THE LITTLE ICE AGE ON MARS : POTENTIAL DETECTION WITH THE INSIGHT HEAT FLOW AND PHYSICAL PROPERTIES PACKAGE (HP^3)

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1. Introduction

The Little Ice Age (LIA) was the period 1200-1900 during which cooling was observed in many regions on Earth, and has been traditionally attributed to drop in solar radiation associated historical records of low sunspot numbers (e.g. the Maunder minimum). More recently, other causes such as volcanic aerosols or changes in thermohaline circulation have been suggested. A discriminator between these hypotheses is that a solar forcing would also cause a ~0.5K cooling on other planetary bodies, most notably the Moon and Mars. This cooling, as on Earth, will influence the subsurface temperature profile which carries a long-term 'memory' of surface temperatures. The forthcoming NASA Insight mission in 2018 intends to measure the temperature profile in the Martian regolith to a depth of 3-5m. While the LIA measurably influences the temperature profile in this depth range, analysis suggests the effect may be too small to detect unambiguously. However, the LIA influence may be the limiting factor on the precision of the retrieval of heat flow measurements from those same data**.**

1. Background

The Little Ice Age (LIA) on Earth saw northern hemisphere surface temperature reductions (e.g. Mann et al., 2009) that were a significant influence on human history (e.g. Fagan, 2001; Lamb, 1972). Traditionally, the LIA was attributed to a reduction in solar output, since the period coincides with sustained drops in the number of observed sunspots (e.g. Eddy, 1976) namely the Maunder and Spörer Minima. However, recent work has suggested alternative explanations, most recently (e.g. Miller et al., 2012) that reflection of sunlight by volcanic aerosols triggered a cooling, which was sustained by climate feedbacks such as changes in the thermohaline circulation (Broecker, 2000). Since the LIA represents a natural experiment on the climate system , it serves as a tool with which to test climate prediction models (e.g. Tung et al., 2008), and it is important to understand the extent to which, if at all, external solar forcing contributed.

The climate data with which the terrestrial LIA is characterized includes meteorological records, historical data such as reports of skating on the River Thames in London, and proxies such as pollen records and isotopic ratios in ice cores. Additionally, borehole temperature profiles, which retain a memory of surface temperatures (e.g. Huang et al., 2010) show a signature of the LIA. Borehole temperature profiles in ice caps, for example, the GRIP borehole temperatures in Greenland (Dahl-Jensen et al., 1998) show a clear dip at a depth of ~60m which is due to the LIA. Although evidence of a cool period is most prolific in the northern hemisphere (motivating non-astronomical causes to be advocated) there is also evidence of a ~0.5K cooling in Antarctic borehole temperatures (Orsi et al., 2012).

Examining the terrestrial climate history, the overall picture is of an irregular dip in temperatures by ~0.5-1K over a period of ~700 years (i.e. a periodic variation with period 1400 years). However, components of the forcing have rather shorter durations - e.g. the Maunder minimum was ~70 years long around 1700, and the Spörer minimum lasted >100 years centered around 1500. We consider here whether such a forcing on Mars might be detectable

2. The Little Ice Age on Other Worlds

In principle, subsurface temperatures other planetary bodies might show a similar signature at comparable depths. To evaluate these, we must first consider the change in solar flux S required to produce the change in terrestrial temperatures. A drop in absorbed insolation ΔF (where $F=(1-A)S/4$, with A the planetary albedo and S the solar constant) will yield a surface temperature perturbation ΔT that depends on the climate sensitivity $\lambda = \Delta T/\Delta F$. For a black body $F=\sigma T^4$, with the Stefan-Boltzmann constant and T the absolute temperature, giving $dF/dT=4\sigma T^3=4F/T$ and thus $\lambda = T/4F$. For Earth with F~250 Wm^{-2} (S=1360 Wm^{-2}, A~0.3) and T~288K, λ ~0.28 K/(Wm^{-2}). Thus a 0.5K drop would mean a drop in insolation F of ~2 Wm^{-2}, or about 0.7%. However, the Earth system features a number of feedbacks, both positive and negative on a range of timescales that influence λ. Two obvious positive feedbacks are the ice-albedo feedback, wherein a lower surface temperature results in more extensive ice sheets, and thus higher albedo and lower absorbed sunlight and the condensable greenhouse feedback, where lower surface temperatures result in less water vapor in the atmosphere and thus lower greenhouse warming : in both cases the temperature drop is amplified over that by the direct forcing. Clouds likely are a major feedback whose magnitude and even sign is not known with certainty. Studies of the earth temperature variation (Tung et al., 2008) over the 11-year solar cycle suggest λ ~0.69-0.97 K/Wm^{-2}, in which case if the 0.5K LIA temperature drop were entirely the result of solar effects, it would need a 0.5-0.7 Wm^{-2} forcing in F, or a 1.4-2 Wm^{-2} forcing in S (i.e. 0.1-0.15%).

