Lecture 10

Paleoclimates and Mars

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Much of the motivation for climate theory stems from the need to understand the current-day climate and the possibility that we are irrevocably changing it by burning fossil fuels, pumping pollutants into the atmosphere and cutting down rainforests. Unfortunately, our pursuit of the problem is hampered by the fact that the record of our quantitative observations of key quantities like sea surface temperature, CO_2 concentration and humidity is far shorter than most of the timescales on which the climate seems to vary intrinsically. Instead we need to rely on, for example, geology and geochemistry to construct proxy signals for the important climate variables that can be traced much further back in time. The reconstruction can then be used to further test and improve our understanding and models of the climate. Somewhat similarly, we need not only focus on our own terrestrial climatological experiment: the atmospheres of Mars and Venus, whilst very different in detail from our own, might also operate with analogous dynamical controls. Thus, one is tempted to assess our understanding by exploring past and extra-terrestrial climates. Unfortunately, this also highlights many other thorny issues which expose our basic lack of understanding and the shortcomings of our models. This lecture mentions some of the open and partiall answered questions that are raised by consideration of climate history and other planets.

1 The Eocene

About two million years ago, glacial cycles were initiated. Before that relatively late epoch of Earth's history, the climate appears to have been in a state of gradual cooling, lasting for some 65 million years. One possible explanation for this trend is based on the gradual reduction of atmospheric CO_2, and therefore the greenhouse effect (the precise cause of this reduction is not clear – the weathering of the uplifting Himalayas, which converts atmospheric CO_2 to carbonate minerals, could be responsible). Whatever the precise cause, this sets the stage for the "Eocene", a period about 55 million years ago when the climate was apparently relatively warm.

The evidence for a warm Eocene climate rests partly on paleoclimate data for oxygen isotopes in marine sediments which suggests that the deep ocean temperature was approximately 10° C (substantially warmer than present-day temperatures). This also suggests that there was no permanent polar ice cover because the melting of any polar ice would immediately flood the deep ocean with water of much lower temperatures. Further evidence is provided by fossil records from the Eocene period which reveal the presence of animal species in geographical regions that would be inhospitable for them today (for example, crocodiles inhabited the Hudson Bay and lemurs lived in Spitzbergen in Scandinavia). The overall conclusion is that the midlatitudes and poles were warmer during the Eocene, a conclusion that is generally accepted by climatologists.

By contrast, foraminifera (plankton) data suggests that the tropical temperature during this time was about 305 K or less, which is more comparable with temperatures experienced today. This highlights a curious puzzle: how can the climate maintain a relatively cool tropical region whilst raising the polar temperatures sufficiently to melt the ice caps? No compelling explanation currently exists. Certainly it is possible to explain elevated temperatures if the CO_2 concentration (and greenhouse effect) was higher. However, one then must increase the latitudinal heat transport significantly over present-day values in order to lower the surface temperature gradient to that required to keep the tropics cool. Unfortunately, detailed, state-of-the-art, coupled atmosphere-ocean models are unable to explain such enhanced heat transports (if anything, in the warmer Eocene temperatures, these models predict lower latitudinal heat transport). One possibility is that the ocean heat transport was enhanced in the Eocene, perhaps as a result of pronounced tidal dissipation or some other physical effect not incorporated into the coupled models. Another is that stratospheric clouds shrouded the tropics and reduced the incoming radiation sufficiently to render the tropics more temperate. Either way, we need some important revision of the climate models in order to solve the puzzle. Alternatively, it is conceivable that the estimated equatorial temperature is simply in error, in which case the interesting puzzle vanishes altogether.

2 The Neoprotozoic Snowball

Even further back in Earth history, about 600 million years, we arrive at another climate conundrum, the possibility that, on two or three occasions, the planet was completely frozen over – the Neoprotozoic "snowball Earth".

There are, in fact, some compelling reasons to believe the Earth was a snowball in the past. First, the C^{12} to C^{13} ratio in ocean sediments implies the ocean was relatively abiotic (devoid of organisms) in the past (the two isotopes of carbon are used differently by marine organisms), and one of the best explanations for an abiotic ocean is that it was frozen over. Second, there is geological evidence that the atmosphere contained high levels of CO_2. Under normal conditions, CO_2 is precipitated out of the atmosphere and removed by the weathering of rock. High CO_2 levels, however, can be built up and maintained by volcanic activity in a snowball Earth in which weatherable rocks are covered by ice. Finally, there is even geomorphological evidence for glaciation at low latitudes.

