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Detecting secular climate change on Mars

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Abstract

Background: The South Polar Residual Cap on Mars, a reservoir of CO_2 ice, appears to be losing mass. Malin et al. (2001) put an upper limit on the loss rate at ~1% of the present atmospheric mass per Mars Decade. Blackburn et al. (2010) model the loss and find an even higher loss rate. We refer to this phenomenon as secular climate change. The atmosphere is the most likely reservoir to take up the cap loss. We seek to detect the signal of this loss in surface pressure data.

Method: We carefully examine the surface pressure data from the Viking, Pathfinder, and Phoenix Landers with due note of their measurement uncertainties. We use standard hydrostatic methods to account for elevation differences, and the NASA/Ames Mars general circulation model to estimate the effect of winds on pressure gradients.

Conclusion: We find that Phoenix surface pressures are ~10 Pa higher than Viking surface pressures after correcting for elevation differences and dynamics. This difference is consistent with the expected change based on the Malin et al. maximum erosion rate, but less than that predicted by Blackburn et al. However, the combined uncertainties in the data and our modeling methodology are large enough that we cannot confirm that secular climate change is occurring on Mars. Should the trend continue, however, future landers that carry well-calibrated pressure sensors with absolute accuracies of ~4 Pa or better could unambiguously detect secular climate change on Mars.

Introduction

The familiar polar caps on Mars that grow and recede with the seasons are composed of frozen carbon dioxide. When these seasonal caps disappear during summer they expose an underlying residual water ice cap in the north, and a permanent CO_2 ice cap in the south. Both of these "residual" caps survive year after year. The north polar residual cap is a major source for atmospheric water vapor and is the main driver of the present water cycle. The south polar residual cap (SPRC), however, may actually regulate the mean annual mass of the atmosphere.

Leighton and Murray (1966) were the first to recognize this possibility. Since the atmosphere is composed mainly of carbon dioxide, they reasoned, it must be in vapor pressure equilibrium with any permanent CO_2 ice deposits on the surface. On centennial time scales, they showed that the heat balance of the SPRC therefore determines the mean annual surface pressure and that if there are vast quantities of CO_2 locked up in the SPRC, changes in orbital parameters could lead to large swings in the mean annual surface pressure. This, of course, implied that Mars had the potential to experience significant climate change on astronomical time scales (*e.g.*, Toon et al. 1980).

We now know that the ability of the SPRC to buffer the atmosphere is much more limited than envisioned by Leighton and Murray. High resolution detailed mapping of the materials in the SPRC show it to consist of no more than about 3% of the present atmospheric mass (Thomas et al. 2009). Thus, if the heat balance favored complete sublimation, the disappearance of the SPRC would raise mean annual surface pressures by less than 20 Pa. Such an increase would have minor climatic effects.

Nevertheless, the existence of the SPRC is a mystery. Though we now have some understanding of why it is offset from the pole (Colaprete et al. 2005), we don't really understand why it exists at all. The somewhat glib answer is that its albedo is coincidentally high enough to prevent complete sublimation every summer season. Perhaps this is the answer. But only a 0.1% change in the long-term albedo can lead to a significant (1%) change in atmospheric mass (Paige 2001), so if the net annual mass of the SPRC is not changing with time, it means the SPRC albedo is finely tuned to a single value that must be maintained by some feedback process against the interannual vagaries of meteorological variability. Such a feedback mechanism may exist, but it would be astonishing if it operated on short enough time scales to yield no annual mean change at all.

Recent observations indicate that the SPRC is, in fact, changing from year to year and that overall it has been losing mass for at least the past 20 Mars years or so, and possibly for as long as a Mars century (Thomas et al. 2009). The loss is not necessarily monotonic. Some years may experience net deposition of SPRC materials (Thomas et al. 2005; Thomas et al. 2009), and global dust storms can cause local changes in frost coverage as well (Bonev et al. 2008; James et al. 2010). But overall, the SPRC appears to be disappearing.

