GEOPHYSICAL RESEARCH LETTERS, VOL. 28, NO. 9, PAGES 1779-1782, MAY 1, 2001

Climate and Interior Coupled Evolution on Venus

Roger J. Phillips

McDonnell Center for the Space Sciences and Department of Earth and Planetary Sciences, Washington University, St. Louis, Missouri

Mark A. Bullock

Department of Space Studies, Southwest Research Institute, Boulder, Colorado

Steven A. Hauck, II

McDonnell Center for the Space Sciences and Department of Earth and Planetary Sciences, Washington University, St. Louis, Missouri

Abstract. Climate-interior coupled evolution is investigated for Venus by merging a partial-melting/parameterized mantle convection model with a gray radiative-convective atmospheric model. A positive feedback process can operate by the release of water to the atmosphere via mantle melting, leading to an increase in atmospheric opacity and the radiative temperature gradient. The resulting amplification of the greenhouse surface temperature raises the mantle temperature leading to an increase in the partial-melting rate. Using thin-lid convection, a coupled model for Venus running over an interval of 3 to 1 Ga shows a significant increase in surface temperature, partial-melting extent, and extrusive magma flux compared to a model where there is no communication between greenhouse-modulated surface temperature and partial-melting in the interior. Coupled and uncoupled models that transition (with variable timing) to stagnant-lid convection either shut down partial melting via lithospheric thickening, or evolve to large amounts of melt due to increased interior temperatures.

1. Introduction

It is possible that during some stage in the evolution of a terrestrial planet, a coupled, positive feedback between climate evolution and interior evolution develops. Specifically, the release of volatiles from partial melting can enhance the atmospheric greenhouse and raise the surface temperature. In turn, the enhanced surface temperature may increase the rate of partial melting in the upper mantle, leading to increased volatile release and further intensification of the greenhouse [*Phillips and Hansen*, 1998]. In this paper we explore this process for the planet Venus.

Climate evolution models of Venus [*Bullock and Grinspoon*, 1996, 2001] have relied on assumptions of magma generation rates that are loosely constrained (e.g., assuming the volume magmatic output of a global resurfacing event). Concurrently, models of solid planet evolution have not considered in any quantitative way the effect of climate change and its greenhouse modulation of surface temperature on interior behavior. Two earlier papers, however, [*Solomon et al.*, 1999; *Anderson and Smrekar*, 1999] have dealt with the potential connection between tectonism and climate evolution on Venus. Depending on the strength of coupling (mainly the effect of surface temperature on interior partial melting), the interior-atmospheric interaction may be one-way (magma flux modulates climate, but the feedback has no significant effect on melting) or two-way. Our intent here is not to model a specific thermal history of Venus, though we will contrast different styles of convection (stagnant lid, thin lid). Rather, we wish to focus on the possible physical interrelationships between convection, partial melting, and climate.

Paper number 2000GL011821.
0094-8276/01/2000GL011821$05.00

2. Approach

Our modeling is one-dimensional in the radial variable, r, and uses partial melting to couple parameterized convection models to gray radiative-convective atmospheric models that employ a single trace greenhouse gas, H_2O. Mantle convection can be characterized by two regimes (plus a transition) — "thin lid" and "stagnant lid" [e.g., *Solomatov and Moresi*, 1996]. A thin lid regime can be equated to lithospheric recycling, and the efficient heat transfer results in a thin lithosphere and a mantle of nearly constant viscosity. If the lithosphere is not recycling, then large mantle viscosity variations and a thick, high-viscosity stagnant upper lid results. An equation for conservation of energy is solved along with a relationship between the Rayleigh number (Ra, the dimensionless ratio of buoyant to viscous forces) and the Nusselt number (Nu, the ratio of total to conductive mantle heat transport) [*Stevenson et al.*, 1983; *Phillips and Hansen*, 1998]. It is assumed that the mantle fraction that undergoes partial melting is part of a broad upwelling from internally-heated convection that is 25 K hotter than average convecting mantle. The excess (over average) heat flow in regions of upwelling mantle will thin the thermal lithosphere (thickness Z_L) by an amount we assume to be 10%. Partial melting extends from the zero-melting level upward to this minimum depth of $0.9Z_L$. The enhanced temperature is transformed adiabatically to $0.9Z_L$ and then converted to potential temperature for partial melting calculations. We use standard partial melting relationships [*McKenzie*, 1984; *McKenzie and Bickle*, 1988] to obtain the mass fraction of partial melt in a column (z_m) extending from $0.9Z_L$ to the depth of zero melt. The rate of magma generation then depends strongly on how convection behaves in the upper mantle. Magma is generated by the flux of convecting mantle that passes through a "melting channel" [*McGovern and Schubert*, 1989; *Reese et al.*, 1999] of thickness z_m and convection "cell" diameter d, which is taken as the thickness of the convecting mantle. The flux of water (kg s^{-1} m^{-2}) due to melting is then given by

