the oceanic realm. Figure 8b demonstrates that the transient shallowing observed by Crosby et al. (2006) is indeed real— it is clearly visible when the global subsidence database of Holdt et al. (2022) is sampled within an identical region of the Pacific Ocean (Figure 9a). The amplitude and wavelength of the recovered shallowing is almost identical (compare Figure 8a and b). Note that the database of Stein and Stein (1992), which includes age-depth measurements from the North Pacific Ocean, also shows transient shallowing between ∼100 and 130 Ma (see their Figure 1b). However, on a global scale, no transient shallowing is observed (see Figure 2b).

Figure 9 demonstrates that, by limiting their detailed study to the Pacific Ocean, Crosby et al. (2006) sampled positive residual depth anomalies in the vicinity of various hotspot tracks within this 80–130 Ma window. Furthermore, residual depth anomalies of the Pacific Plate between 0 and 80 Ma are generally all slightly negative, which tends to exacerbate the shallowing trend for older ages Figure 9c. In this specific case, shallowing of this region becomes emergent because only one anomalous flank of the cooling plate has been sampled. Several other regions were also used by Crosby et al. (2006) to demonstrate transient shallowing. Figure 9 demonstrates that these smaller regions suffer from the same age-depth sampling pathology, giving rise to apparent shallowing between 80 and 130 Ma. These different regions all feature mantle plumes: the Bermuda Swell of the NW Atlantic Ocean; the Cape Verde, Madeira and New England swells of the NE Atlantic Ocean; and the Angolan and Vema swells of the SW Atlantic Ocean. Figure 9c also demonstrates that these regions do not include significant downwellings

which lie within the 80–130 Ma window (e.g. Argentine Abyssal Plain, Gulf of Mexico, Bay of Bengal).

Although long wavelength temporary shallowing is not supported by the revised and augmented global subsidence observations of Holdt et al. (2022), an overall flattening from approximately 80 Ma onward is observed, which tends to a value of 6 km. This flattening is matched by a flattening of heatflow measurements which tend towards 0.05 W m^{-2} at old plate ages. Parsons and McKenzie (1978) suggest that oceanic lithosphere consists of mechanical and thermal boundary layers. As the plate cools, both of these layers thicken with age. From a fluid dynamical perspective, the thermal boundary layer should become unstable and convectively overturn for plate ages that exceed about 60 Ma. The Rayleigh-Taylor instability acts to stabilise lithospheric thickness. We suggest that this instability occurs on smaller length scales and on shorter timescales than hitherto expected, such that it lies within scatter produced by residual depth anomalies.

We end this contribution by addressing the rheological structure of cooling oceanic lithosphere. Figure 10a and b show the evolving thermal structure of oceanic lithosphere calculated for the HWR-2 and HWR-3 models, respectively. In both cases, the distribution of intraplate and outer rise earthquakes analyzed by Craig et al. (2014) occur at temperatures of $< 700°$ C in broad agreement with some, but not all, previous studies (McKenzie et al., 2005; Craig et al., 2014; Richards et al., 2018). This revised estimate is now consistent with laboratory creep experiments carried out upon dry polycrystalline olivine aggregates (Goetze & Evans, 1979; Boettcher et al., 2007).

7 Conclusions

In this study, a revised and augmented global database of accurate age-depth measurements is combined with an existing database of filtered heatflow measurements to revisit the well-known plate cooling model. Our primary goal is to identify the simplest plate model that yields the smallest misfit between combined age-depth and heatflow measurements. In order to accomplish this task, we have explored a variety of analytical and numerical plate models which respectively assume that thermal conductivity, thermal expansivity and specific heat capacity are constant or vary as a function of temperature and pressure using experimental constraints.