The main evidence for this comes from the scarp retreat rates of circular depressions in the SPRC that have been imaged in successive years. Malin et al. (2001) examined high resolution Mars Observer Camera (MOC) images of the SPRC taken from the Mars Global Surveyor (MGS) during the 1999 and 2001 viewing opportunities and found the scarps to be retreating at a rate of 1-3 m per Mars year. Given these retreat rates, a range of plausible values of the CO_2 ice density, and a calculation of the total scarp perimeter, Malin et al. (2001) estimated the net annual loss to be between 0.008% to 0.08% of the present atmospheric mass per Mars year or, equivalently, a change in global mean annual surface pressure of between ~0.5 and ~5 Pa per Mars Decade (MD). Follow on imaging by the Context Imager on the Mars Reconnaissance Orbiter (MRO) has expanded the spatial and temporal coverage begun by Malin et al. (2001) and confirms that the SPRC continues to erode (Thomas et al. 2009). While the cause of the erosion is not yet known, these observations clearly indicate that if a stabilizing feedback mechanism does exist, it operates on time scales much longer than a Mars year.

More recently, however, Blackburn et al. (2010) model the stability of the SPRC and find its average thickness to be decreasing at a rate of ~0.4 m per Mars year. This is equivalent to a net annual loss of ~13 Pa/MD, more than double the maximum rate estimated by Malin et al. (2001). Blackburn et al. (2010) find several examples where MOC and HiRISE images overlap that reveal a change in thickness almost identical to their model prediction. Thus, multiple observations and at least one model point to an eroding SPRC with loss rates of between 0.5 - 13 Pa/MD.

The purpose of this paper is to search for the signal of an eroding SPRC primarily in surface pressure data, which contain a great deal of information about the meteorology and climate of Mars (*e.g.*, Leovy 1981). We use the term "secular climate change" to describe this phenomenon and define it to mean a multi-decadal non-periodic change in surface pressure driven by an unstable SPRC. We begin with a modeling/theoretical exercise to make the case that most of the disappearing CO_2 must be going into the atmosphere. As a consequence the most promising way to corroborate the imaging observations is to find an increase in surface pressure from year to year. For this purpose we carefully examine the available surface pressure data making due note of the challenges involved in detecting such a small signal. We conclude that while there does appear to be a signal, it does not rise above the uncertainties in the measurements and/or our methodology. Therefore, we cannot say that the available pressure data confirm the erosion story.

What we can say, however, is that in principle it is possible to detect secular climate change on Mars from surface pressure measurements and we show how this can be done. Our main conclusion is that we can apply these methods to future landers and therefore strongly recommend that they carry stable, well-calibrated pressure sensors. The Mars Science Laboratory (MSL) represents the next opportunity to conduct surface pressure measurements, and these can further test the notion of secular climate change on Mars.

Where is the CO_2 going?

There are three likely exchangeable reservoirs for uptake of the disappearing CO_2: the atmosphere, the seasonal polar caps, and the regolith. For several reasons, we believe the regolith is not likely to be a major sink. On the short time scales considered here (~1 MD), regolith pore sizes would have to be unrealistically large for CO_2 to diffuse to significant depths and adsorb on subsurface grains (Toon et al. 1980). More problematic, however, is the fact CO_2 would have to compete with water for adsorption sites (Zent and Quinn 1995). Given that the atmosphere contains some 10-15 pr-microns on average (Farmer and Doms 1979; Smith 2008), water molecules will occupy many of the seasonally accessible adsorption sites. Furthermore, water ice is known to be within centimeters of the surface at the middle and high latitudes (*e.g.*, Boynton et al. 2002; Feldman et al. 2002; Mitrofanov et al. 2002), which would preclude any CO_2 adsorption below those depths at all. Thus, an actively exchanging CO_2 regolith reservoir is unlikely.

The seasonal polar caps could take up some of the released CO_2, but they too are not likely to be a major sink. As the surface pressure increases, more CO_2 will condense on the winter caps because the caps will warm. This counterintuitive result is a consequence of the fact that to good approximation, the latent heat released during CO_2 condensation balances the net radiative loss of the caps, which is dominated by thermal emission (*i.e.*, $\varepsilon\sigma T^4$, where ε is the cap emissivity, σ is the Stephan-Boltzmann constant, and T is the cap temperature. See Paige and Ingersoll (1985) for a complete discussion of the cap heat balance.) Since T is set by the vapor pressure relationship, higher vapor pressures mean higher cap temperatures and, hence, higher condensation rates. However, the change in T for the expected changes in surface pressure is small enough that even with the condensation rate dependent on the fourth power of the temperature, the increased mass of the seasonal caps is likely to be small fraction of the total loss of material from the SPRC.