There are three main questions regarding the snowball Earth during this period. First, what are the conditions necessary for the Earth to become a snowball? Second, how would the Earth get out of a snowball climate as models suggest that, if it exists, the snowball is a stable state? Third, if the snowball Earth is commensurate with large quantities of atmospheric CO_2, what happened to all this CO_2, and could the high CO_2 level allow one to discriminate against models that then predict a runaway greenhouse climate? It is often impractical to address such issues with GCMs, so we attempt to offer some answers to these questions using a one-dimensional energy balance model.

As in lecture 7, the main ingredient to the model is the balance between the solar heat input and the outgoing long-wave radiation (OLR). The solar heating is determined by the total incoming radiation (S), which was 6% less during the neoprotozoic than today, and the albedo (α), which is in turn a function of land fraction and cloud parameterization. The OLR is a function of the Earth's surface temperature (T), the atmospheric CO_2 (about 20% for the neoprotozoic) and the (constant) relative humidity. Unlike in lecture 7, we simplify the model somewhat by fixing the latitudinal structure of the temperature to be uniform over the tropics and parabolic in $y = \sin\phi$ (with ϕ latitude) elsewhere; see figure 1. The equator-to-pole temperature gradient (DT) and the latitude of the ice margin are then variables of the model. However, as in lecture 7, we treat the latitude of the ice margin as a free parameter, compute the thermal structure of the model and search for the special location of the ice margin that gives the corresponding temperature to be 273 degrees Kelvin. The pole-equation temperature difference DT is thereby determined by the global energy balance.

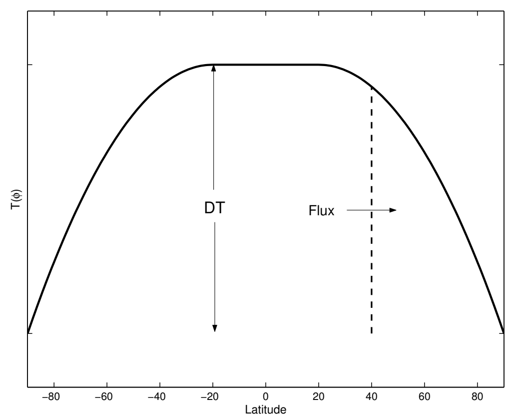

Figure 1: Temperature profile used in the one dimensional energy balance model for the snowball Earth.

Figure ??? shows the temperature at the ice margin for computations with different levels of atmospheric CO_2 and different cloud parameterizations. *This figure is not yet available.* As in lecture 7, for several cases there are multiple equilibria consisting of solutions with polar ice caps (with a finite ice margin latitude whose temperature is $273K$), and a completely ice-covered snowball (for which the ice margin is at zero latitude and the temperature is less than $273K$). Other computations reveal no equilibria other than the snowball; a frozen planet would inevitably result under the corresponding conditions. Unfortunately, as with all climate models, the parameterization of clouds represents the largest source of error and uncertainty. However, the model suggests that a snowball Earth is possible. The main physical ingredient needed seems to be a significant contribution to the albedo from the clouds.

If the Earth does becomes a snowball, how does it escape this snowball climate? In order to deglaciate, the ambient conditions must change so that the system is kicked away from the snowball solution and proceeds to another solution such as a stable ice-cap solution or

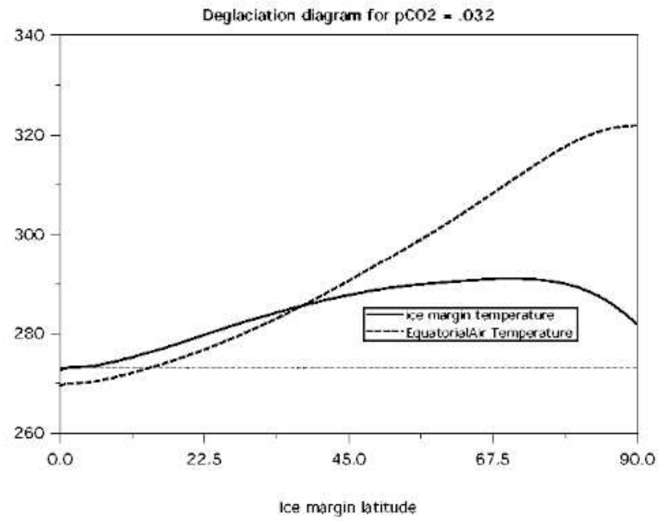

Figure 2: The possibility of deglaciation: A completely snowball Earth subject to large levels of CO_2.

an ice-free one. For example, as shown in figure 2, by building up the atmospheric CO_2, we can induce the disappearance of the snowball solution and a runaway deglaciation. The computation shown is on the brink of deglaciation – the slightest perturbation or further increase in CO_2 would open up an ice-free equatorial region that would permit more heating and deglaciate the planet completely. The equatorial temperature would eventually settle down to a warm $320K$.

