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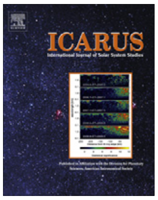

Coupling the atmosphere with interior dynamics: Implications for the resurfacing of Venus

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ABSTRACT

We calculated 2D and 3D mantle convection models for Venus using digitized atmosphere temperatures from the model of Bullock and Grinspoon (Bullock, M.A., Grinspoon, D.H. [2001]. Icarus 150, 19–37) to study the interaction between interior dynamics and atmosphere thermal evolution. The coupling between atmosphere and interior occurs through mantle degassing and the effect of varying concentrations of the greenhouse gas H_2O on the surface temperature. Exospheric loss of hydrogen to space is accounted for as a H_2O sink. The surface temperature enters the mantle convection model as a boundary condition.

Our results suggest a self-consistent feedback mechanism between the interior and the atmosphere resulting in spatial–temporal surface renewal. Greenhouse warming of the atmosphere results in an increase in the surface temperature. Whenever the surface temperature reaches a critical value, the viscosity difference across the lithosphere becomes smaller than about 10^5 and the surface becomes locally mobile. The critical surface temperature depends on the activation energy for mantle creep, the stress exponent in the non-Newtonian mantle rheology law, and the mantle temperature. Surface renewal together with surface lava flow may explain why the surface of Venus is young on average, i.e. not older than a few hundred million years.

The mobilization of the near-surface lithosphere increases the rate of heat removal from the mantle and thereby the interior cooling rate. The enhanced cooling results in a reduction of the water outgassing rates. As a consequence of decreasing water concentrations in the atmosphere, the surface temperature decreases. Our model calculations suggest that Venus should have been geologically active until recently. This is in agreement with several lines of observational evidence from thermal emissivity measurements and crater distribution analyses.

1. Introduction

The thermal evolution of a terrestrial planet involves heat transfer from the interior by convection (Schubert et al., 2002; Turcotte and Schubert, 2002), volcanism, and chemical differentiation. On one-plate planets like Mars, convection takes place below a stagnant upper layer and most of the heat is transported via conduction through this layer. On Earth, convection incorporates the surface layers through plate tectonics which results in the interior being cooled more efficiently than in the case of stagnant lid convection (e.g. Schubert et al., 2002). Plate tectonics can be regarded as a specific form of mobile lid convection where the mobile lid being mechanically stiff moves across the surface of the mantle as (several, distinct) plates. The climatic evolution of a planet is mainly controlled by the solar flux and the amount of greenhouse gases in the atmosphere. The latter may change significantly over time as greenhouse gases are released into the atmosphere by volcanic outgassing, leading to an increase in surface temperature. High surface temperatures may reduce the plate-like behaviour of the lithosphere – by making it more ductile – as was suggested for Venus by Lenardic et al. (2008), which in turn influences the climate since plate tectonics can recycle atmospheric CO_2 into the mantle. Higher surface temperatures may also result in an increase in mantle temperature and a corresponding increase in partial-melting and outgassing rates.

Phillips et al. (2001) investigated the coupling effect of the surface temperature on mantle dynamics for Venus by using simple parameterized convection models. They considered the outgassing of the greenhouse gas H_2O using a gray atmosphere model (Wildt, 1966; Sagan, 1969). Their model revealed a positive feedback mechanism: an increase in surface temperature leads to an increase in partial melting and hence an increase in atmospheric density and surface temperature. This triggers a positive runaway effect, which destabilizes the climate of the planet. The mechanism is illustrated in Fig. 1.

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Parameterized models are limited by their inability to adapt the convective regime (e.g. stagnant lid, mobile lid, plate tectonics) to changing boundary conditions such as the surface temperature, and because there is no lateral variation e.g. in lid thickness or convective stress. Whereas average lid thickness is self-consistently calculated during a parameterized thermal evolution calculation as long as the surface is in a stagnant-lid regime, transitions from a stagnant-lid to a locally mobile regime cannot be simulated. This limitation does not exist in 2D or 3D interior models which simulate fully convective mantle behaviour, so that the feedback mechanism might be different than illustrated in Fig. 1.

For the work presented in this paper we use the spherical convection simulation code GAIA (Hüttig and Stemmer, 2008) to which we added an atmosphere module that uses the digitized average surface temperature from Bullock and Grinspoon (2001) (their Fig. 6) as a function of the water vapour concentration in the atmosphere. We modify these surface temperatures for varying solar luminosities following Gough (1981). It is widely accepted that the solar luminosity has increased by about 40% in the past 4.5 Gyr. The concentration of water in the atmosphere is calculated from the mantle outgassing rate while employing observed exospheric loss rates. The influence of mantle outgassing on the atmosphere as well as the feedback effect on the interior through the surface temperature can then be calculated self-consistently.

Venus is a particularly good candidate to which to apply such a model. Venus today has a surface temperature of about 740 K, which is too high for liquid water to exist on the surface. The temperature is caused mainly by the greenhouse gases CO_2 and H_2O (Lewis, 2004; Grinspoon, 1993; Bullock and Grinspoon, 2001) of which H_2O is particularly effective with relatively small variations in concentrations having significant effects on the temperature as demonstrated by the models of Pollack (1969), Pollack et al. (1980) and Bullock and Grinspoon (2001).