On the other hand, the caps do occupy a considerable fraction of the planet's surface. To get a reasonable quantitative estimate of this partitioning, we therefore ran a fast C-grid version (*v*1.7.3) of the Ames Mars General Circulation Model (GCM) for 30 Mars years. The model was run at 5°×6° latitude-longitude resolution (with 24 vertical layers) using the parameters that Haberle et al. (2008) found gave a good fit to the Viking pressure data when subsurface water ice was present. Kahre et al. (2006) describe this version. We placed an infinite reservoir of CO_2 ice at the model grid points that most closely match the observed distribution of the SPRC (see Figure 1) and, to reduce computer time, gave this artificial reservoir an albedo (0.76) which resulted in a net annual global equivalent loss of 60 Pa per MD, a rate 10 times faster than suggested in Malin et al. (2001). After 5 Mars years, the model reached a quasi-steady state in which the global mean annual surface pressure was increasing by a steady rate of 48 Pa/MD, while the seasonal caps were taking up the remaining 12 Pa/MD. Thus, the atmosphere takes up about 80% of the CO_2 annually released by the SPRC in this simulation. (The change in surface temperature with surface pressure for the last 10 Mars years of this simulation can be found in tchange.txt.)

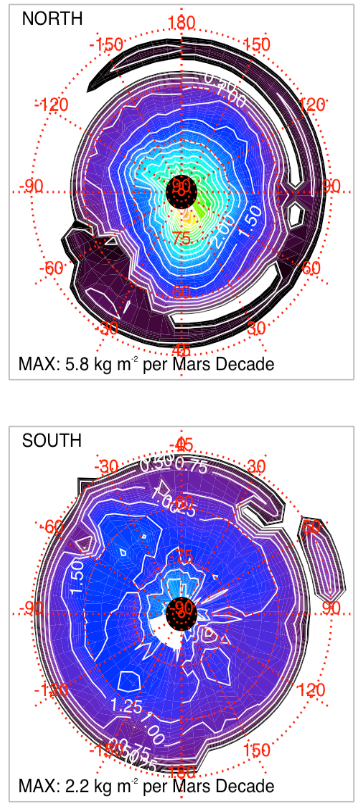

Figure 1. Change in cap mass at winter solstice (kg m^{-2} per Mars Decade). Top: North cap at L_s=270; bottom: South cap at L_s=090. This version of the model does not have a grid point at the poles (filled black circles). White fan shaped area near the South Pole represents the SPRC. (figure1.png)

Are these changes detectable?

Polar caps

While most of the CO_2 is going into the atmosphere, the change in seasonal cap mass is worth discussing. As shown in Figure 1, the mass of the seasonal polar caps at their maximum extent increases at all latitudes for both hemispheres. As expected, the increase is greatest at the poles, which have a longer condensation season, and monotonically decreases toward the equator. There is, however, a distinct hemispheric difference: most of the CO_2 is going into the north cap. In this simulation it accounts for about 70% of the mass taken up by the seasonal caps. Elevation differences account for this asymmetry. In the model, the north polar region lies as much as 5.5 km below the south polar region and therefore has a higher mean annual surface pressure. Higher surface pressures increase the frost point temperatures, the net radiative loss (through σT^4), and total condensation. Thus, the north cap is favored for accumulating CO_2 as was noted, for example, by James et al. (1992).

