Approaches to the relatively hot Altiplano plateau

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INTRODUCTION

The Altiplano-Puna Plateau is characterized by a heat flow of 78 ± 28 mWm^{-2} when geochemical estimates of heat flow are not taken into account (Hamza & Muñoz, 1996). This variation in heat flow has large consequences in the temperature distribution in a crust of about 60 – 70 km thickness between 17 °S and 24 °S (Götze et al., 1994; Wigger et al., 1994; Beck & Zandt, 2002). Several models have been forwarded for explaining thickening and warming of the Altiplano assuming a weak temperature dependence of thermal conductivity that is contrary to experimental results. In this contribution, some geotherms where temperature-pressure dependence of conductivity based on improved experimental observations has been taken into account are presented and discussed.

GEOTHERMS, STRENGTH OF THE LITHOSPHERE AND ZONES OF PARTIAL MELTING

Steady state continental geotherms where temperature-pressure dependence of conductivity has been considered are presented in Fig. 1 for surface heat flow of 65, 70 and 80 mWm^{-2}. These geotherms are useful for comparison with the evolution of the thermal equilibration of the lithosphere resulting from models and –in conjuction with these models– for discussing other geophysical observations. The geotherms are calculated by an iteration algorithm that follows the variations of conductivity and heat generation in the crust and upper mantle. Conductivity at ambient conditions of upper crustal rocks is taken to be 3.0–3.3 Wm^{-1}°C^{-1}; for the lower crust a value of 2.6 Wm^{-1}°C^{-1} is used. In Fig. 1, for geotherms 1, 3 and 4, the temperature dependence of thermal conductivity in the crust follows the results of Zoth & Haenel (1988) as described by Seipold (1998), and for geotherm 2 the results obtained by Sass et al. (1992) are used. The pressure dependence of conductivity in the crust is taken from Chapman & Furlong (1992). In the upper mantle, thermal and pressure effects on conductivity are also taken into account, following a model based on phonon lifetimes from infrared reflectivity (Hofmeister, 1999). In 1, 2 and 4, the conductivity at 20-30 km depth is 2.2-2.0 Wm^{-1}°C^{-1}, and 2.5-2.3 Wm^{-1}°C^{-1} in geotherm 3; the conductivity decreases from these values to about 1.5 Wm^{-1}°C^{-1} at the crust/mantle boundary. In the mantle lithosphere, the conductivity results in values of about 2.9-3.0 Wm^{-1}°C^{-1}. Radiogenic heat generation (A) in a layered crust for geotherms 1, 2 and 4 is taken to be 1.0-1.5, 0.6-0.8 and 0.4-0.5 μWm^{-3} for the upper, middle and lower crust, respectively; in geotherm 3, a first layer with a step variation in A giving a weighted value of 1.3 μWm^{-3} and a mafic lower crustal layer 8 km thick with A= 0.2 μWm^{-3} have been assumed. At a depth of 20 km, temperature ranges from about 420 °C to 550 °C; at 30 km depth, it ranges from about 580 °C to 750 °C, and at the crust/mantle boundary (CMB) it can reach values from about 1050 °C to 1200 °C. The radiogenic heat generation in the upper mantle is taken to be 0.02 μWm^{-3}, and the thermal boundary layer at 1250 °C is reached at 90-130 km depth in geotherms 1, 2, and 4 and at 85 km depth in geotherm 3. Heat flow from the mantle results in values of 8-10 mWm^{-2} for 1, 2 and 4, and of 21 mWm^{-2} for geotherm 3.

