The early heat loss evolution of Mars and their implications for internal and environmental history

Javier Ruiz

Departamento de Geodinámica, Facultad de Ciencias Geológicas, Universidad Complutense de Madrid, 28040 Madrid, Spain.

The time around 3.7 Ga ago was an epoch when substantial changes in Mars occurred: a substantial decline in aqueous erosion/degradation of landscape features; a change from abundant phyllosilicate formation to abundant acidic and evaporitic mineralogy; a change from olivine-rich volcanism to olivine-pyroxene volcanism; and maybe the cessation of the martian dynamo. Here I show that Mars also experienced profound changes in its internal dynamics in the same approximate time, including a reduction of heat flow and a drastic increasing of lithosphere strength. The reduction of heat flow indicates a limited cooling (or even a heating-up) of the deep interior for post-3.7 Ga times. The drastic increasing of lithosphere strength indicates a cold lithosphere above the inefficiently cooled (or even heated) interior. All those changes experienced by Mars were most probably linked and suggest the existence of profound interrelations between interior dynamics and environmental evolution of this planet.

Paleo-heat flows based on determinations of lithospheric strength are useful for obtaining information on the thermal evolution of a planetary body. So, Ruiz et al. used paleo-heat flow estimates for 22 martian regions of different periods and geological contexts, derived from the effective elastic thickness of the lithosphere or from faulting depth beneath large thrust faults, to constrain the thermal history of Mars. The results obtained by these authors suggest that Mars has been losing less heat than predicted from most thermal history models. The martian mantle could even have been heating-up during a substantial portion of its evolutionary history.

An interior of Mars inefficiently cooled is, as pointed out by Ruiz et al., consistent with several independent geophysical and geological observations: the present-day existence of a (at least partly) fluid core, and the simultaneous absence of an endogenic magnetic field, are in agreement with high mantle temperatures reducing core cooling, and weakening or suppressing core convection; a lower amount of global contraction since the Early Noachian than expected from thermal history models, as deduced from thrust faults recorded on the surface, because lack of mantle cooling limits the thermal contraction that can drive surface contraction; the evidence for recent extensive volcanism also suggest the retention of a substantial amount of internal heat. On the other hand, a limited interior cooling would be in agreement with a recent thermal history modeling, which take into account phase transitions and compositional changes due to mantle melting.

I add here that recent petrologic modeling of magma production of magmas originating shergottite meteorites (which typically have ages younger than 0.6 Ga) than for magmas originating Gusev and Meridiani basaltic rocks, which are Noachian in age. (This period corresponds to ages older than 3.7 Ga, and was followed by the Hesperian; e.g., Ref. 15) Shergottite meteorites could have originated in a hot-plume environment, and therefore be not representative of the average mantle, but a long-term heating-up mantle would be a solution to the problem of the high-temperature shergottites.

The relatively low paleo-heat flows deduced for Mars from lithospheric strength are mostly related to post-Noachian times. Otherwise, most of paleo-heat flows obtained for Noachian times are high or lower limits, and in the second case the heat flows could also be high; this would be consistent with an early time of intense interior cooling. In this work I investigate the key question of timing the change from high heat flows to comparatively low heat flows, through both (1) the comparison between radioactive heat production and surface heat flow as a function of time, and (2) the temporal pattern of lithospheric strength evolution, which relates to the thermal evolution of the lithosphere. I then compare this timing with other major events in the martian history, and discuss the implications of the heat loss history of Mars for the evolution of mantle temperature, core dynamo, and climate of this planet.
Results

The evolution of the interior heat budget as constrained by paleo-heat flows. The Urey ratio \( (U_r) \), the ratio between the total radioactive heat production and the total heat loss through the surface of a planetary body, constitutes an excellent means to visualize the internal heat budget of a planetary body. Thus, information on both paleo-heat flows and global planetary composition can be used to constrain the evolution of the Urey ratio, and hence the thermal evolution of a planetary body.