In principle, these changes could be detected in the gravity field and/or ice cap depths. Smith et al. (2009) report the detection of time variations in four Mars years of MGS gravity data that they relate to the seasonal CO_2 cycle. They even find hints of interannual variability in these data. However, four Mars years would lead to an increase in maximum cap mass well below the precision of their data. A much longer time series is needed. Alternatively, the changes in cap mass might be detectable through changes in ice cap depth. Assuming a CO_2 snow density of 1000 kg m^{-3}, the maximum increase in the depth of the ice caps would be about 6 mm per Mars decade in the north, and several mm per Mars decade in the south (note we are accounting for the fact that the model RSPC is losing mass at 10 times the expected rate). MOLA measured changes in ice cap depths during the MGS mission (Smith et al. 2001). Unfortunately, its accuracy is not capable of measuring elevation changes on the order of mm per Mars year, and it did not survive long enough to search for multi-year changes in ice cap depths. Thus, the MGS gravity and MOLA elevation data sets either do not have the longevity or precision to detect evidence for secular climate change.

A final point here is that though the poles are clearly the place to look for yearly changes in cap thickness, there are longitudinal variations in cap growth worth noting. In the model simulation for the northern polar region, the longitudinal sector centered on the prime meridian accumulates more CO_2 than at other longitudes. In the south polar region, the accumulations are more zonally symmetric, though there is a preference for the western hemisphere between 0° and -90° longitude. Hellas also is a favored site because of its low elevation.

Atmosphere

If most of the disappearing CO_2 is going into the atmosphere, then it should be detectable in surface pressure data. Under hydrostatic conditions, an excellent assumption for the Martian atmosphere, surface pressure provides a measure of the mass in an atmospheric column. Thus far, there have been three landed missions to Mars with payloads containing pressure sensors: Viking, Pathfinder, and Phoenix. The pertinent information on these missions is summarized in Table 1.

Table 1. Mars lander missions with pressure sensors.

Mission	Operating Dates	Mars Year[a]	Location	Elevation (m)	Pressure Change[b] (Pa)
VL-1	07/20/76 to 11/13/82	12.31 to 15.67	22.27°N 312.05°E	-3637	0.13-3.6
VL-2	09/03/76 to 04/12/80	12.38 to 14.29	47.67°N 134.28°E	-4495	0.08-2.1
Pathfinder	07/04/97 to 09/27/97	23.45 to 23.58	19.10°N 326.75°E	-3682	0.45-12
Phoenix	05/25/08 to 11/02/08	29.24 to 29.48	68.22°N 234.25°E	-4126	0.69-18

[a]Mars years are numbered according to a calendar proposed by Clancy et al. (2000) with Mars Year 1 beginning on April 11, 1955 (L_s=0). [b]Range of expected pressure changes (given to two significant figures) since beginning of VL-1 mission (see text for details).

The elevation data were obtained by carefully registering HiRISE lander images onto THEMIS thermal images and then to the MOLA grid. The gridded MOLA data were then contoured to the 20 m level and interpolated to get the lander elevations. The use of HiRISE images significantly reduced the uncertainties in lander locations and therefore improved elevation estimates to ± 5 m, which corresponds to an uncertainty in surface pressure of ± 0.3 Pa. Thus, uncertainties in the lander elevations are small compared to uncertainties in sensor calibration.

The final column of Table 1 gives the range of expected changes in global mean surface pressure since VL-1 began surface operations. We arrive at these numbers by multiplying the Malin et al. (2001) lower limit for the erosion rate (0.5 Pa/MD) and the Blackburn et al. (2010) modeling estimate (13 Pa/MD) by the time interval between the beginning of VL-1 surface operations and the end of operations of the mission in question. We then multiply by 0.8 to account for the fact that some of the CO_2 is going into the seasonal polar caps as suggested by our simulations.

Given these assumptions, the maximum expected pressure change during the Viking mission would be less than 4 Pa. This is smaller than the observed interannual variability (Paige and Wood 1992) and roughly comparable to the expected accuracy of sensors (thought to be better than 4 Pa (Chamberlain et al. 1976; Tillman et al. 1993)). Thus, it is not possible to search for a secular trend in the Viking data.

The Pathfinder mission ended 11.27 Mars years after VL-1 began surface operations, which would correspond to an increase during that time of between 0.45 and 12 Pa. Unfortunately, the Pathfinder pressure sensor had major calibration issues and cannot be used for the purpose of detecting absolute changes in surface pressure. Haberle et al. (1999) give the details in the appendix of their paper.