For Mars, with S~ 587 Wm^{-2} and A~0.25 and annual average temperature ~202K, the black body sensitivity λ~0.5 K/Wm^{-2} is somewhat smaller than Earth (due to the dependence of the sensitivity on the cube of temperature). In principle, an ice-albedo feedback may operate on Mars (due to the CO_2 frost cycle which forms seasonal frost caps visible from Earth), although simulations suggest this is small (e.g. a ~100 Wm^{-2} forcing in S is needed to expand the ice cap by only a few degrees of latitude - see Nakajima and Tajika, 2001). Similarly, the greenhouse effect on Mars is rather small (~5K, e.g. Pollack, 1979) , so there is limited scope for amplification by feedback . Thus we assume the black body climate sensitivity on Mars holds, and given the solar forcing above (i.e. 0.1-0.15% drop in F, or 0.1 Wm^{-2}) , the Martian surface temperatures a few centuries ago was most likely about 0.5K lower than present, and certainly no more than 1K lower.

3. LIA Signature in the Temperature Profile

The planetary subsurface temperature profile is a classic problem in geophysics (e.g. Kelvin, 1861; Turcotte and Schubert, 1982 ; Carslaw and Jaeger, 1959). The depth to which a previous surface temperature perturbation has influence relates to the skin depth, which is a function of the thermal properties of the crustal rock or ice (see later). For solid rock or ice with thermal conductivity of ~2$Wm^{-1}K^{-1}$, the skin depth for millennial forcing is of the order of 100m. Larsen and Dahl-Jensen (2000) consider temperature profiles in the Martian ice cap forced by longer period (125kyr) Croll-Milankovich cycles, but do not examine the LIA influence.

In porous regolith the thermal conductivity is much lower (0.02-0.1 Wm-1K-1) than solid ice and the associated skin depths are much shallower, by a factor of ~10. It is this shallow depth at which the annual thermal wave is damped that makes planetary heat flow measurements feasible with measurements only a few meters below the surface, in contrast with the ~100m length scales more typical for terrestrial boreholes.

Shallow boreholes were drilled in the lunar regolith by astronauts on Apollo 15 and 17 and strings of temperature sensors inserted to determine the geothermal heat flow of the moon (e.g. Langseth et al., 1972; see also Kömle et al., 2011 and Keiffer, 2012). These strings reached 1.5m depth, too shallow to be sensitive to the LIA. However, NASA recently selected the InSight mission (to perform geophysical measurements at Mars, including an experiment intended to measure the heat flow. This is to be achieved with a self-hammering drill (informally, a 'mole' - formally the 'Heat Flow and Physical Properties Package' or 'HP3') which will drag a string of temperature sensors to a depth of 3-5m (Spohn et al., 2001; Spohn et al., 2012). Such depths are required to minimize the effect of the annual variation in surface temperature which has a signal that is superposed on the linear increase of temperature with depth due to the heat flow. However, these depths are such that a LIA signal may be present. Since skin depth varies as time^0.5, the signature of a ~200 Mars year event propagates ~14 times deeper than the annual wave.

Previous analysis (Grott et al., 2007) showed that errors in retrieved heat flux due to surface climate change of a few years (e.g. due to the interannual variability in Martian dust storms), or due to climate change forced by Croll-Milankovich orbital cycles on Myr timescales, would be small. However, that same analysis (Grott et al., 2007, figure 7) shows that errors for periods of ~10-1000 Martian years could be significant.