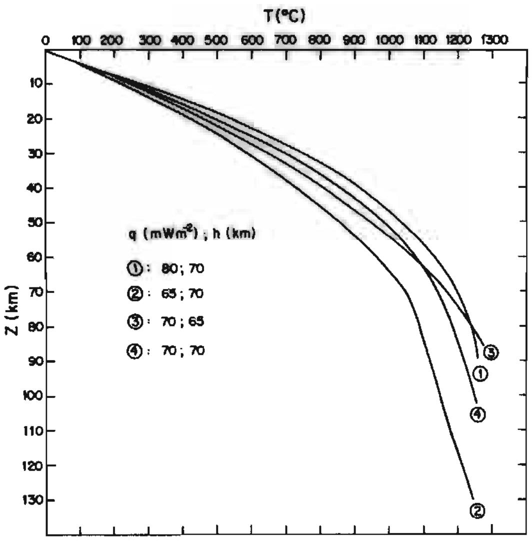

Figure 1. Steady state geotherms for surface heat flow (q) and crustal thickness (h) with temperature-pressure dependence of thermal conductivity.

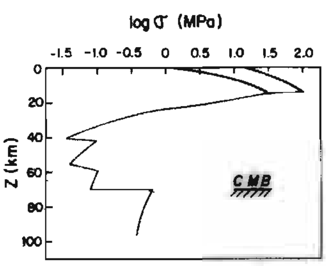

Figure 2. Strength of the lithosphere beneath the Altiplano for a layered crust. Upper thick curves indicate the domain of the brittle regime (*left*: extension; *right:* compression). The thin curve shows the domain of the ductile regime. CMB is the crust/mantle boundary.

Very high surface heat flow in the Altiplano (100-180 mWm^{-2}) can be seen as the effect of near-surface effects as also observed for the Tibetan plateau (Francheteau et al., 1984). If the presently observed surface heat flow of 65-80 mWm^{-2} in the Altiplano is due to the thermal relaxation of a transient geotherm in 30-50 My, crustal temperatures should be decreased from the former values by some few tens of degrees at 20-30 km depth and by about 100-250 °C at the crust/mantle boundary. In this case, temperature at 20 km depth could range between about 400 °C and 500 °C, and between 550 °C and 700 °C at 30 km depth. Taking into account the results of Lebedev & Khitarov (1964) and of Liu et al. (2001), this implies that partial hydrous melting of granitic rocks is possible below 25 km in some areas of the Altiplano, which is in satisfactory agreeement with a crustal electrical conductor obtained in a magnetotelluric study carried out in the southern Altiplano (Brasse et al., 2002) and with depths of seismic low-velocity zones detected beneath the plateau (Beck & Zandt, 2002; *ANCORP Working Group*, 2003). Temperature at the crust/mantle boundary could range between 800 °C and 1000 °C, with the largest temperature possibly representing the thermal state in the deeper crust of the southern Altiplano at about 21 °S where a strong electrical conductivity anomaly has been observed (Brasse et al., 2002). To the north (17 °S-18 °S) this anomaly is not encountered (Brasse, 2004); beneath an adjacent terrane the observed seismic velocities are most consistent with a felsic composition and temperatures between 700 °C and 800 °C near the base of the crust (Beck & Zandt, 2002). A low integrated strength of the lithosphere is obtained in any case (Fig. 2), ranging generally between 5.0×10^{11} Nm^{-1} and 8.0×10^{11} Nm^{-1}, and not larger than 3.0×10^{12} Nm^{-1}. The crust below 15-20 km is in the ductile regime.

A different thermal parametrization of the crust has led to a hotter upper and middle crust when compared with results obtained by Springer (1999) where also a high thermal conductivity (4.4 $Wm^{-1}°C^{-1}$) was assumed for subcrustal rocks. In numerical models for the thermal evolution of the Altiplano described by Babeyko et al. (2002) it is suggested that neither radiogenic heat production in a thickening crust, nor heating due to other processes as shear during deformation or intrusion of arc magmas into the middle crust, can heat the middle crust to the degree and within the time suggested by observations. The preferred scenario, which results in convection heating by bulk flow of the crust, corresponds to a felsic crust and to a heat flow at the CMB higher than 60 mWm^{-2} during the whole thermal evolution of the plateau. But it should be noted that Babeyko et al. (2002) assume almost no temperature-pressure dependence of conductivity, with $\lambda[Wm^{-1}°C^{-1}] = -0.38 \times \ln(T\,[°C]) + 4.06$, and fixing the conductivity at 2.5 $Wm^{-1}°C^{-1}$ for temperatures higher than 50 °C (S.V. Sobolev, personal communication); for intra-crustal heat sources, temperature at 20-25 km depth and surface heat flow are always decreasing for a time span of 30 My or more. Le Pichon et al. (1997) have obtained a thermal model to compute the evolution of the density of the crust with time that explains that part of the uplift of a plateau can occur independently of the tectonic shortening, presumable responsible for its thickening (for this controversial question, see also Husson & Sempere, 2003; Hindle et al., 2004).

