AMES GRANT
IN-91-CR
190164
33P

A Heat-pipe Mechanism for Volcanism and Tectonics on Venus

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(NASA-CR-184791) A HEAT-PIPE MECHANISM FOR VOLCANISM AND TECTONICS ON VENUS (Cornell Univ.) 33 p CSCL 03B

N89-19224

Unclas
G3/91 0190164

Abstract

A heat-pipe mechanism is proposed for the transport of heat through the lithosphere on Venus. This mechanism allows the crust and lithosphere on Venus to be greater than 150 km thick. A thick crust and thick lithosphere can explain the high observed topography and large associated gravity anomalies. For a 150 km thick lithosphere, the required volcanic flux on Venus is 200 km^3/yr; this is compared with a flux of 17 km^3/yr associated with the formation of the oceanic crust on the earth. A thick basaltic crust on Venus is expected to transform to eclogite at a depth of 60 to 80 km; the dense eclogite would contribute to lithospheric delamination that returns the crust to the interior of the planet completing the heat-pipe cycle. Topography and the associated gravity anomalies can be explained by Airy compensation of the thick crust. The principal observation that is contrary to this hypothesis is the mean age of the surface that is inferred from crater statistics; the minimum mean age is about 130 Myr and this implies an upper limit of 2 km^3/yr for the surface volcanic flux. If the heat-pipe mechanism was applicable on the earth in the Archean it would provide the thick lithosphere implied by isotopic data from diamonds.

Introduction

Venus, in terms of size, is the planet in the solar system most similar to the earth. However, studies carried out to date show major geological differences. The elongated trenches and ridge structures associated with plate tectonics on the earth are absent on Venus. Linear volcanic trends associated with mantle plumes are also absent on Venus. Several general explanations can be given for these differences:

i.) The surface of Venus is about 450 ^{o}K hotter than the surface of the earth. The interior temperature in Venus may be somewhat higher than in the earth, but the strong exponential dependence of the solid-state viscosity on temperature would require the difference to be considerably less than the difference in surface temperature. Thus, it is concluded that the temperature difference across the lithosphere on Venus is about two-thirds of the difference on the earth. For a similar conductive heat flux through the lithosphere, its average thickness would also be reduced by about a third. The higher surface temperature on Venus implies a weaker lithosphere and the reduced negative bouyancy of the lithosphere could impede plate tectonics on the planet.

ii). The lack of oceans eliminates the hydrologic cycle. Without extensive erosion, major tectonic features would be expected to last for long periods of time. The lack of erosion and hydrologic transport could also preclude the formation of thick silicic continents. Certainly the bimodal hypsometric curve on earth is provided by continental erosion to sea level.

A large fraction of the geological activity on earth can be directly associated with plate tectonics. Some authors associate topographic features on Venus with plate tectonics [Brass and Harrison, 1982; Head and Crumpler, 1987; Crumpler and Head, 1988]. But most authors argue that plate tectonics does not occur or is so inefficient that it only transfers a small fraction of the heat from the interior to the surface [Kaula, 1984; Kaula and Phillips, 1981; Grimm and Solomon, 1987]. Certainly neither a global set of ridges (spreading centers) or trenches (subduction zones) are observable from either the altimeter or radar observations of the surface.

One of the major surprises from the altimeter data [Pettingill et al., 1980] was the presence of several extensive regions of high topography. Aphrodite Terra and Beta Regio have elevations in the range 4-5 kms over the mean reference surface and Ishtar Terra has elevations as high as 12 km. These elevations are comparable to the highest elevations on the earth.

Another surprise was that the major topographic features on Venus could be directly correlated with large gravity anomalies [Sjogren et al., 1983; Mottinger et al., 1985]. This is generally not the case on the earth where major mountain belts such as the Himalayas, Andes, or Rockies have small gravity anomalies. The gravity anomalies on Venus are small compared with those expected from uncompensated topography, but require depths of compensation of 135 km for Beta Regio [Esposito et al., 1983] and up to 150 km for Ishtar [Sjogren et al., 1984]. Reasenberg et al. [1982] suggest at least two compensation mechanisms are required.