Phoenix, on the other hand, did carry pressure sensors whose calibration uncertainties are small enough to warrant a close look. The major calibration issue with the Phoenix sensors concerns diurnal pressure variations as the sensor thermal environment was much different than the pre-launch calibration environment. Specifically, the pressure data are thought to degrade when the sensor head temperature exceeds 0°C (Taylor et al. 2010). However, the unexpected thermal environment has less of an effect on daily-averaged surface pressures and worst case estimates for sensor accuracies are 7 Pa at the beginning of the mission and 15 Pa at the end of the mission. Further testing and analysis may reduce the uncertainties to less than 10 Pa, which is small enough to see the change predicted by Blackburn et al. (2010), though not by significant margin. Nevertheless, the Phoenix data merit close examination.

The daily averaged surface pressure data from Phoenix and VL-2 for the same seasonal period are shown in Figure 2. We chose VL-2 for the initial comparison because it is closest to Phoenix, has a smaller elevation differential, and is in approximately the same thermal and dynamical regime as Phoenix. Note that, as expected, VL-2 pressures are higher than Phoenix pressures, because VL-2 sits at a lower elevation than Phoenix (see Table 1). However, when we correct for elevation differences assuming a constant scale height of 10 km (~200 K), the Phoenix data are systematically higher than VL-2 data by ~10 Pa. In other words, there appears to be an offset in mean surface pressures between Viking and Phoenix. Taylor et al. (2010) found a similar systematic offset. This offset is consistent

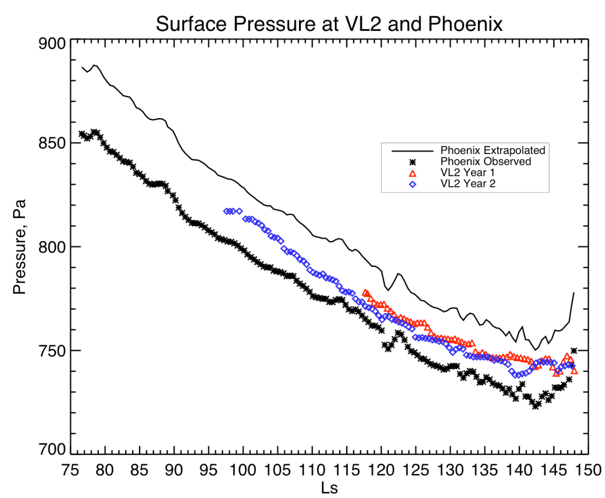

Figure 2. Daily-average surface pressure (Pa) at the Phoenix and VL-2 lander sites. Symbols denote Phoenix data (black asterisks), VL-2 year 1 data (red triangles), and VL-2 year 2 data (blue diamonds). Solid line is the Phoenix data hydrostatically extrapolated to the VL-2 site assuming a 10 km scale height. (figure2.png)

with the expected secular pressure change between the Viking and Phoenix missions if the SPRC is eroding at the maximum rate estimated by Malin et al. (2001) (~ 5 Pa/MD), but somewhat less than that expected using the Blackburn et al. (2010) rate.

However, this correction for elevation differences is a rough approximation. In reality the scale height changes with season, and dynamical processes can have an effect as well (Hourdin et al. 1993; Hourdin et al. 1995). The scale height changes with season because temperatures change, and dynamical processes affect pressures because winds are driven by pressure gradients. The combined effect can be significant. Therefore, a better approach to estimating the pressure correction would be to use output from a Mars GCM. GCMs are at least self-consistent in the wind and temperature fields they predict, and to the extent they reproduce the relevant observations, surface pressures and air temperatures, they give us an improved method to estimate the correction.

Again, we use the Ames GCM for this purpose and show in Figures 3 and 4 how the model compares to the Viking and Phoenix data. (This version of the model is based on our B-grid as described in Haberle et al. 2008.) The model data are taken from the grid point nearest the lander sites. Surface pressures are hydrostatically adjusted to account for the elevation difference between the model grid point and actual lander elevation. The adjustment uses the model-predicted daily averaged temperature of the lowest layer (~5 m) to compute the scale height. This daily averaged simulated temperature, and hence the computed scale height, varies with season and is assumed to be independent of altitude when correcting the surface pressures. Note that the measured temperatures were made several meters below the predicted temperatures, and thus the comparison is not completely straightforward. Lapse rates in this part of the atmosphere can be quite high in late afternoon and early morning (Schofield et al. 1997). These lapse rate effects are minimized when averaged over a day, however, so that the change in daily averaged temperature with altitude over such a small distance should be much less than a Kelvin.