I have calculated Urey ratios (or more exactly, paleo-Urey ratios) for Mars from paleo-heat flows derived from lithospheric strength and the total radioactive heat production (function of age) derived from the global compositional model of Wänke and Dreibus\(^\text{16}\). The Urey ratio values were calculated extrapolating heat flows obtained for a given region to the entire planet. It is obvious that those “local” paleo-heat flows are not necessarily representative of globally average values, but so-derived “equivalent” \( U_r \) values are certainly informative, as I show below. For \( U_r > 1 (U_r < 1) \) the interior, as an average, is heating-up (cooling-down). Thus, for a given feature and time I have selected the upper limit surface heat flow (Figure 1), which in turn gives a lower limit for \( U_r \), being therefore the less favorable cases for a limited cooling or a heating-up of the martian interior, and making more robust my conclusions.

The used paleo-heat flows (Figure 1) are mostly based on the upper limits obtained by Ruiz et al.\(^3\), although I have expanded the calculations for Olympus Mons, Arsia Mons, and Elysium rise in order to take into account an higher age range for the time of loading of these volcanic edifices. Previous work\(^3\) considered the time of loading of Olympus Mons, and Tharsis Montes to be generally Amazonian (i.e., younger than \( \sim 3 \) Ga), in accordance with the ages of surface-covering lavas. However, crater counts of volcano flanks suggest that the main edifice, or at least a substantial portion, of those volcanoes was already emplaced around \( 3.5-3.6 \) Ga (Ref. 17), and therefore, the time of loading could have been mostly Hesperian\(^18\). Similarly, flexure modeling and crater counts suggest that most of Olympus Mons would have constructed between 3.67 and 2.54 Ga (Ref. 19).

Figure 2 shows lower limit estimates of Urey ratio for Mars as a function of time. Curve lengths indicate uncertainty related to feature age, not temporal evolution. For clarity, I have simplified the representation to show only the lower Urey ratio value obtained for a given region. Heat flows for post-Noachian volcanic regions would be consistent with \( U_r < 1 \) if these regions were loaded in the Amazonian, although volcanic areas are expected to be associated with heat flows higher than planetary averages. However, as stated above, refined crater counts suggest relatively early ages for the emplacement of Elysium and Tharsis volcanoes\(^17\), and therefore a Hesperian age for the associated loading\(^17\), and hence \( U_r > 1 \) values are favored from these features for the Hesperian. Moreover, all post-Noachian non-volcanic regions give \( U_r \) values higher than, or close to, 1; similarly, the present-day upper limit heat flow derived for the North Pole region is slightly above the present-day heat production, consistent with an upper limit \( U_r \) close to 1 (Ref. 3; see also Figure 1). The conclusion of a limited interior cooling seems therefore inescapable for post-Noachian times; moreover, the case for an interior heating-up is robust, at least during the Hesperian.

For the Noachian the situation is unclear, because many heat flow estimates for the Noachian southern highlands are lower limits\(^3\) (and so they are not shown in Figure 1), based on upper limit effective elastic thicknesses. Therefore a very high surface heat flow, and hence a very low Urey ratio, cannot be ruled out for the time when the large-scale topography of these regions was formed (for illustrating this, an Urey ratio close to zero for the southern highlands has been represented in Figure 2). Thus, most of the heat flows derived for Noachian terrains could be consistent with \( U_r < 1 \) and a substantial interior cooling. Relatively low heat flows deduced from the depth of large thrust faults at Amethyst Rupes and Warrego rise could suggest a highly variable heat loss across the martian surface in the Noachian, although the age of these thrust faults would be, respectively, between \( 3.6 \) and \( 3.8 \) Ga and between \( 3.7 \) and \( 4.0 \) Ga (Refs. 20, 21), whereas the