On the earth, mantle convection is responsible for plate tectonics. The plates are the cold thermal boundary layers of mantle convection cells. Without plate tectonics, alternative means for the transport of heat to the surface must be found. This problem has been considered in some detail by Solomon and Head [1982], Phillips and Malin [1983, 1984], and Morgan and Phillips [1983]. Two alternative mechanisms have been proposed:

i). Conductive transport through a relatively thin, rigid lithosphere. Solid-state convection within the interior of Venus would transport heat to the base of the lithospheric cap. The primary objection to this hypothesis is the presence of the large correlated topography and gravity anomalies on Venus. It is extremely difficult to hypothesize a mechanism that can explain these anomalies with a thin lithosphere.

ii). Surface volcanic flows. Pressure release melting associated with solid-state convection at depth would produce magmas that would rapidly ascend to the surface in magma driven fractures producing thin extrusive flows. The solidification and cooling of these flows acts as a heat-pipe mechanism for the transport of heat and can lead to a relatively thick lithosphere. This mechanism has been proposed for the cooling of Io by O'Reilly and Davies [1981]. Previous authors have questioned the applicability of this mechanism for Venus due to the very large flux of magma required.

The purpose of this paper is to examine the heat-pipe mechanism for Venus in some detail and to discuss its implications. It will also be argued that this mechanism may have been important for the earth, particularly in the Archean.

The Heat-Pipe Hypothesis

Our basic hypothesis is that a substantial fraction of the heat lost from the interior of Venus is transported through the lithosphere by magma. Magma is generated within the interior of Venus in the ascending limbs of mantle convection cells. This magma pools at the base of the lithosphere and then migrates to the surface in magma-generated fractures. The surface eruptions of the magma solidify rapidly without significantly heating the surface. Some heat is lost during transit through the lithosphere and crust, this loss is estimated in a later section. This is a heat-pipe mechanism. The melting at depth and surface solidification is an efficient means of transporting heat.

The standard plate tectonic mechanism for heat transport is illustrated in Figure 1. The oceanic lithosphere is the cold thermal boundary layer of mantle convection cells. Conductive heat loss from the surface of the plate is the primary means of heat transport from the interior of the earth at the present time. An essential feature of this process is the subduction of the oceanic crust and its subsequent mixing into the earth's mantle. This prevents complete fractionation of the earth and returns substantial quantities of the heat-producing elements to the interior.

The heat pipe mechanism is illustrated in Figure 2. The various stages of volcanic addition and loss of crust by delamination are illustrated as a sequence in time, horizontal surface displacements are not implied. Surface volcanic flows accumulate and thicken the crust. The crust is sufficiently thick so that its base transforms to eclogite. Based on the experimental work of Green and Ringwood [1972] this phase transformation is likely to occur at pressures of 1.5 GPa to 2.0 GPa; on Venus this corresponds to depths of 60 to 80 km. The dense eclogite phase is gravitationally unstable with respect to the mantle below; when the mantle lithosphere and eclogite layer are sufficiently thick delamination can occur. There is a reasonable basis for hypothesizing that delamination occurs at the basalt-eclogite phase change boundary as this is likely to be a zone of weakness.

Delamination on the earth has been suggested by Bird [1978, 1979] and Bird and Baumgardner [1981]. These authors model delamination in a manner similar to subduction, the lithosphere bends and sinks into the mantle. Large gravity and topographic anomalies would be associated with such a process, these anomalies are not observed on the earth. Turcotte (1983) has proposed an alternative mechanism for delamination, lithospheric stoping in which the lithosphere fails along pre-existing zones of weakness and blocks of lithosphere break away and sink into the mantle. McKenzie and O'Nions [1983] have suggested that delamination occurs at island arcs. Delamination has been proposed by Head [1986] as a possible explanation for the elevated topography of Ishtar Terra on Venus.