The optimal analytical model yields a basal temperature of 1326 ± 50°C and an equilibrium plate thickness of 105 ± 10 km for a zero-age depth of 2.60 ± 0.3 km. This value of basal temperature is consistent with independent constraints. A more sophisticated numerical model fits subsidence and heatflow observations equally well and yields almost identical values of basal temperature and equilibrium plate thickness. Spatial analysis of our global database of age-depth measurements shows that previously described transient shallowing for plate ages of 80–130 Ma is a product of geographic sampling. This lack of shallowing suggests that the Rayleigh-Taylor instability, which is thought to maintain constant plate thickness, probably acts on shorter timescales and smaller length scales. Finally, our revised plate model can be used to calculate the thermal structure of the lithosphere as a function of time, to estimate that the maximum depth of intraplate earthquakes is bound by the 700° C isothermal surface, and to accurately estimate residual depth anomalies which represent maximum observable mantle dynamic topography.

Table 1. Symbols and Values of Parameters.

	Symbol	Parameter	Value	Units
Thermal parameters	α	Thermal expansivity	-	K^{-1}
	C_p	Heat capacity	-	$J°C^{-1}$
	k	Thermal conductivity	-	$W\ m^{-1}\ K^{-1}$
Recovered parameters	T	Mantle temperature	-	°C
	T_b	Mantle basal temperature	-	°C
	T_p	Mantle potential temperature	-	°C
	z_p	Equilibrium lithospheric thickness	-	km
	z_r	Water-loaded ridge depth	-	km
Observed subsidence	w	Water-loaded basement depth	-	km
	C_c	Crustal correction	-	km
	C_s	Sediment correction	-	km
	z_{sb}	Depth of seabed	-	km
Observed heat flow	H	Heat flow	-	$W\ m^{-2}$
	$\Delta T/\Delta z$	Thermal gradient	-	°C / m
	F	Perturbation of geotherm	-	dimensionless
	κ_s	Thermal diffusivity of sediment	0.25	$mm^2\ s^{-1}$
	t_s	Duration of sedimentation	-	Ma
	U	Sedimentation rate	-	$km\ Ma^{-1}$
General	t	Time	-	Ma
	z	Depth	-	km
	T	Temperature	-	°C
	P	Pressure	-	MPa
	X	Composition	-	
	ρ	Density	-	$Mg\ m^{-2}$
Calculated subsidence; calculated heat flow	ρ_m	Density of mantle at 0°C	3.30	$Mg\ m^{-2}$
	ρ_w	Density of water	1.03	$Mg\ m^{-2}$
	T_0	Surface temperature	0	°C
	κ	Thermal diffusivity	-	$mm^2\ s^{-1}$
	ρ_b	Density at depth of compensation	-	$Mg\ m^{-2}$
	Δt	Time interval between nodes	-	Ma
	Δz_0^n	Depth interval between nodes	-	km
	k_0^n	Surface conductivity	-	$W\ m^{-1}\ K^{-1}$
	T_0^n	Surface temperature	-	°C
	T_1^n	Temperature at next deepest node	-	°C
Misfit calculation	χ_s	Subsidence misfit	-	dimensionless
	χ_h	Heat flow misfit	-	dimensionless
	χ_t	Joint misfit	-	dimensionless
	w_i^o	Observed subsidence		km
	w_i^c	Calculated subsidence		km
	N_s	Number of subsidence age bins	65	dimensionless
	σ_i	Interquartile range of subsidence age bin		km
	H_i^o	Observed heat flow	-	$W\ m^{-2}$
	H_i^c	Calculated heat flow	-	$W\ m^{-2}$
	N_h	Number of heat flow age bins	65	dimensionless
	σ_i^*	Interquartile range of heat flow age bin		$W\ m^{-2}$

Table 2. Reference models for evolution of oceanic lithosphere. Sub. = subsidence measurements. HF = heatflow measurements. Pac. = Pacific Ocean. Atl. = Atlantic Ocean. V = variable parameters. Bold values denote recovered parameters.