$$F_{H_2O} = 2\rho u z_m \chi C_{H_2O} / d \qquad (1)$$

where ρ is the density of fluxed mantle, u is the convective velocity obtained from boundary layer theory, χ is the average mass fraction of melt in the channel, and C_{H_2O} is the mass fraction of water in the melt. The value used for C_{H_2O} is the present-day estimate of 50 ppm [*Grinspoon*, 1993], which is conservative, as the venusian mantle must have lost water over geologic time. The melting model assumes a dry solidus, which is adequate to describe melt production for the low concentrations of water assumed here [*Hirschmann et al.*, 1999]. The delivery (kg/s) of water to the atmosphere, Q_{H_2O}, is then set by the surface area of the planet and by assumptions about the fraction of magma that reaches the surface (10%) [*Crisp*, 1984], areal fraction of the planet affected by convective upwelling and partial melting (25%), and the fraction of water associated with intrusive magmas that diffuses to the surface (75%).

The model radiative temperature profile in the atmosphere is given by

$$T_r(z) = T_e\{\tau(z) + \tfrac{1}{2}\}^{1/4} \qquad (2)$$

where T_e is the effective radiating temperature of the atmosphere (232 K) and $\tau(z)$ is the total infrared opacity given by

$$\tau(z) = \frac{H}{k_B T_s}\left(\kappa_{CO_2} P_{CO_2} + \kappa_{H_2O} P_{H_2O}\right) e^{-\frac{z}{H}} \qquad (3)$$

where k_B is the Boltzmann constant, T_s is the surface temperature, H is the atmospheric pressure scale height (6.5 km), κ is an infrared absorption coefficient, and P is an atmospheric partial pressure.

P_{CO_2} is taken to be 89 bars [*Oyama et al.*, 1980], while P_{H_2O} is obtained by scaling a 92 bar atmosphere with the time-varying model volume fraction of atmospheric water. κ_{CO_2} is derived from a detailed nongray greenhouse model constraint that surface temperature should be 522 K with no water in the atmosphere [*Pollack et al., 1980*]; κ_{H_2O} is found by requiring that the model give T_s = 740 K with 30 ppmv H_2O in the atmosphere [*Pollack et al.*, 1993; *de Bergh et al.*, 1995; *Meadows and Crisp*, 1996]. In the troposphere, the convective temperature profile is given by

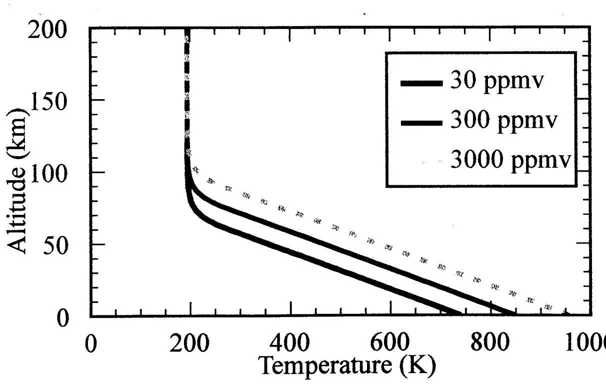

Figure 1. Temperature profiles in the gray atmosphere model for 30 ppmv, 300 ppmv, and 3000 ppmv water in the atmosphere. The lower linear portion of a profile defines the troposphere.