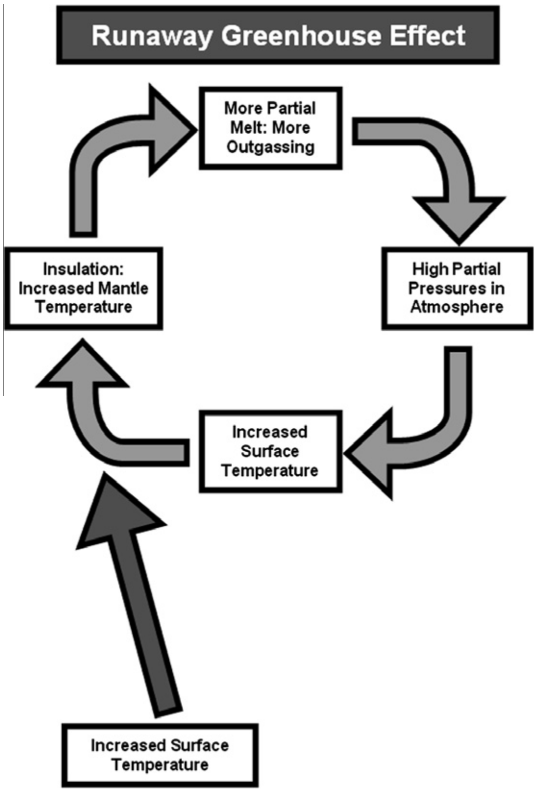

Fig. 1. Sketch illustrating the runaway greenhouse effect on a planet like Venus according to Phillips et al. (2001). Increases in surface temperature (e.g. due to outgassing) reduce mantle heat flux and increase mantle temperatures and melting rates, thus further increasing the surface temperature.

Another singular characteristic of Venus is the age of its surface, which is 300 Myr to 1 Gyr on average (McKinnon et al., 1997; Schaber et al., 1992; Strom et al., 1994). Romeo and Turcotte (2010) investigated the number and distribution of impact craters as well as the proportion of modified craters, concluding that these might indicate a catastrophic resurfacing event. In their model they assumed a gradual decrease in volcanic activity, so that about 40% of the surface would have been resurfaced in the last 750 Myr. Other studies (Schaber et al., 1992; Strom et al., 1994) identify a catastrophic resurfacing event, followed by much-reduced resurfacing activity.

For such a (possibly episodic) catastrophic global resurfacing event, several geodynamical explanations have been proposed so far. The event could have been exclusively magmatic (Reese et al., 1999), or caused by lithosphere thickening leading to episodic subduction and mantle overturn (Schubert et al., 1997; Fowler and O'Brian, 1996; Turcotte et al., 1999; Turcotte, 1993), or by a depleted layer on top of the mantle sinking down due to negative thermal buoyancy (Parmentier and Hess, 1992). Another explanation considers phase transitions in the mantle, which might lead to prolonged layered convection. At some point, however, a transition from layered to whole-mantle convection might occur, giving rise to a catastrophic resurfacing event (Steinbach et al., 1993). Further potential explanations include a cessation of plate tectonics 500 Myr ago (Schubert et al., 1997, 2002) or even episodic brittle mobilization due to a higher friction coefficient compared to Earth (Moresi and Solomatov, 1998; Stein et al., 2010). Basilevsky and Head (1998) explain geological features at the surface by compression and tension caused by variations in surface temperature over several hundred Myr but do not entirely rule out plate tectonics to be the cause of these features at an earlier stage of evolution.

Even on Earth, where plate tectonics is active, a large proportion of the surface – the continental crust – is several billion years old. We know from Magellan data that on Venus, too, some surface features (tessera terrains (Ivanov and Head, 1996)) seem to be much older than the mean values mentioned above (e.g. Hansen and López, 2010) and may even be comparable in age to Earth's continental crust. These investigations support another resurfacing history in which small patches are resurfaced at different times (Phillips et al., 1992; Phillips and Hansen, 1998; Guest and Stofan, 1999); features like coronae and plains basins seem to be much younger than the crustal plateaus. Consequently, catastrophic resurfacing seems less likely but cannot be ruled out. However, as these areas are quite small it is not possible to estimate their age reasonably by the crater counting method (Hauck et al., 1998; Hansen, 2000).

Smrekar et al. (2010) interpreted VIRTIS emissivity data as indicating recent volcanism on Venus and argued for gradual resurfacing. Recent volcanism in the last tens of Myr is also needed to explain the high concentration of SO_2 in the atmosphere (Bullock and Grinspoon, 2001). Stofan et al. (2005) conclude in their study that two thirds of the venusian surface is covered by volcanic material (e.g. coronae or large volcanoes) while one third has no identifiable source. These plains are interpreted as having a volcanic source but might as well be of different (unknown) origin.

Today, a highly sluggish or stagnant-lid regime is thought to prevail on Venus (Schubert et al., 1997; Nimmo and McKenzie, 1998). When surface temperatures were much higher in the past, Venus might have had a mobile surface, however. Reese et al. (1999) estimated that surface temperatures must exceed a critical value of about 1000 K to allow a transformation from a stagnant-lid to a global mobile regime. Higher surface temperatures than today – in the range of 850–1000 K – have been suggested by Phillips and Hansen (1998) and Ruiz (2007) to explain the formation of