Thus with different settings for the cloud physics and CO_2, it certainly seems plausible to evolve the Earth into a snowball state, and then deglaciate it. One unappealing feature of the model is that the settings required for each event are incompatible. The main conclusion from the toy model is therefore that the neoprotozoic snowball scenario is conceivable, if critically dependent on the cloud parameterization. Such a sensitivity does not bode well for the robustness of results from GCMs, which use a variety of such parameterizations.

3 Early Mars

Now we turn to the climates of other planets, and focus on Mars for which recent space missions have provided a wealth of new information. One of the most significant results is that there is now fairly conclusive evidence that there was once a flowing liquid on the surface of this planet. For example, Figs. 3 and 4 show photographs of features reminiscent of river valleys and catastrophic flood plains, and Fig. 5 outlines the large, flat polar area that resembles an ocean floor. Also notable are apparently glacial landforms.

The flood features can be dated to be about 4 billion years old. But Mars today is too cold to have running water, so could it have been warm enough in the past, particularly given that the sun's output then was approximately 70% of its current value? The answer might

Figure 3: Figure showing possible water features on Mars.

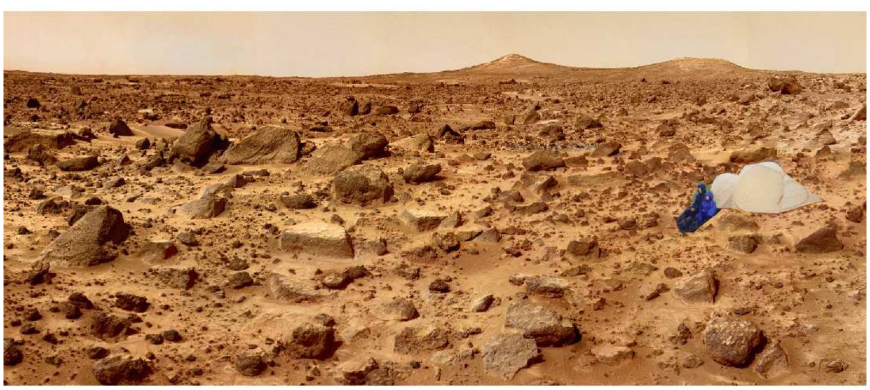

Figure 4: Figure showing the Martian surface (complete with principal lecturer and family).

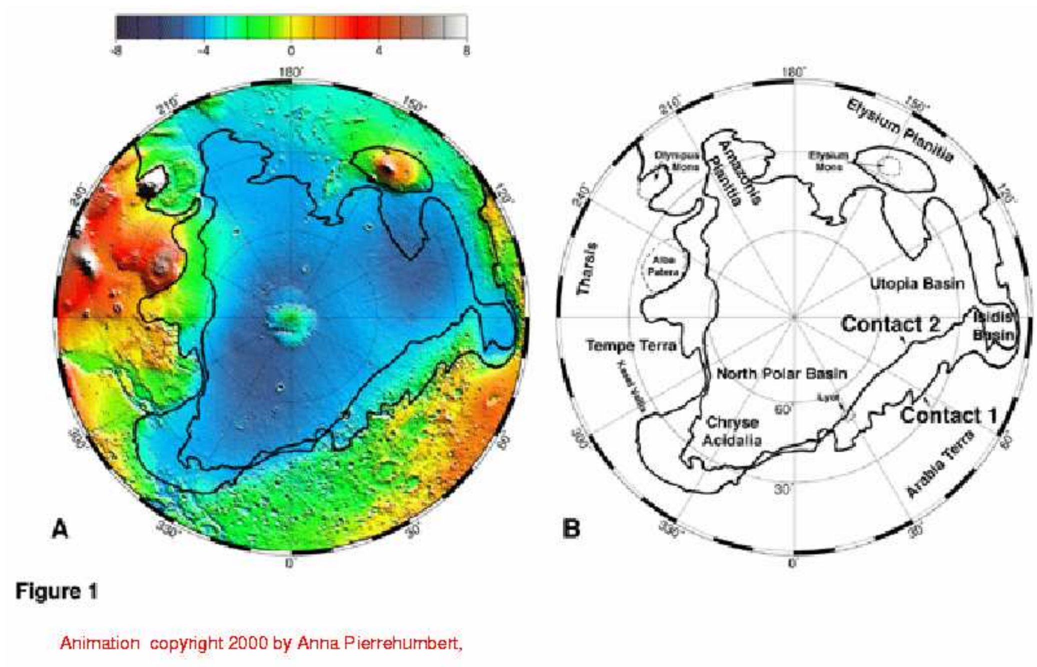

Figure 5: Laser altimetry data showing a possible polar ocean.