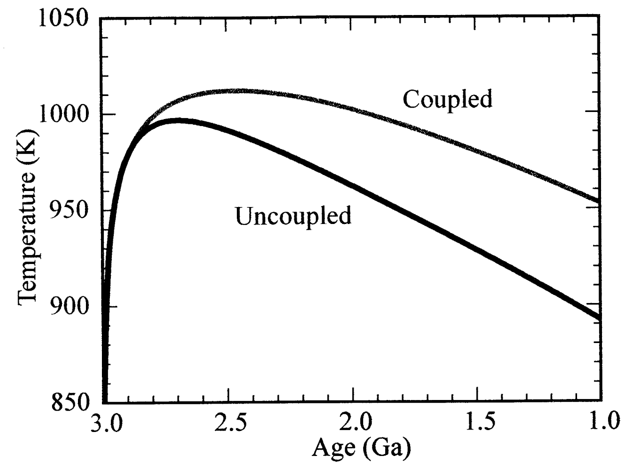

Figure 2. Temperature histories for coupled and uncoupled models illustrating the effects of climate-interior coupling.

$$T_c(z) = T_s - \Gamma z \qquad (4)$$

where Γ is the adiabatic lapse rate (7.7x10^{-3} K/m). The atmospheric temperature profile (and thus T_s) is determined by finding a tropopause where convective and radiative temperature and temperature gradients both match. Figure 1 shows the atmospheric temperature profiles resulting from 30 ppmv, 300 ppmv, and 3000 ppmv of H_2O in the atmosphere. The tropopause occurs where the lower atmosphere profile departs from the linear adiabat.

The opacity is altered by changes in the amount of H_2O in the atmosphere through delivery by magmatism and loss by exospheric escape. The total amount of water in the atmosphere, $N(t)$, at time t is given by

$$N(t) = \{N(0) - \tau_{ex} Q_{H_2O}(t)\}\exp(-t/\tau_{ex}) + \tau_{ex} Q_{H_2O}(t) \qquad (5)$$

where τ_{ex} is the exospheric escape time for hydrogen, taken as 160 Myr [*Grinspoon*, 1993]. The partial pressure of water in equation (3) is directly proportional to $N(t)$.

3. Results for Thin-lid Convection

The specifics of the thermal model follow *Phillips and Hansen* [1998] for Venus, and for a thin lid

$$\mathrm{Nu} = a\mathrm{Ra}^{1/3} \qquad (6)$$

where a is a constant. Heat producing isotopes of potassium (K), uranium (U) and Thorium (Th) drive the mantle convection system. The present-day value of uranium concentration is 25.7 ppb, with Th/U = 3.8, and K/U = 2x10^4 [e.g., *Barsukov et al.*, 1992]. The crust is 25 km thick with a fractionation efficiency of 10.

Figure 2 shows the surface temperature, T_s, over a 2 Gyr interval of venusian history for coupled and uncoupled models. A "coupled" model allows feedback between atmospheric opacity changes and partial melting by permitting changes in the surface temperature to affect the thermal model, and hence partial melting. An "uncoupled" model allows water delivery to the atmosphere to increase the greenhouse surface temperature, but this temperature is not transmitted to the thermal model. Contrasting these models isolates the effects of coupling. Both models commence at 4.6 Ga with T_s = 740 K and 30 ppmv H_2O in the atmosphere; the model runs end at 1 Ga. Partial melting and exospheric escape are initiated at 3 Ga, and a rapid increase in T_s results from the addition of new water to the atmosphere. If the heat

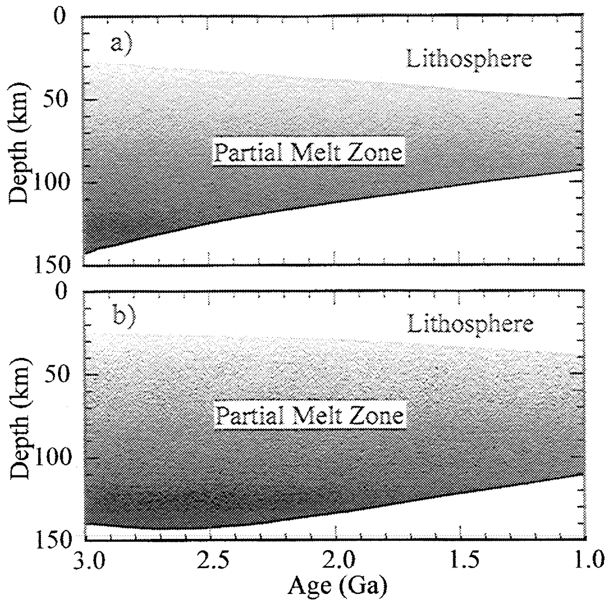

Figure 3. Partial melt zones in the interior corresponding to the surface temperature models of Figure 2. Melting extends from near the base of the lithosphere to the depth of zero melt and defines the melting channel. Lithospheric thickness (from top of melting zone to surface is 90% of thickness) also evolves with time.

