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ISOSTATIC CALCULATION

We have characterized the thermal structure of the lithosphere by means of combined thermal data with local isostasy equilibrium. The continental geotherm in the region has been approximated by assuming steady-state conditions and that the base of the lithosphere is defined by the 1350ºC isotherm, disregarding any fluid circulation. According to heat conduction theory we determine the distribution of temperature in depth by means of the surface heat flow (q_0) and surface temperature (T_0):

$$T = T_0 + \frac{q_0}{k} y - \frac{A}{2k} y^2$$

where A is the radiogenic heat production and k is thermal conductivity. Unfortunately, no further transient thermal modeling can be done due to the scarcity of constraints on the Neogene history of the lithosphere.

Surface heat flow is the sum of incoming heat flow from the asthenosphere and that produced by radiogenic heat production. From the point of view of modeling, the asthenospheric heat flow is controlled by the structure, composition and thickness of the continental lithosphere. The radiogenic heat production mostly comes from the upper crust, where it is commonly assumed to be an exponential decay (e.g., Chapman and Furlong, 1992):

$$0 = k \frac{d^2T}{dy^2} + \rho H_0 e^{-y/h_r}$$

Although we accept this model as a general and smooth approximation to radiogenic heat production in the crust, we are aware that some other models assess that heat production may be related to composition and not to depth suggesting a pattern of radioelement distribution that is inconsistent with the downward-decreasing exponential function predicted from modeling of surface heat flow data (e.g., Mareschal and Jaupart, 2004).

To better constraint the continental geotherm we have used a non-linear relationship between absolute elevation, surface heat flow, crustal and lithospheric mantle thickness (Lachenbruch and Morgan, 1990). We used heat flow, crustal thickness and typical thermal parameters for the region (Table DR1) according to Fernàndez et al. (1998) and Soto et al. (2003). Theoretical elevation and lithospheric thickness are iteratively calculated assuming Airy local isostasy. Once calculated elevation fit observed topography, we assess lithospheric structure and validate thermal modeling. With this procedure, we are aware that any possible contribution of dynamic topography to net elevation in relation to deep density heterogeneities is disregarded.

Absolute elevation (E) in each lithosphere column has been calculated following Lachenbruch and Morgan (1990). It is compared the buoyancy force of a lithospheric column with that of a reference column at the mid-oceanic ridges where average elevation and lithospheric structure are well known. The elevation is calculated as:

$$E = \frac{\rho_a - \rho_L}{\rho_a} L - H_0$$

where ρ_L is the lithospheric density (kg m^{-3}), ρ_w water density (1050 kg m^{-3}), ρ_a asthenospheric density (taken as 3200 kg m^{-3}), and H_0 a calibration constant (−2400 m). The calculation of absolute topography requires implicitly that the asthenospheric density is the same as underneath mid-ocean ridges. Lithosphere thickness (L) is taken as the depth where T= 1350 ºC.

Once the depth of the lithosphere is established, we modeled the effects of an instantaneous removal of the different thicknesses of eroded sediments in each sub-basin and the subsequent uplift.

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TABLE DR1. THERMAL MODELING PARAMETERS

	Heat Production ($mW\ m^{-3}$)	Thermal Conductivity ($W\ m^{-1}\ K^{-1}$)	Density ($kg\ m^{-3}$)
Sediments	1	2.1	2000
Crust	2.1 exp(-z/12000)	2.5	2800
Lithosphere mantle	0	3.2	*

* Mantle density is temperature dependent, being equal to 3200 $[1+3.8\times10^{-5}\ (T_a-T(z))]$ and T_a=1350 ºC (Lachenbruch and Morgan, 1990).