As illustrated in Figure 2 basaltic volcanism is hypothesized to be continuous; however, the heat loss from the ascending magma to the lithosphere is sufficiently low that the lithosphere continues to thicken. The extrusive volcanics thickens the crust and leads to subsidence; this subsidence causes the basalt in the lower crust to cross into the eclogite stability field. Since the eclogite is more dense than the mantle rock it is not only gravitationally unstable, but can also result in net subsidence of the surface. When the negative buoyancy of the eclogitic crust and mantle lithosphere become sufficiently large, they delaminate at the basalt-eclogite phase boundary and sink into the mantle; the result is a period of rapid uplift. The regions of high topography on Venus are associated with this uplift. The cycle then repeats.

Volcanism

The heat-pipe mechanism certainly can be an effective means of heat transport from a planetary interior. However, an essential question is whether the required volume flux of surface volcanics is compatible with observations. Before this volume flux can be determined the mean heat flow to the surface of Venus must be specified. The mean heat flow to the surface of the earth is $\bar{q} = 72$ mW/m^2 and is quite well constrained. However, the fraction of this heat loss that is attributed to radiogenic isotopes (the Urey number U) can only be estimated with values ranging from 0.5-0.85; the remainder is due to secular cooling. A reasonable hypothesis is to simply scale the surface heat flux from the earth to Venus so that $\bar{q} = 66$ mW/m^2 for Venus.

If this heat flowed through a lithosphere that acted as a rigid-conductive lid, the thickness of the lid would be

$$y_L = \frac{k\Delta T}{\bar{q}} \tag{1}$$

It is reasonable to estimate that the transition from rigid to fluid behavior in the mantle of Venus occurs at a temperature of 1560°k. With a mean surface temperature of 735°k we find $\Delta T = 825$°k. With $k = 3.3$ W/m°k we find that the mean thickness of the lithosphere on Venus would be $y_L = 41.25$ km.

Based on our analogy, the total heat loss from the interior of Venus is $Q = 3.03 \times 10^{13}$ W. The required volcanic flux $\dot{V}$ if the heat-pipe mechanism transported all this heat is

$$\dot{V} = \frac{Q}{\rho(L + c_p \Delta T)} \tag{2}$$

where L is the latent heat of fusion of basalt (4×10^5 J/kg), c_p is the specific heat at constant pressure (1 kJ/kg ^oK), and $\Delta T = 825\ ^oK$. The value of the required volcanic flux is $\dot{V} = 270\ km^3/yr$. Clearly a fraction of the surface heat loss on Venus can be attributed to conduction through a lithosphere. Combining (1) and (2) gives

$$\dot{V} = \frac{Q - \dfrac{4\pi a^2 k \Delta T}{y_L}}{\rho\ (L + c_p \Delta T)} \tag{3}$$

where a is the radius of Venus. The dependence of the volcanic flux on the mean thickness of the lithosphere on Venus is given in Figure 3.

Based on the discussion in the previous section the minimum lithospheric thickness for delamination to occur at the basalt-eclogite phase boundary is about 100 km. Our determination of the maximum thickness in the next section is 356 km. As a reasonable value we take $y_L = 150$ km. From Figure 3, the required volcanic flux is $\dot{V} = 200\ km^3/yr$. Twenty-five percent of the heat flux from the interior is by conduction and seventy-five percent is by the heat-pipe mechanism.

It is of interest to compare the required volcanic flux for Venus with various volcanic fluxes on the earth. The largest source of volcanism on the earth is at mid-ocean ridges. This volcanic flux is about 17 km^3/yr, or about an order of magnitude less than the required flux for Venus. A comparison with intraplate fluxes may be more directly relevant. The long-term volcanic flux for the Hawaiian hot spot is about 0.02 km^3/yr so that Venus would require 1,000 Hawaiian hot spots. However, there is evidence that hot spot volcanism on the earth has been considerably higher in the past. The volcanic flux associated with the formation of the Deccan Traps is 4.9 km^3/yr [McLean, 1985]. The volcanic flux associated with the formation of the Mid-Pacific Mountains in the Cretaceous is 3.8 km/yr [Arthur et al., 1985]. The flux required for Venus is certainly high compared with present rates of volcanism on the earth, but this is not a difficulty in itself.