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Figure 1 | Surface heat flow upper limits for several regions and times of Mars, indicating the geological category of the respective features. Curve lengths indicate uncertainty related to loading/deformation age (and hence to radioactive heating in the lithosphere), not temporal evolution. Most of heat flows shown in this figure are from Ruiz et al.\(^3\), although the calculations have been expanded in order to account for an higher age range for the time of loading of the lithosphere supporting Olympus Mons, Arsia Mons, and Elysium rise (see text for details).
The evolution of lithospheric strength and thickness. The effective elastic thickness, \( T_e \), is related to the strength of the lithosphere, integrating contributions from brittle and ductile layers and from elastic cores of the lithosphere (for a review see Ref. 22); when considering curvature due to lithospheric flexure, \( T_e \) can in fact be used as a measure of the total strength of the lithosphere. Mantle rocks are stronger than crustal rocks, but under higher temperatures, which in turn reduces their strength. For a sufficiently high heat flow the upper mantle does not contribute to \( T_e \), and the lithosphere is thin and restricted to the crust. Conversely, if the heat flow is sufficiently low, the upper mantle is cold and strong, and contributes to the strength of the lithosphere, and hence to the effective elastic thickness, which is therefore higher. Thus, the evolution of the effective elastic thickness is very interesting as an indicator of the evolution of the thermal state, strength and mechanical behaviour of the lithosphere.

Table 1 shows a summary of effective elastic thickness values derived for martian terrains loaded in different times, compiled from several works (see Methods Section). Much of these \( T_e \) estimates are not adequate for performing well-constrained heat flow calculations (and for that reason heat flow values for the respective regions are not shown in Figure 1), because curvatures of the equivalent elastic layer and curvature are abundant evidence for a substantial decline in aqueous erosion/degradation of landscape features, indicating a reduction of the amount or lifespan of liquid water on the surface; a change from abundant phyllosilicate formation to abundant acidic and evaporitic mineralogy occurred some time previously to the end of the Noachian, indicating a transition from wetter to drier environmental conditions; a change from olivine-rich to olivine-pyroxene volcanism, indicating a modification in magma composition that could be due to a reduction of mantle temperature or to melts formed under higher pressure, and therefore from a deeper magma source below a thicker lithosphere. Similarly, recent careful examinations or Hesperian times, as a function of the age derived from model-ages based in crater counts; for representation clarity, Figure 3 shows only the central value of the range of \( T_e \) for each feature.

\( T_e \) estimates proposed for Noachian times are lower than 25 km, which are low relative to values obtained for post-Noachian terrains, suggesting a comparatively high heat flow in that time. Regions loaded (or that could have been loaded) in Late Noachian or Early Hesperian times (corresponding to an age range between 3.8 and 3.6 Ga; Ref. 15), show a wide dispersion of \( T_e \) values, generally higher than for Noachian, and very high values appear. Otherwise, most of \( T_e \) values for Hesperian or Amazonian times are higher than ~40–50 km, implying low heat flow and a significant contribution of the upper mantle to the strength of the lithosphere. Indeed, as indicated above, if the uppermost mantle is sufficiently cold then it contributes to the strength and effective elastic thickness of the lithosphere, and this results in an increased \( T_e \) (e.g., Refs. 22,23). Moreover, if such a strong mantle lithosphere is welded to the crust (which occurs if the strength at the crust base is sufficiently high) then there is a drastic increasing of \( T_e \) with respect to the case with mechanically decoupled crust and mantle lithosphere.21–24

Thus, the transition from low to high values of \( T_e \) occurred around the Noachian/Hesperian boundary, with a substantial increasing of \( T_e \) after 3.6–3.7 Ga. This transition is related to a substantial contribution of the mantle to the strength of the lithosphere, and maybe to crust-mantle lithosphere welding. Thermal evolution models20,24 have shown that a local substantial increasing in \( T_e \) does not necessarily require a change in heat flow decline; but the timing of \( T_e \) increasing can largely vary from a region to another as a consequence of differences in the local lithospheric thermal regime (including variations in other local factors, such as curvature and crustal thickness, should make more important this effect), but Figure 3 shows a general increase in \( T_e \) is a relatively narrow temporal window. The timing for this dramatic lithospheric strength increasing is similar to that for the inferred starting of \( Ur > 1 \) values. (Under appropriate conditions a reduction of heat flow may simultaneously produce interior heating of the convective mantle and cooling and thickening of the lithosphere.) In any case, this lithospheric strength increasing was probably not simpler, since the dispersion of \( T_e \) values for Late Noachian or Early Hesperian times; this is not surprising, as lithosphere strength depends, among other factors, on crustal thickness and curvature.22–24