lie in CO_2: It is thought that early Mars may have had an almost pure $\mathrm{CO_2}$ atmosphere, which may have been instrumental in raising the temperature sufficiently to allow water to flow on the surface. Somewhat later, Mars must have lost this atmosphere, leaving the planet as it is today.

To decide how tenable such an explanation is, we need to think more carefully about the structure of the early Martian atmosphere and the possible climate dynamics associated with it. One important ingredient is that $\mathrm{CO_2}$ condenses (to solid dry ice) within the temperature range of the Martian atmosphere. By analogy with the atmospheric structure of the Earth's tropics, we might then expect that the thermal stratification would follow some analogue of the moist adiabat in the layers where CO_2 can condense. More specifically, the Clausius-Clapeyron relationship for $\mathrm{CO_2}$ between the condensation temperature, T_c, and the pressure, p, can be written in the form,

$$T_c(p) = \frac{3148}{23.02 - \ln p} \tag{1}$$

(p in mbar), which plays the role of the moist adiabat for early Mars.

By suitably modifying the radiative energy balance models described in earlier chapters to incorporate this and other physics of the Martian atmosphere, we can proceed to explore whether the surface was ever warm enough to support liquid water. The results suggest that no matter how much $\mathrm{CO_2}$ is put into the Martian atmosphere, the temperature never rises above the freezing point of $\mathrm{CO_2}$ on the surface; the highest attainable temperature is 220K with 2 bars of surface pressure. However, these computations ignore the Martian analogue of clouds.

The condensation of CO_2 could in principle also generate clouds of dry ice. Such clouds scatter large amounts of infrared radiation, and might significantly warm the surface. Indeed, the incorporation of parameterizations of dry-ice clouds in the radiative balance models suggests that for a surface pressure of 2 bar it is possible to warm the Martian surface temperature to 300K. (It is not necessary to include water vapor and nitrogen in such a model, however, the increase in temperature is more dramatic if they are included.) Thus, with dry-ice clouds, we might be able to achieve our goal of warming the surface above freezing so that water flow can shape the early Martian landscape.

The next step is to explain how early Mars evolved. The cloudy model requires a surface pressure of at least 2 bars to achieve sufficiently high surface temperatures. But Mars today has only 6 mb of surface pressure, so where did it all go? One explanation resides in the dynamics of immense CO_2 glaciers that may have existed at the Martian poles.

Given current Martian surface temperatures, one expects that CO_2 only condenses well above the surface near the equator. However, surface temperatures decrease with latitude, and eventually fall beneath the condensation temperature. Poleward of this margin, CO_2 snow falls from the atmosphere, and the planet surface could become covered by dry ice. In principle, large quantities of $\mathrm{CO_2}$ could be stored in such ice caps (in fact, the whole atmosphere, if condensed, could be contained in a 1km high glacier), leading to a delicate mass balance between the atmosphere and the glaciers. This delicate balance could easily be upset by greenhouse and albedo feedback effects, the result of which could be the runaway to the current Martian climate (in which neither ice cap is pure CO_2, and the north polar ice cap is, in fact, predominantly H_2O).

The glaciers are, however, restricted in thickness: At typical subglacial temperatures, and for pressures greater than about 5 bars, instead of solidifying to dry ice, $\mathrm{CO_2}$ is forced into its liquid form. Such pressures first occur at the base of a glacier with a depth of about 100m due to the weight of the overlying dry ice. This liquid $\mathrm{CO_2}$ flows toward the ice-margin until the pressure decreases sufficiently for it to re-freeze. However, the liquid layer lubricates the base of the glacier, and should the ice become any thicker, the glacier may well slide freely (surge) to lower latitudes. This action redistributes the mass of the glacier, causing the shape to become more rectangular, and provides a dynamical control on the glacier thickness. It is also notable that, unlike water and ice, the density of solid $\mathrm{CO_2}$ is greater than that of liquid CO_2. Thus glacier fragments sink into the liquid and perhaps melt rather than float like terrestrial icebergs. All this suggests the intriguing possibility that an entirely different kind of glaciology existed in the early Martian Chronicles.