production in the mantle were constant, then the uncoupled model would reach (in a time controlled by $\sim\tau_{ex}$) an equilibrium surface temperature with a constant amount of atmospheric water ($\tau_{ex}Q_{H_2O}$) set by delivery from magma and loss by exospheric escape. However, loss of parent isotopes in K, Th, and U cools the planet over time so that there is less partial melting and less water delivered to the atmosphere. Thus, after reaching quasi-equilibrium, the surface temperature declines in a steady fashion. The feedback between atmospheric opacity and partial melting is clearly evident in the coupled model. About 200 Myr into the model run, the surface temperature departs from the uncoupled model and reaches a maximum of 1010 K at 2.25 Ga. The surface temperatures in the two models slowly diverge with time; the difference at the end of the model run (1 Ga) is 60 K.

Figure 3 shows the evolution of partial melt column and lithospheric thickness. Changes in surface temperature will diffuse into the interior and affect the zone of partial melting

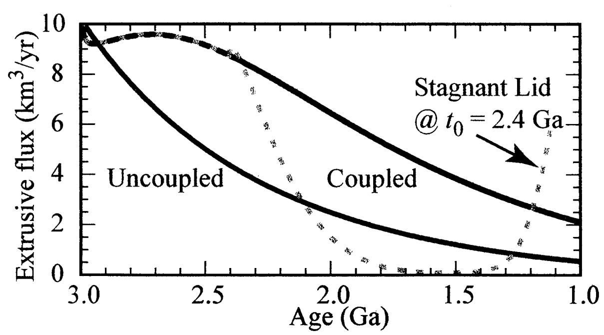

Figure 4. Extrusive melt fluxes corresponding to surface temperature models of Figure 2, plus a coupled model that commences a transition to stagnant lid convection at t_0 = 2.4 Ga.

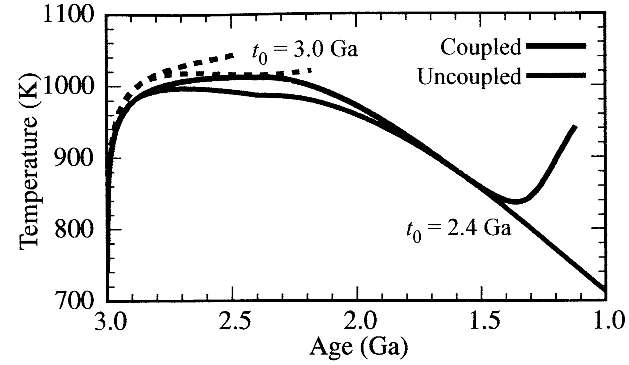

Figure 5. Surface temperature histories for coupled and uncoupled stagnant-lid models with conversion from thin to stagnant lid commencing at 3.0 and 2.4 Ga. Truncation of curves indicates an instability due to runaway partial melting.

with a time lag of several tens of millions of years. As this is much less than the time constant for exospheric escape of hydrogen (160 Myr), the feedback loop between climate change and partial melting should operate. In the uncoupled model (Figure 3a), the partial melt column thins and the lithosphere thickens monotonically with time. In the coupled model (Figure 3b), the lithosphere thickens more slowly, while the melt column thickens to a maximum at about 2.65 Ga, and then thins. At 1 Ga, melt column thickness is 73 and 44 km for the coupled and uncoupled models, respectively.

Figure 4 shows the extrusive volcanic rate for the uncoupled and coupled models. Both models start with a rate of around 10 km^3/yr. The uncoupled model decreases monotonically to 0.5 km^3/yr at the end of the model run. The coupled model has a local maximum at 2.7 Ga and at the end of the model run has four times the extrusive flux as the uncoupled model, attesting to the sensitivity of partial melting. to only moderate changes in potential temperature.