Name	Reference	Location	Sub.	HF	k (W m^{-1}K^{-1})	α (x10^{-5}K^{-1})	C_p (J°C^{-1})	T (°C)	z_p (km)	z_r (km)
PSM	Parsons and Sclater (1977)	N.Pac., N.Atl.	377	29	3.138	**3.20**	1171.52	**1350**	**125**	2.50
GDH1	Stein and Stein (1992)	N.Pac., NW.Atl.	Grid	5,539	3.138	**3.10**	1171.52	**1450**	**95**	2.60
CHABLIS	Doin and Fleitout (1996)	Global	Grid	853	3.100	**4.20**	1124.00	**1310**	**105**	2.50
HW05	Hillier and Watts (2005)	N.Pacific	Grid	Grid	3.370	**2.77**	1171.52	**1363**	**120**	3.01
MJP05	McKenzie et al. (2005)	N/A	N/A	N/A	V	V	V	1315	**106**	2.50
CMS06	Crosby et al. (2006)	N.Pacific	Grid	N/A	V	V	V	1315	90	2.60
CM09	Crosby and McKenzie (2009)	Global	Grid	N/A	V	V	V	N/A	90	2.60
RHWC18	Richards et al. (2018)	Global	2,028	3,597	V	V	V	**1302**	**136**	**2.64**
HWR-1	This study	Global	10,863	3,573	3.138	3.20	1171.52	**1503**	**98**	**2.40**
HWR-2	This study	Global	10,863	3,573	3.138	3.20/3.90	1171.52	**1326**	**105**	**2.60**
HWR-3	This study	Global	10,863	3,573	V	V	V	**1326**	**96**	**2.80**

Figure 1. Global databases. (a) Distribution of water-loaded depth to oceanic basement measurements, averaged within 1° bins (Holdt et al., 2022). Circles = accurate measurements obtained by applying both sedimentary and crustal corrections; upward/downward pointing triangles = lower/upper estimates obtained by applying sedimentary correction and regional crustal correction. (b) Distribution of heatflow measurements (Hasterok et al., 2011). Colored/gray circles = measurements that pass/fail filtering process. (c) Augmented age grid of Seton et al. (2020) that includes modifications based upon regional studies compiled by Hoggard et al. (2017) and Holdt et al. (2022).

Figure 2. Subsidence and heatflow measurements as function of oceanic plate age. (a) Water-loaded measurements of depth to oceanic basement plotted as function of plate age and averaged within 1° bins (Holdt et al., 2022). See Figure 1a for global distribution. (b) Same measurements binned every 2.5 Ma. Horizontal bars with vertical gray boxes = median values with interquartile ranges. (c) Filtered and corrected heatflow measurements as function of plate age (Hasterok et al., 2011). See Figure 1b for global distribution. (d) Same measurements binned every 2.5 Ma. Horizontal bars with vertical gray boxes = median values with interquartile ranges.

Figure 3. *Caption overleaf.*

Figure 3. Analytical plate model using Parsons and Sclater (1977) parameterization (HWR-1).
(a) Horizontal bars with vertical gray boxes = median values with interquartile ranges of age-depth measurements; red line = optimal fit between observed and calculated subsidence obtained by jointly fitting age-depth and heatflow measurements where $T = 1503°$C; $z_p = 98$ km; $z_r = 2.40$ km. (b) Horizontal bars with vertical gray boxes = median values with interquartile ranges of heatflow measurements; red line = optimal fit between observed and calculated heat flow obtained by jointly fitting age-depth and heatflow measurements. (c) Misfit between observed and calculated subsidence, χ_s, plotted as function of basal temperature and plate thickness for $z_r = 2.60$ km. Black cross = locus of global minimum; red cross = locus of global minimum shown in panel c; open blue circle = locus of global minimum obtained by Parsons and Sclater (1977); pink line = loci of minima when basal temperature is fixed at $1333\pm19°$C. (d) Misfit between observed and calculated heat flow, χ_h. (e) Joint misfit between observed and calculated subsidence and heat flow, χ_t, for $z_r = 2.40$ km. Red cross = global minimum at $T = 1503°$C; $z_p = 98$ km; $z_r = 2.40$ km; Blue circle = locus of global minimum obtained by Stein and Stein (1992). (f) Black circles = original age-depth measurements of Parsons and Sclater (1977) from North Pacific and North Atlantic Oceans; red line = optimal fit between observed and calculated subsidence obtained by jointly fitting subsidence and heatflow measurements of Parsons and Sclater (1977) where $T = 1490°$C; $z_p = 105$ km; $z_r = 2.40$ km; black line = optimal fit between observed and calculated subsidence obtained by Parsons and Sclater (1977) where $T = 1350°$C; $z_p = 125$ km; $z_r = 2.50$ km. (g) Black circles = original heatflow measurements of Parsons and Sclater (1977); red line = optimal fit between observed and calculated heat flow obtained by jointly fitting subsidence and heatflow measurements of Parsons and Sclater (1977) where $T = 1490°$C, $z_p = 105$ km and $z_r = 2.40$ km. (h)–(j) Same as panels (c)–(e) where misfit is calculated using subsidence and heatflow measurements shown in panels (f) and (g).