The model compares very well with the measured pressures and temperatures at VL-2. The main differences are in the degree of interannual and day-to-day variability. In the observations, year 1 pressures are several Pa higher than year two, and year 1 temperatures are several Kelvin lower than year 2. Daily average temperatures fluctuate by as much as 10 K at VL-2 during the early part of the summer season in year 2. The model produces very little interannual variability (not shown), and virtually no day-to-day variability at any point during the Phoenix season. Variations in dust loading, and sensor drift, are possible explanations for the observed interannual variability, while small-scale circulations (which are not represented in the model) may account for the day-to-day variability.

At the Phoenix site the model pressures compare very well

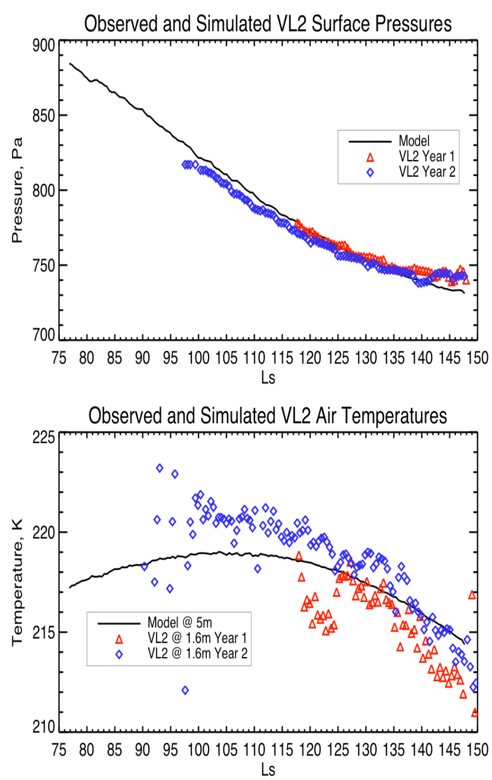

Figure 3. (a) Daily-average GCM surface pressure (Pa) and (b) 5-m air temperature (K) from the grid point closest to the VL-2 lander site. Symbols represent observations. (figure3.png)

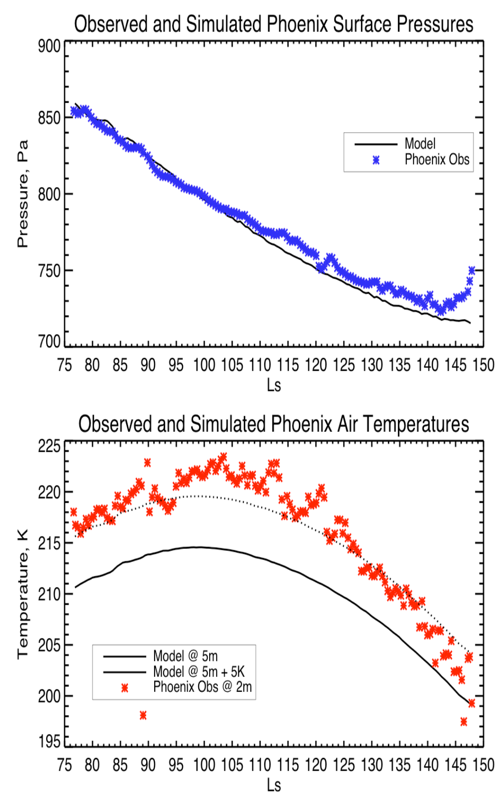

Figure 4. (a) Daily-average GCM surface pressure (Pa) and (b) 5-m air temperature (K) from the grid point closest to the Phoenix lander site. Symbols represent observations. (figure4.png)

with the observations, however temperatures are systematically about 5 K colder than observed. We are not certain about the reason for this discrepancy, but tentatively attribute it to the use of incorrect aerosol scattering properties (*e.g.*, we do not account for water ice clouds, and our dust particles are too absorbing), an excessively high surface albedo, or errors in the nature and distribution of subsurface ice at this particular site (which affects the soil thermal conductivity). The dotted line in Figure 4, which does fit the data well, represents the model temperatures increased by 5 K. We use these "adjusted" temperatures to compute the scale heights. While this adjustment to the model temperatures is arbitrary, it is justified for the present purpose.