We quantify the effect by using the classic heat conduction solution (e.g. Turcotte and Schubert, 1982). The contributions to temperature T(z) at depth z in a uniform half-space are the long-term average surface temperature To, plus the increment due to the geothermal temperature gradient plus increments from various periodic surface temperature forcings, written as

$$T(z)=T_o+(Fz/k)+\Delta T_a\exp(-z/\delta_a)\cos(\theta_a-z/\delta_a)+\Delta T_l\exp(-z/\delta_l)\cos(\theta_l-z/\delta_l)$$

Each periodic forcing has an angular rate $\omega=2\pi/P$ and an associated skin depth $\delta=(2\kappa/\omega)^{0.5}$ where $\kappa=k/\rho c_p$. As a function of time, the surface temperature forcing is written $\Delta T\cos(\omega t+\theta)$ where θ is a phase term to denote time of day, season or phase in any periodic insolation variation. For the purpose of the present discussion, we omit the diurnal heat wave which is attenuated at very shallow depths and consider only the annual wave ΔTa=20K, Pa=687 days=5.9E7s. Over the course of a Martian year, θa will vary over the range 0-2π. We model the effect of a LIA insolation drop as ΔTl=1K and Pl=200-1400yrs = 6.2E9-4.3E10s and, since the LIA perturbation is taken as zero at present and rising, θl=3π/2.

For ρ=1500 kg/m3, cp=600 J/kg/K (following Grott et al., 2007) and k=0.02-0.1 $Wm^{-1}K^{-1}$, we have κ=2.2E-8 to 1.3E-7 m^2s^{-1}, and δa=0.72-1.6m and δl=19-43m for 1400yr forcing, but only δl=7-16m for 200yr forcing. Thus, since heat flow measurements will be derived predominantly from temperature measurements in the 3-5m range (to get a couple of annual skin depths down), they will be sensitive to LIA contributions.

4. Results

Inspection of the results (e.g. figure 1) shows that if the thermal conductivity k of the regolith is perfectly known (assumed to be 0.02-0.1 $Wm^{-1}K^{-1}$), and the heat flow F is calculated from the difference in temperatures between 5m and 3m depth (T5,T3) as F=k(T5-T3)/2, then the percentage error without accounting for the unknown LIA contribution is ~3ΔTlk/F, i.e. for ΔT=1K, k=0.1 $Wm^{-1}K^{-1}$ and F=0.02 Wm^{2}, the error is 10% and is likely the dominant term in the error budget ; in the more likely scenario of ΔT=0.5K, the error is 5% The error is only weakly dependent on the duration of the LIA forcing, ranging in this case from 7% at 2000 years to 12% at 200 years.

For weakly conducting regolith, k=0.02 $Wm^{-1}K^{-1}$, the LIA signal is less attenuated than for the higher k (since the measurements at 3-5m are a smaller fraction of an LIA skin depth down) but the geothermal temperature gradient is some 5 times higher and thus the percentage error due to an unknown LIA contribution is small.

Ideally, measurements (as on Earth) would be made over a much wider depth range to get below the LIA signal altogether. For a shallow profile necessitated by practical limitations on spacecraft, with enough depth resolution and measurement precision, it might be possible to separate out the curved LIA signal from the linear geothermal one : likely (as is done on Earth) Monte Carlo and other probabilistic methods have to be used to perform the inversion, and even then it may well be impossible to separate the effects. It should be noted that while temperature measurements that are progressively more widely spaced towards greater depth are common in thermal profiles (since the diurnal and annual variations get damped out at shallow depths), there may be value in retaining a more uniform spacing for InSight to detect the curvature of a LIA signature on Mars.

5. Conclusions

If the LIA had a primarily solar forcing, a similar ~0.5K temperature drop may have affected the Moon and Mars, and is in principle detectable in borehole temperature profiles as on Earth. In the case of shallow profiles in low-conductivity regolith, the effect is likely difficult to isolate, but is a non-negligible (and possibly dominant) contributor to the error budget in planetary heat flow measurements.