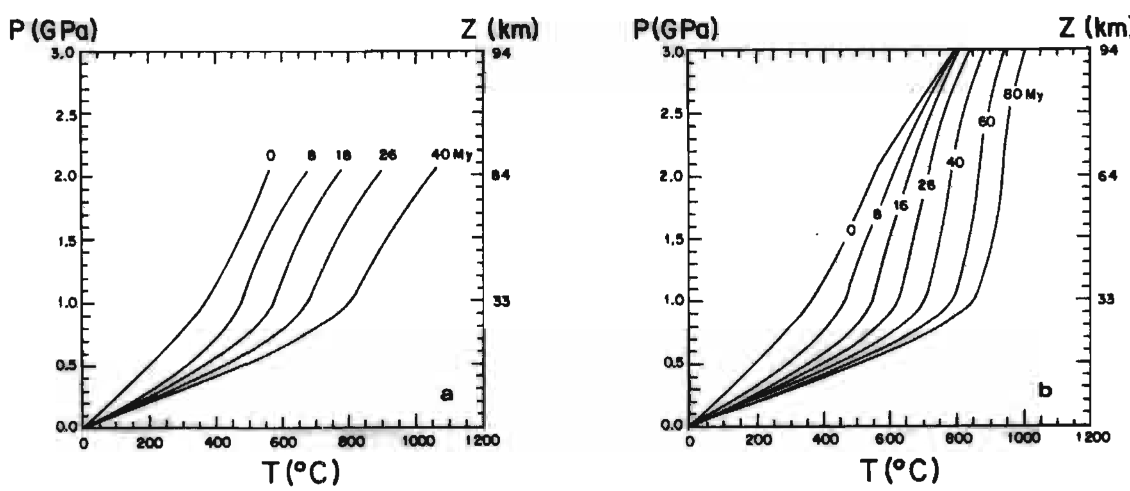

Figure 3. Evolution of the geotherm with time for a 70 km thick crust with an intermediate composition for the lower crust (from Le Pichon et al., 1997). (a) Homogeneous crustal thickening of a 35 km thick continental crust with surface heat flow of 68.3 mWm^{-2} (constant heat flow at the crust/mantle boundary of 24 mWm^{-2}). (b) Homogeneous crustal thickening, with attached lithospheric mantle (very low heat flow at the crust/mantle boundary).

The evolution of the geotherm obtained by Le Pichon et al. (1997) is shown in Fig. 3 corresponding to two model cases. The results presented formerly have affinity with those of Le Pichon et al. (1997) especially when it is considered that in Fig. 3 the last stages of the geotherm evolution indicate a surface heat flow of about 120 mWm^{-2} whereas here it has been assumed that the present heat flow is of about 70-80 mWm^{-2} corresponding nearly to an initial crust of moderate surface heat flow (50-60 mWm^{-2}); also, the thermal gradients of geotherms in Le Pichon et al. (1997) for pressures P larger than 1.0 GPa should be increased because they have assumed a very weak temperature dependence of conductivity for the lower crust.

CONCLUSIONS

The Altiplano evolved from a crust of moderate surface heat flow of about 50-60 mWm^{-2} generating during the thickening process a rather heterogeneous thermal structure. Radiogenic heating of the thickened crust is a main cause of electrical conductivity and seismic anomalies observed beneath some areas of the plateau. Further thermal studies of the sedimentary basins, new heat flow data and a comprehensive view of fluid circulation are needed to establish more properly its thermal evolution and the interaction with heat flow and thermal processes in the mantle lithosphere.

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