Perhaps the more important comparison is the pressure difference between lander sites. By comparing simulated pressure differences with observed differences we can eliminate model offset errors (*i.e.*, model errors in the absolute value of surface pressure) and these can be applied to the observations to provide an improved elevation correction. The observed and simulated pressure differences between VL-2 and Phoenix are shown in Figure 5. Model pressures are first hydrostatically adjusted to the actual lander elevation using the 5-m daily averaged air temperatures to calculate the scale height, and then differenced. The model-predicted pressure differences are systematically greater than the observed differences, particularly at the beginning of the Phoenix mission. The model shows a systematic decline and small day-to-day variations in pressure differences during this season, whereas the observations have rather large day-to-day variations and very little overall trending.

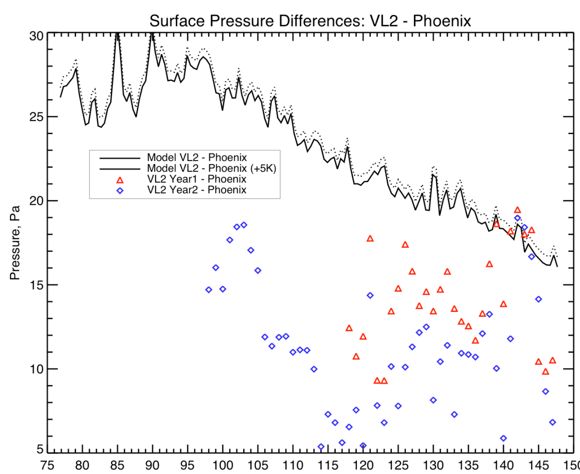

Figure 5. Observed and simulated pressure differences between VL-2 and Phoenix (Pa). Symbols are the observations, which were first binned into 1° L_s bins, then differenced. Solid and dotted lines are model pressure differences. Solid line is for scale heights calculated from predicted model temperatures; dotted line is for scale height calculated from model temperatures increased by 5 K, which gives better agreement with the Phoenix temperature data shown Figure 4. (figure5.png)

Adding the model-predicted pressure differences to the Phoenix data gives our best estimate of the change since Viking. As can be seen in Figure 6, surface pressures at the VL-2 site would be systematically higher than in the late 1970s when VL-2 began operations with an average difference over the Phoenix season of +9.5 Pa. A similar exercise extrapolating Phoenix to the VL-1 site, shown in Figure 7, gives an average difference of +13.6 Pa, though the uncertainties here are greater because of the greater elevation difference. In general, these results imply that the model cannot simultaneously match the Viking and Phoenix data as would be expected if there has been no change in mean annual pressure between the two missions.

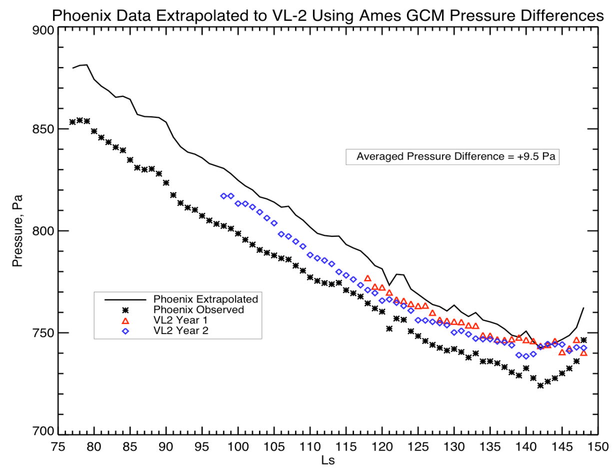

Figure 6. Phoenix daily-average surface pressures (Pa) extrapolated to the VL-2 site using the model pressure differences given by the solid line in Figure 5. For the interval of time during which there are VL-2 and Phoenix data (~L_s = 98 - 147) the average difference is 9.5 Pa. (figure6.png figure6.txt)