On the other hand, the frequently stated increasing of \( T_e \) with age through the martian history (e.g., Ref. 25) is not evident after Early Hesperian, but uncertainty in loading time for most of \( T_e \) estimates for post-Noachian terrains preclude us of obtaining a definitive conclusion with respect to this point. The effective elastic lithosphere loaded by the north polar cap, a nearly current structure, has been inferred to be thicker than 300 km (Ref. 26), but the case for the lithosphere under the south polar cap, whose present-day effective elastic thickness could be as low as 110 km, is more poorly constrained.27

Discussion

The time around the Noachian/Hesperian boundary was an epoch when substantial changes in Mars occurred (Figure 4): there is abundant evidence for a substantial decline in aqueous erosion/degradation of landscape features, indicating a reduction of the amount or lifespan of liquid water on the surface; a change from abundant phyllosilicate formation to abundant acidic and evaporitic mineralogy occurred some time previously to the end of the Noachian, indicating a transition from wetter to drier environmental conditions; a change from olivine-rich to olivine-pyroxene volcanism, indicating a modification in magma composition that could be due to a reduction of mantle temperature or to melts formed under higher pressure, and therefore from a deeper magma source below a thicker lithosphere. Similarly, recent careful examinations...
Table 1 | Effective elastic thickness of the lithosphere derived for several martian terrains

| Featurea | Surface age (Periods)b | Surface age (Ga)c | $T_e$ (km)d |
|----------|------------------------|------------------|-----------|
| North Pole | Current | Current | >300 |
| South Pole | Current | Current | >110 |
| Valles Marineris (1) | H-A | 3.6–1.8 | 60 |
| Alba Patera (2) | H-A | <3.5 | 43–65 |
| Arsia Mons (3) | H (or A) | −3.6 (or lower) | 20–35 |
| Pavonis Mons (4) | H (or A) | −3.6 (or lower) | 50–100 |
| Ascraeus Mons (5) | H (or A) | −3.6 (or lower) | 50–80 |
| Olympus Mons (6) | H-A | 3.7 to 2.5 | >70 |
| Elysium rise (7) | H | 3.7–3.0 | 36–45 |
| Syrtis Major (8) | H | 3.7–3.0 | 10–15 |
| Solis Planum (9) | H | 3.7–3.0 | 24–37 |
| Isidis Planitia | H | 3.7–3.0 | 100–180 |
| Amenthes, dichotomy (10) | N-H | 3.8–3.6 | 15–35 |
| Aeolis region (11) | N-H | 3.8–3.6 | 10–15 |
| Hellas S rim (12) | N-H | 3.8–3.6 | 20–120 |
| Hellas W rim (13) | N-H | 3.8–3.6 | <20 |
| Amazonis Planitia (14) | N-H | 3.7–3.6 | <45 |
| Acidalia Planitia (15) | N-H | >3.6 | 13–30 |
| Arcadia Planitia (16) | N-H | >3.6 | <23 |
| Thaumasia Highlands (17) | N-H | 3.8–3.6 | 19–23 |
| Tyrrhena Patera (18) | N-H | 3.9–3.3 | <28 |
| Corax Fossae (19) | N-H | 3.9–3.5 | 10–13 |
| Acheron Fossae (20) | N-H | 3.9–3.6 | 9–11 |
| Amenthes, highlands (21) | N | >3.8 | <25 |
| NE Arabia Terra (22) | N | >3.8 | <16 |
| Noachis Terra (23) | N | >3.8 | <12 |
| Terra Cimmeria (24) | N | >3.8 | <12 |
| Hellas Basin (25) | N | 4.1–3.9 | <13 |

The numbered features refer to those represented in Figure 3.