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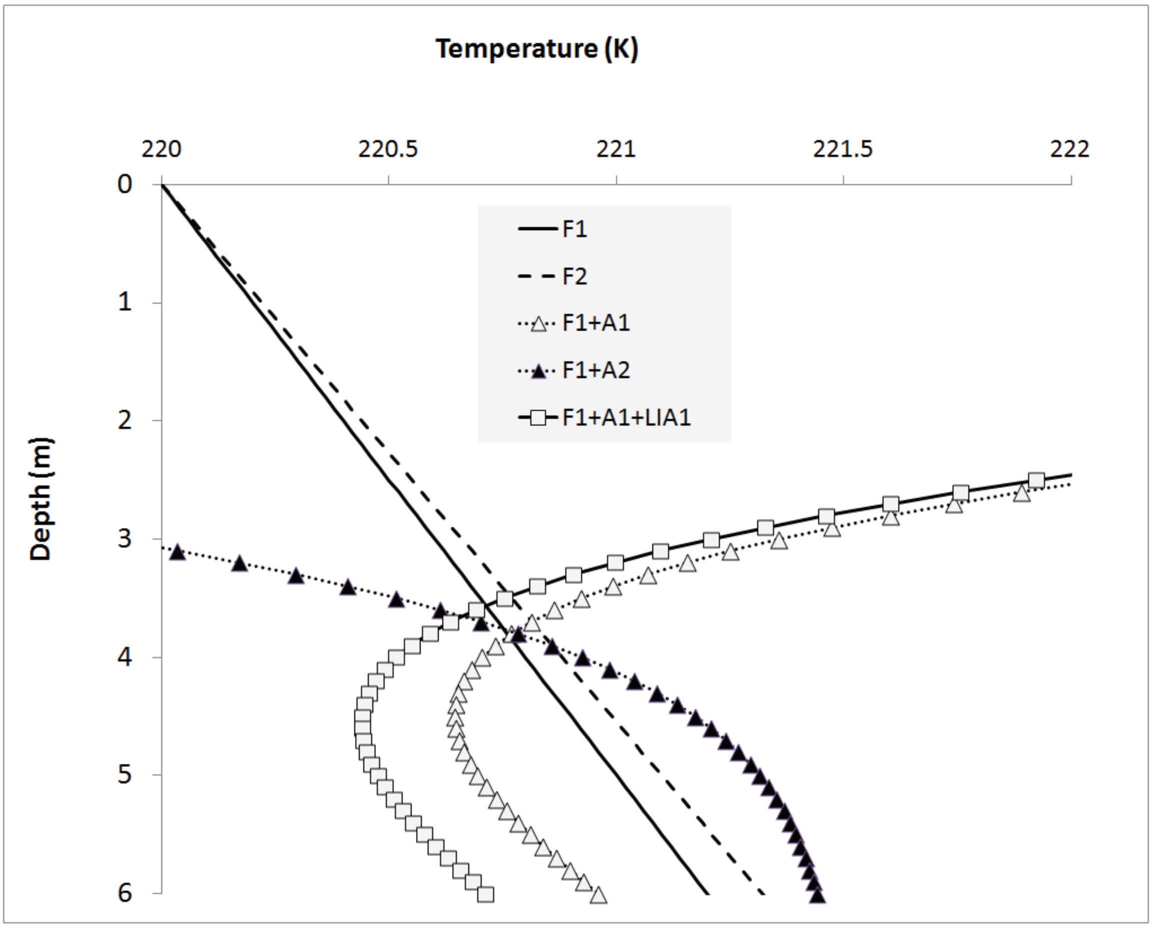

Figure 1. Example subsurface temperature profiles to show the different magnitude of the different contributions at depth. All are for a uniform thermal conductivity of 0.1 W/m/K. The solid line F1 is the reference profile for heat flow only with F=0.02 Wm^{-2}. Dashed line F2 shows the profile sensitivity to the desired parameter - this is the ideal profile for F=0.022 Wm^{-2}, thus a 10% difference in heat flow and represents roughly the desired performance of the measurement. Dotted lines with triangles show the perturbation to F1 for an annual surface temperature perturbation A1 of 10K, one half-year apart (and thus the filled and open triangles represent opposite phases of the annual forcing). In principle this contribution can be estimated from a 0.5-1 year long series of measurements. The line with open squares is the profile with a 1K LIA signal (period 200 years) superposed on the annual profile. The dominant effect over 4-6m is an offset, but with some growth in amplitude towards greater depth - the LIA contribution therefore masquerades as a decrease in inferred heat flow.