Given these rough ideas, we can also build a simple energy balance model for Mars including the CO_2 ice caps. The methodology is somewhat similar to the models of the neoprotozoic snowball Earth, but with the additional novelties that a change in the location of the ice margins also changes the amount of atmospheric CO_2 (and therefore the OLR), and that the temperature of the ice margin should be given by the condensation temperature appropriate to the specific surface pressure, $T_c(p)$, as in (1).

Results from such a model are shown in Figs. 6 and 7. A twist in the solution of the current model is that the ice margin is specified by assuming the toe of the glacier to be at the freezing temperature. Then the energy budget is calculated - the albedo and greenhouse effect being affected by the ice margin's location. Fig.6 shows the global surface pressure plotted against ice-margin position. This plot confirms the expected mass balance - when

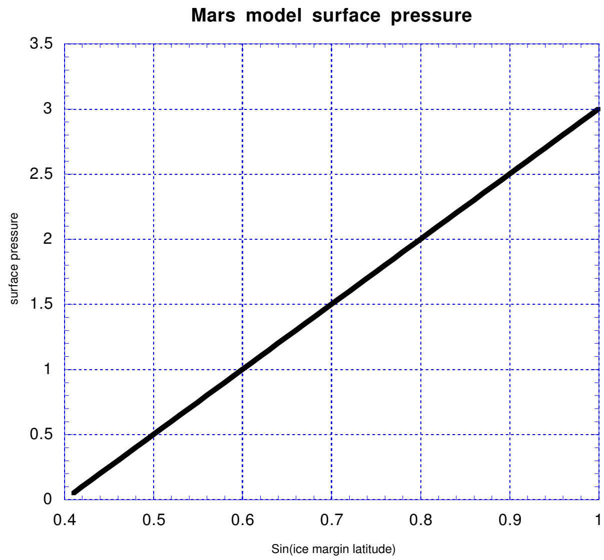

Figure 6: Plot of surface pressure against sine of the latitude of the ice margin.

the ice margin is at the equator, practically all the CO_2 is captured in the ice sheet and thus the pressure of the atmosphere approaches zero. As the ice-sheet recedes back to the pole, CO_2 is reinjected into the atmosphere and the pressure increases accordingly.

Fig.7 shows the global net heat flux (incoming minus outgoing radiation) against the position of the ice margin. The fixed points of the system are where the plot crosses the line of zero net flux. There are two such equilibria; the right-hand fixed point is unstable, whereas the left-hand one, a low-latitude glacier, is stable. These stability characteristics follow because a positive flux perturbation at the high-latitude glacial state corresponds to an increase in the latitude of the glacier margin. But such a flux perturbation also leads to a warming of the climate which provokes further recession of the glacier to the pole. Conversely, for the low-latitude equilibrium glacier, the introduction of a similar flux excess increases the glacier margin, and the subsequent heating of the climate melts the margin back to its original position.

To summarize, it is possible that early Mars was warmer and wetter in the past. With less CO_2 stored in the glaciers, the combined effect of greenhouse gases, reduced albedo and cloud dynamics can produce a ground temperature high enough to allow liquid to flow over the surface to shape the land. Subsequently, the dynamics of the CO_2 glaciers could have played an important role in the evolution of the Martian climate to its current state.

Notes by Matt Spydell, Fiona Eccles, and Helén Andersson

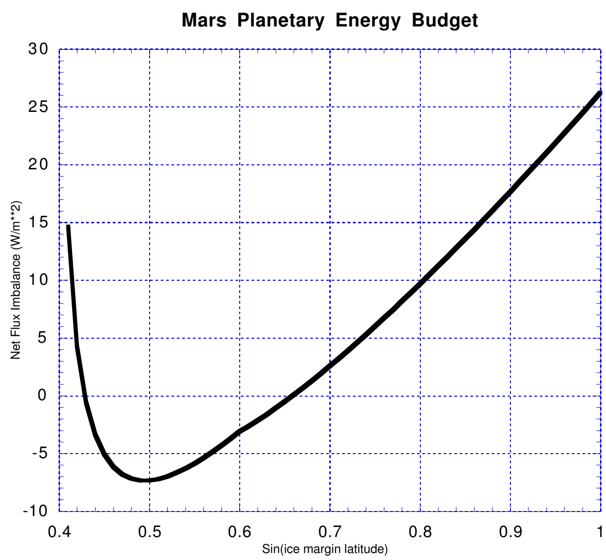

Figure 7: Plot of net heat flux (incoming-outgoing radiation) against sine of the latitude of the ice margin.