Age relations based on Refs. 2, 17, 20, 49, 51, 56–60.

For determination of $T_e$ ranges see Methods Sections.

Figure 3 | Effective elastic thickness of the lithosphere ($T_e$) for martian terrains loaded (or that could have been loaded) in Noachian or Hesperian times, as a function of age. Numbers referring to given geological features are as in Table 1. For clarity, for a given feature only the central $T_e$ value of the range in Table 1 is shown. Line lengths indicate uncertainty related to loading age.
I have shown here that Mars also experienced profound changes in its internal dynamics in the same approximate time, including a reduction of heat flow and a drastic increasing of lithosphere strength and thickness. The reduction of heat flow, when compared with global radioactive heat production, indicates a limited cooling (or even a heating-up, consistent with the high temperature of formation of shergottite meteorites) of the deep interior for post-Noachian times. The drastic increasing of lithosphere strength indicates a cold lithosphere (including a strong uppermost mantle and maybe crustal-mantle lithosphere welding) above the inefficiently cooled (or even heated) interior.

The, at least rough, time coincidence of the above mentioned changes (Figure 4) seems unlikely to be accidental. Those changes were most probably linked, suggesting profound interrelations between interior dynamics and environmental evolution of this planet. The reduction of the heat flow from the deep interior produced a cooling and thickening of the lithosphere, evidenced by the drastic increasing in effective elastic thickness deduced for the same approximate time. This in turn would be consistent with a change in magma mineralogy due to a magma source located below a thicker lithosphere.

The rise of mantle temperature caused by an interior heating-up would have reduced the thermal gradient between mantle and core, weakening or even stopping core convection and dynamo\(^29\). Even if one accepts the earlier chronology for dynamo cessation, it could have been the first sign of the big changes experienced by Mars around the Noachian/Hesperian boundary. The cessation of the dynamo would eliminate the magnetic shielding of the atmosphere, contributing to their severe erosion, and therefore heavily affecting the climatic and hydrologic evolution of Mars\(^29,30\). As a consequence, the erosion/degradation for liquid water would be greatly diminished, and a wetter environment permitting clay formation would transition to more arid conditions favouring the formation of acidic and evaporitic mineralogy.

Low post-Noachian heat flow values could be indicative of an important reduction in mantle convection efficiency, which could in turn be related to inefficient water recycling in a stagnant-lid planet (e.g., Ref. 40). Thus, it is possible that internal water cycles and environmental evolutions were profoundly interrelated throughout the history of Mars. A more profound understanding of these interrelations is necessary for our knowledge of the biological potential of this planet, past, present, or even future.

Methods
Calculation of heat flows from effective elastic thickness. The effective elastic thickness of the lithosphere can be converted to heat flow following the equivalent stress envelope formalism. Here I use an implementation of this methodology which takes into account the stress contributions from the crust and the lithospheric mantle\(^41\), in order to calculate upper limit heat flows for the Hesperian for Olympus and Arisia Montes, and for the Early Hesperian for Elysium rise. For calculating heat flows I use estimates of both the effective elastic thickness of the lithosphere and the curvature of the equivalent elastic layer\(^3\). I use values widely used for Mars\(^3,33\) for the Young’s modulus (100 GPa), Poisson’ ratio (0.25), crustal density (2900 kg m\(^{-3}\)) and mantle lithosphere density (3500 kg m\(^{-3}\)). The admittance modeling\(^1\), from which derived the used effective elastic thicknesses, was performed for a mean crustal thickness of 50 km, whereas the updated crustal thickness map of Neumann\(^32\) was constructed for a mean crustal thickness of 45 km. Thus, in my calculations I assume local crustal thicknesses based on the regional trends of those authors\(^32\), but increased in 5 km, in order to be consistent with the values of the effective elastic thicknesses: 55, 85 and 45 km for, respectively, Olympus Mons, Arisia Mons, and Elysium rise. For calculating brittle strength I assume zero pore pressure, which serves to put an upper limit to the obtained surface heat flow.