Figure 4. Revised analytical plate model (HWR-2). (a) Horizontal bars with vertical gray boxes = median values with interquartile ranges of age-depth measurements; red line = optimal fit between observed and calculated subsidence obtained by jointly fitting age-depth and heatflow measurements where $T = 1326°$C; $z_p = 105$ km; $z_r = 2.60$ km. (b) Horizontal bars with vertical gray boxes = median values with interquartile ranges of heatflow measurements; red line = optimal fit between observed and calculated heat flow obtained by jointly fitting age-depth and heatflow measurements. (c) Misfit between observed and calculated subsidence, χ_s, plotted as function of basal temperature and plate thickness for $z_r = 2.60$ km. Black cross = locus of global minimum; red cross = locus of global minimum shown in panel c; pink line = loci of minima when basal temperature is fixed at $1333\pm19°$C. Note that subsidence measurements between 0 and 5 Ma are excluded. (d) Misfit between observed and calculated heat flow, χ_h. Note that heatflow measurements between 0 and 5 Ma are excluded. (e) Joint misfit between observed and calculated subsidence and heat flow, χ_t, for $z_r = 2.60$ km. (f)–(h) Same as panels (c)–(e) except that both subsidence and heatflow measurements between 0 and 5 Ma are now included. Note that locus of global minimum obtained by jointly fitting subsidence and heatflow measurements shifts very slightly to $T = 1325°$C; $z_p = 105$ km; $z_r = 2.60$ km.

Figure 5. Plate model parameterization. (a) Thermal conductivity plotted as function of temperature. Solid/dashed black lines = forsterite lattice conductivity measurements $\pm 1\sigma$ at constant pressure of 0.1 MPa, redrawn from Pertermann and Hofmeister (2006); red line = constant thermal conductivity assumed for HWR-1 and HWR-2 (subsidence only) models; red circle = constant thermal conductivity assumed for HWR-2 model (heat flow only); black circle = mean thermal conductivity value calculated between 0 and 1300°C, based upon parameterization of Pertermann and Hofmeister (2006). (b) Thermal expansivity plotted as function of temperature. Solid/dashed black lines = thermal expansivity measurements $\pm 1\sigma$ for constant pressure of 0.1 MPa, taken from Figure 3 of Bouhifd et al. (1996); red line = constant thermal expansivity assumed for HWR-1 and HWR-2 models; black circle = mean thermal expansivity value calculated between 0 and 1300°C, based upon parameterization of Bouhifd et al. (1996). (c) Heat capacity plotted as function of temperature. Solid/dashed black lines = heat capacity measurements $\pm 1\sigma$ using parameterization of Berman (1988), which assumes 89% forsterite and 11% fayalite; red line = constant heat capacity assumed for HWR-1 and HWR-2 models; black circle = mean heat capacity value calculated between 0 and 1300°C, based upon parameterization of Berman (1988).