4. Conversion to Stagnant Lid Convection

It has been proposed that prior to ~1 Ga, Venus might have switched from thin-lid to stagnant-lid convection as convectively-induced stresses dropped below the yield stress of the lithosphere (with declining mantle heat production) and lithospheric recycling stopped [*Solomatov and Moresi*, 1996; *Phillips and Hansen*, 1998]. We assume a steady-state stagnant lid condition [*Moresi and Solomatov*, 1995; *Solomatov and Moresi*, 1996]

$$\mathrm{Nu} = a'\theta^{-1.02}\mathrm{Ra}^{0.2} \qquad (7)$$

where a′ is a constant and θ is defined by $\Delta\eta = \exp(\theta)$, where $\Delta\eta$ is the viscosity contrast across the convecting system (the interior convecting viscosity is used in the Rayleigh number). We illustrate the initiation of a stagnant lid at times t_0 = 3.0 and 2.4 Ga, and the conversion time to the thicker stagnant lid is controlled by the thermal diffusion thickening time of the lithosphere.

Initiation of a stagnant lid (Figure 5) leads to a thickening of the lithosphere and an increase in mantle temperature (due to less efficient heat transfer), opposing effects for the state of partial melting. At t_0 = 3.0 Ga, the increasing mantle temperature raises the melting rate to the point that an instability develops when the mantle temperature does not pass below the solidus with increasing depth. In actuality, such a melting runaway would be inhibited by a number of factors, including the effects of latent heat, extraction of heat-producing elements, and a change in the style of heat transfer. The coupled and uncoupled cases are similar. With t_0 = 2.4 Ga, surface temperature difference for the two cases declines as partial melting decreases due to thickening lithosphere, which in turn attenuates the feedback effect. For the uncoupled case, the melting goes to zero, and the absence of water flux into the atmosphere drives the surface temperature towards the no-water value of 522 K. For the coupled case, rapid lithospheric thickening drives the melt flux to nearly zero by 1.7 Ga (Figure 4). Increasing mantle temperature then increases the melting rate, and an instability develops.

5. Discussion and Conclusions

These very simple models demonstrate a complex interplay between a planet's convective evolution, partial melting state, and atmospheric radiative physics. To bring these models closer to reality, we should consider the reservoir exchange of greenhouse gas volatiles (e.g., H_2O, SO_2, CO_2), and their exogenic gain and loss. We should also evaluate *a*) the role of ^{40}Ar and ^{4}He in constraining the outgassing history [*Namiki and Solomon*, 1998; *Kaula*, 1999], *b*) volatile effects on partial melting relationships, *c*) volatile diffusion times through the lithosphere, *d*) crustal sequestration of heat sources, *e*) the finite lithospheric diffusion time of surface temperature changes, and *f*) the style and energetics of magmatism.

CO_2 delivery to the atmosphere by interior processes can also be modeled. Additionally, gas-solid and solid-solid mineral reactions at the surface (involving CO_2, CO, COS, etc.) can be driven either way by variations in the greenhouse-modulated surface temperature brought about by changes in abundances of the trace constituents [*Bullock and Grinspoon*, 2001]. Further, the simple gray atmosphere model used here can be replaced with one providing outgassing of both H_2O and SO_2 and employing high-resolution spectral absorption calculations and the formation and dissipation of sulfuric acid clouds [*Bullock and Grinspoon*, 2001]. Application to the real evolution of Venus will require consideration of all of these processes.

Acknowledgments. This work was supported by NASA Grant NAG5-4448 to Washington University and NASA Grant NAG5-8723 to Southwest Research Institute. Comments by three anonymous reviewers were very useful.

References

Anderson, F. S, and S. E. Smrekar, Tectonic effects of climate change on Venus, *J. Geophys. Res., 104*, 30,743-30,756, 1999.

Barsukov, V. L., Venus igneous rocks. In *Venus Geology, Geochemistry and Geophysics: Research Results from the Soviet Union*, eds., V. L. Barsukov, A. T. Basilevsky, V. P. Volkov and V. N. Zharkov, University of Arizona Press, Tucson, 165-176, 1992.

Bullock, M. A., and D. H. Grinspoon, The stability of climate on Venus, *J. Geophys. Res., 101*, 7521-7529, 1996.