For creep parameters of the Martian crust I use the flow law of wet diabase\(^43\), which is consistent with extensive evidence of water-related geological activity in early Mars. The ductile strength of the mantle lithosphere is calculated for dry olivine dislocation creep rheology (a flow law obtained for artificially dried dunites\(^44\)), which give an upper limit to the surface heat flow, which is useful to obtain a lower limit for the Urey ratio. A strain rate of \(10^{-14} \text{s}^{-1}\) was used in the calculations, based on the faster strain rates in the range usually considered for Mars\(^3\), which is again useful to obtain a heat flow upper limit; to make a comparison, terrestrial strain rates are typically \(\sim 10^{-16} \text{s}^{-1}\) in Earth active continental interiors; using slowly strain rates would result in lower heat flows and higher Urey ratios. The base of the mechanical lithosphere is here defined as the depth at which the ductile strength reaches a low value of 0.09 MPa, and below which there are no further significant increases in strength, although the exact value selected does not produce significant changes in the calculations due to the exponential dependence of ductile strength on temperature. The strength level at the crust base determining if mantle lithosphere and crust are mechanically welded is taken as 10 MPa; again, the exact value selected does not produce significant changes in the obtained results.

For a complete description of this procedure see Ruiz et al.\(^1\).

Temperature profiles. I calculate temperature profiles following the procedure described by Ruiz et al.\(^2\), and using crustal heat-producing elements (HPE) abundances based on Mars Odyssey GRSE measurements. The significant homogeneity of elemental abundances measure by Mars Odyssey GRSE suggests that the Martian crust is much less geochemically varied than the Earth’s crust. Although it is possible that crustal HPE abundances decrease with depth, a homogeneous distribution gives higher surface heat flow and is therefore useful for my upper limit calculations. For the mantle lithosphere I use HPE abundances 0.1 times, per unit mass, the crustal average, consistent with a ratio between crustal and primitive mantle HPE abundances higher than −10 for Mars. Heat dissipation rates were calculated for standard decay constants.

For the thermal conductivity I use a constant value of 2 W m\(^{-1}\) K\(^{-1}\) for the crust, appropriate for intact (non-porous) basaltic rocks, and a temperature-dependent thermal conductivity appropriate for olivine for the mantle lithosphere. The surface temperature is assumed as 220 K, the present-day mean surface temperature on Mars, which is consistent with the inference of low near-surface temperatures deduced for most of the past 4 Ga from ALH84001 thermochronology\(^43\).

For a complete explanation and discussion of the values used in the calculations see Ruiz et al.\(^1\).

Urey ratio calculations. I derive paleo-Urey ratios by calculating the ratio between the two radioactive heat production, for a given time, obtained from the compositional model of Mars of Wänke and Dresibus\(^2\) and standard decay constants, and the surface heat flow derived from lithospheric strength. Thus, the paleo-Urey ratio is given by \(R(t) = \frac{\text{uranium}}{\text{uranium + thorium}}\), where \(U(t)\) is the mean heat production rate by mass unit, \(U_{\text{uranium}}\) and \(U_{\text{uranium}}\) are, respectively, the mass and the mean radius of Mars, and \(R(t)\) is the surface heat flows derived for a given feature and age. The so-derived \(U(t)\) values are in fact “equivalent” \(U(t)\) values, since they are based on “local” heat-flow heat flows.

Ranges of effective elastic thickness of the lithosphere. Values of effective elastic thickness have been compiled from literature\(^2,18,20,26,46–55\). In the case of features for

| Age (Ga) | Effective Elastic Thickness (km) |
|---------|---------------------------------|
| 3.5     | 20                              |
| 3.6     | 30                              |
| 3.7     | 40                              |
| 3.8     | 50                              |

Figure 4 Time range for several important changes that occurred in Mars around or close to the Noachian/Hesperian boundary.
which two or more different T, ranges have been proposed (as is the case for the Tharsis volcanic province). I have consigned the range of elastic thicknesses consistent with all the estimates, which are therefore considered equally valid.

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