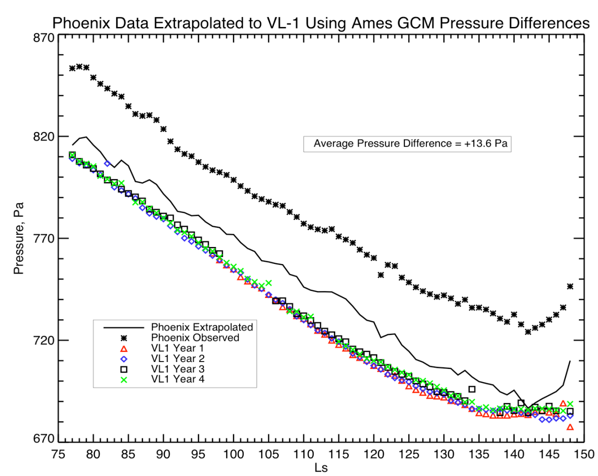

Figure 7. Phoenix daily-average surface pressures (Pa) extrapolated to the VL-1 site using the model pressure differences between VL-1 and Phoenix (not shown). For the interval of time during which there are VL-1 and Phoenix data (~L_s = 77 - 147), the average difference is 13.6 Pa. (figure7.png figure7.txt)

Discussion and conclusion

While the measured pressures and modeled pressure differences imply a roughly 10 Pa increase in surface pressures since the Viking mission, it is not clear how to interpret this result. The 10 Pa offset is less than the presently estimated worst case uncertainties in the Phoenix data (< 15 Pa), and when coupled with the uncertainties in the Viking data (< 4 Pa), lander site elevations (< 0.3 Pa), and model hydrostatic corrections for the difference between the grid point and lander elevations (< 4 Pa (estimated)), it does not rise enough above these combined uncertainties to draw firm conclusions about secular climate change on Mars. However, it does appear inconsistent with the Blackburn et al. (2010) erosion rate. Further refinement of the Phoenix pressure data will help clarify these issues, though the ultimate test will come from future landers.

This exercise clearly points out yet another benefit of measuring surface pressure on Mars. Surface pressure sensors are primarily flown to collect meteorological data and monitor the seasonal CO_2 cycle. Here we have shown that given a long enough record of surface pressure data, it is possible to detect secular climate change on Mars. For this reason, we strongly recommend that future landers carry stable well-calibrated pressure sensors. If spacecraft resources permit, air temperatures should also be measured to help constrain the hydrostatic pressure adjustments that will be needed to account for elevation differences. Wind measurements would also be helpful to constrain dynamical effects. But at a minimum, *pressure sensors should be part of all future landed payloads*; they are small, light, need little power, require minimal data, do not require orientation or deployment, are relatively easy to accommodate, and provide exceptionally high science value, particularly if they can be thermally stabilized.

The next landed mission to Mars, MSL, has a pressure sensor that ground tests have shown may have the needed accuracy to address this issue provided there is little post-launch drift. Since MSL begins surface operations in the fall of 2012, the signal for secular climate change will have risen even further and it should be readily detectable if it rises to the ~15 Pa level. Follow on landers are now being seriously considered for the 2016 (ExoMars Trace Gas Mission), and 2018 (ExoMars Rover Mission) opportunities, with perhaps a geophysics network mission in the early 2020s. With pressure sensors on each of these missions, it should be possible to detect secular climate change on Mars.

Directory of supporting data

root directory
- haberle_mars_2010_0003.pdf
- tchange.txt Surface temperature change vs. pressure
- Fig. 1 figure1.png full-resolution image
- Fig. 2 figure2.png full-resolution image
- Fig. 3 figure3.png full-resolution image
- Fig. 4 figure4.png full-resolution image
- Fig. 5 figure5.png full-resolution image
- Fig. 6 figure6.png full-resolution image
  - figure6.txt numerical data
- Fig. 7 figure7.png full-resolution image
  - figure7.txt numerical data

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