Figure 6. Plate model with temperature- and pressure-dependent parameters (HWR-3). (a) Horizontal bars with vertical gray boxes = median values with interquartile ranges of age-depth measurements; red line = optimal fit between observed and calculated subsidence obtained by jointly fitting age-depth and heatflow measurements where $T = 1326°$C; $z_p = 96$ km; $z_r = 2.80$ km. (b) Horizontal bars with vertical gray boxes = median values with interquartile ranges of heatflow measurements; red line = optimal fit between observed and calculated heat flow obtained by jointly fitting age-depth and heatflow measurements. (c) Misfit between observed and calculated subsidence, χ_s, plotted as function of basal temperature and plate thickness for $z_r = 3.00$ km. Black cross = locus of global minimum; red cross = locus of global minimum shown in panel e; pink line = loci of minima when basal temperature is fixed at $1333\pm19°$C. (d) Misfit between observed and calculated heat flow, χ_h. (e) Joint misfit between observed and calculated subsidence and heat flow, χ_t, for $z_r = 2.80$ km.

Figure 7. Temperature and plate thickness estimates for range of reference models. (a) Joint misfit between observed and calculated subsidence and heat flow plotted for revised analytical plate model HWR-2 as function of basal temperature and plate thickness (see Figure 4e). Open/filled red circles = HWR-1/HWR-2 models; filled red square = HWR-3 model; open black circle = Parsons and Sclater (1977) model; filled black circle = Stein and Stein (1992) model; black cross = CHABLIS model by Doin and Fleitout (1996); open black square = Hillier and Watts (2005) model; open black triangle = McKenzie et al. (2005) model; filled black triangle = Crosby et al. (2006) model; filled black square = Richards (2018) model; vertical pink band = range of mantle potential temperatures from independent petrologic/geochemical studies shown in panel b. (b) Independent estimates of mantle potential temperature. K04 = 1250–1350°C (Katsura et al., 2004); H07 = 1280–1400°C (Herzberg et al., 2007); L09 = 1350±50°C (Lee et al., 2009); D14 = 1314–1464°C (Dalton et al., 2014); and M16 = 1318^{+44}_{-32} °C (Matthews et al., 2016).

Figure 8. Analysis of Crosby et al. (2006) observations. (a) Age-depth measurements as function of oceanic plate age from portion of Pacific Ocean analyzed by Crosby et al. (2006). Horizontal bars with vertical gray boxes = median values with interquartile ranges of age-depth measurements binned every 10 Ma. Note plate shallowing centered on 100 Ma. (b) Age-depth measurements as function of oceanic plate age extracted from revised and augmented database of Holdt et al. (2022). Gray points = age-depth measurements from same portion of Pacific Ocean (see Pacific box of Figure 9 for location). Horizontal bars with vertical gray boxes = median values with interquartile ranges of age-depth measurements binned every 10 Ma. Note recovery of plate shallowing centered on 100 Ma.

Figure 9. Regions selected in plate model study of Crosby et al. (2006). (a) Global distribution of residual depth measurements averaged over 1° bins. Circles = measurements with both sedimentary and crustal corrections; triangles = measurements for which only sedimentary corrections are applied. Red polygons = large igneous provinces and seamounts Coffin et al. (2006). Black box with thick outline = region of Pacific Ocean used to generate reference model by Crosby et al. (2006). Black boxes with thin outline = other regions where transient shallowing is observed. (b) Augmented and revised age grid of Seton et al. (2020). Dashed contours = 80 and 130 Ma isochrons. (c) Spherical harmonic representation of residual depth measurements where $l_{max} = 40$. Transparent gray fill = oceanic crust < 80 Ma and > 130 Ma (i.e. region without transient shallowing observed by Crosby et al. (2006)). Red/blue circles = mantle plumes/drawdowns.

Figure 10. Thermal structure of oceanic lithosphere. (a) Thermal structure of oceanic lithosphere calculated using HWR-2 model. Contour lines = isothermal surfaces; green/white circles with errors = oceanic intraplate/outer rise earthquakes taken from Craig et al. (2014) where small, medium and large circles = M_b <5.5, 5.5–6.5, and >6.5; bold contour line = 700°C isothermal surface, which identifies temperature above which oceanic earthquakes occur. (b) Same for HWR-3 model.

Open Research Section

The data described in this study will be made available on Mendeley Data upon publication of the manuscript.

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