Bullock, M. A., and D. H. Grinspoon, The recent evolution of climate on Venus, *Icarus*, in press, 2001.

Crisp, J. A., Rates of magma emplacement and volcanic output, *J. Volcan. Geotherm. Res., 20*, 177-211, 1984.

de Bergh, C., B. Bezard, D. Crisp, J. P. Maillard, T. Owen, J. Pollack and D. H. Grinspoon, Water in the deep atmosphere of Venus from high resolution spectra of the night side, *Advances in Space Research, 15*, (4)79-(4)88, 1995.

Grinspoon, D. H., Implications of the high D/H ratio for the source of water in Venus' atmosphere, *Nature, 363*, 428-431, 1993.

Hirschmann, M. M., P. D. Asimow, M. S. Ghiorso, and E. M. Stolper, Calculation of peridotite partial melting from thermodynamic model of minerals and melts, III, Controls on isobaric melt production and the effect of water on melt production, *J. Petrology, 40*, 831-851, 1999.

Kaula W. M., Constraints on Venus evolution from radiogenic argon, *Icarus, 139*, 32-39, 1999.

McGovern, P. J., and G. Schubert, Thermal evolution of the Earth: Effects of volatile exchange between atmosphere and interior, *Earth Plan. Sci. Lett., 96*, 27-37, 1989.

McKenzie, D., The generation and compaction of partially molten rock, *J. Petrol., 25*, 713-765, 1984.

McKenzie, D., and M. J. Bickle, The volume and composition of melt generated by extension of the lithosphere, *J. Petrol., 29*, 625-679, 1988.

Meadows, V. S., and D. Crisp, Ground-based near-infrared observations of the Venus night side: The thermal structure and water near the surface, *J. Geophys. Res., 101*, 4595-4622, 1996.

Moresi, L.-N., and V. S. Solomatov, Numerical investigation of 2D convection with extremely high viscosity variations, *Phys. Fluids, 7*, 2154-2162, 1995.

Namiki, N., and S. C. Solomon, Volcanic degassing of argon and helium and the history of crustal production on Venus, *J. Geophys. Res, 103*, 3655-3677, 1998.

Oyama, V. I., G. C. Carle, F. Woeller, J. B. Pollack, R. T. Reynolds, and R. A. Craig, Pioneer Venus Gas Chromatography of the Lower Atmosphere of Venus, *J. Geophys. Res., 85*, 7891-7902, 1980.

Phillips, R. J., and V. L. Hansen, Geological evolution of Venus: Rises, plains, plumes, and plateaus, *Science, 279*, 1492-1497, 1998.

Pollack, J. B., O. B. Toon, and R. Boese, Greenhouse models of Venus' high surface temperature, as constrained by Pioneer Venus measurements, *J. Geophys. Res., 85*, 8223-8231, 1980.

Pollack, J. B., J. B. Dalton, D. H. Grinspoon, R. B. Wattson, R. Freedman, D. Crisp, D. A. Allen, B. Bezard, C. de Bergh, L. P. Giver, Q. Ma, and R. H. Tipping, Near infrared light from Venus' nightside: A spectroscopic analysis, *Icarus, 103*, 1-42-1993.

Reese, C. C., V. S. Solomatov, and L.-N. Moresi, Non-Newtonian stagnant lid convection and magmatic resurfacing on Venus, *Icarus*, 139, 67-80, 1999.

Solomatov, V. S., and L.-N. Moresi, Stagnant lid convection on Venus, *J. Geophys. Res., 101*, 4737-4753, 1996.

Solomon, S. C., M. A. Bullock, and D. H. Grinspoon, Climate change as a regulator of tectonics on Venus, *Science, 286*, 87-90, 1999.

Stevenson, D. J., T. Spohn, and G. Schubert, Magnetism and thermal evolution of the terrestrial planets, *Icarus, 54*, 466-489, 1983.

R. J. Phillips and S. A. Hauck, II, Department of Earth and Planetary Sciences, Washington University, Campus Box 1169, St. Louis, MO 63130
(e-mail: phillips@wustite.wustl.edu)
M. A. Bullock, Department of Department of Space Studies, Southwest Research Institute, Boulder, CO

(Received May 25, 2000; revised January 30, 2001; accepted February 6, 2001.)