In order to get a better feeling for the implications of $\dot{V} = 200$ km^3/yr for Venus, we make a few simple calculations. If volcanism uniformly covered the entire surface of Venus, the mean interval between eruptions τ_i is given by

$$\tau_i = \frac{4\pi a^2 h_f}{\dot{V}} \tag{4}$$

where h_f is the mean thickness of the erupted magma. For $\dot{V} = 200$ km^3/yr the dependence of τ_i on h_f is given in Figure 4. For example, an eruption 10 m thick that covers the entire surface of Venus is required every 23,000 years. Also given in Figure 4 is the time required to solidify the extrusive eruption τ_c. It is a reasonable approximation (Carslaw and Jaeger, 1959, pp. 283-289) to assume that

$$\tau_c = \frac{h_f^2}{4\kappa} \tag{5}$$

with $\kappa = 0.8$ mm^2/s. In all cases the cooling time is several orders of magnitude shorter than the interval time. This shows that the flow can be quite localized and solidification will still take place in the intervals between flows. If it is assumed that volcanism is restricted to the areas of high topography, surface flows would cover about 6% of the entire surface with an area of about 30 x 10^6 km^2. If these areas have a 10 km thickness of volcanic flows, they would be formed in only 1.5 Myr. Quite an impressive rate.

There is certainly extensive evidence for recent surface volcanism on Venus [Schaber, 1982; Esposito, 1984; Ksanfomaliti, 1985; Taylor and Cloutier, 1986; Head and Wilson, 1986]. However, the essential question is the volumetric flow.

An important constraint on the rate of surface volcanism is the density of impact craters. Crater counts have been obtained from Arecibo radar data [Campbell and Burns, 1980] and from Veneras 15 and 16 radar data [Barsukov et al., 1986]. Frequency-size statistics and morphology favor an impact origin for a majority of the observed craters. Using statistics from other planetary bodies, it is inferred that the mean age of the Venus surface where data are available is between 0.5 and 1 Gyr. Grimm and Solomon [1987] took a minimum mean surface age of 130 Myr and concluded that the upper limit on the rate of volcanism is V = 2 km^3/yr. For an older mean surface age the inferred rate of volcanism would be less. These observations could be compatible with the heat-pipe mechanism if:

i) Volcanism on Venus is very localized. Observational evidence does not favor this.

ii) A large fraction of the craters have a volcanic origin. It would not be surprising if both impact and volcanic craters obey scale-invariant, power-law statistics.

iii) A substantial fraction of the volcanism occurs as shallow intrusions.

On balance, however, the density of impact craters is evidence against the heat-pipe mechanism transporting a large fraction of the heat from the interior of Venus.

Magma Transport

The only way in which magma can be transported through a thick, cool lithosphere is by magma fracture. This process has been studied by Spence and Turcotte [1985], Emerman et al. [1988] and Spence et al. [1987]. It is appropriate to hypothesize that the magma ascends upwards due to its buoyancy in thin cracks of width w and horizontal length ℓ. For the applicable turbulent flow, the mean ascent velocity $\bar{u}$ is given by

$$\bar{u} = 4.7 \frac{w^{5/7} \quad (\Delta\rho \;\, g)^{4/7}}{\rho_m{}^{3/7} \quad \eta^{1/7}} \tag{6}$$

where ρ and η are the density and viscosity of the magma and $\Delta\rho$ is the density difference between the magma and the lithosphere through which it is ascending. The volume of magma in an eruption V is given by

$$V = \ell \; \bar{w} \; u \; \tau_f \tag{7}$$

where ℓ is the horizontal length of the crack and τ_f is the time of eruption.

As a typical example, we take $w = 1$ m, $\Delta\rho = 300$ kg/m^3, $\rho = 3{,}000$ kg/m^3, $\eta = 1$ Pa s and find from (6) that $\bar{u} = 13.5$ m/s. Taking $\ell = 100$ km and $\tau_f = 1$ day, we find from (7) that $V = 117$ km^3. This would be a flow 100 m thick covering an area 1,000 by 1,200 km. To provide the required mean flux an average of two of these eruptions would have to occur each year. The length and width of the assumed crack are typical of dikes found in shield areas of the earth. The flow rates are also consistent with observations in Hawaii and Iceland. An alternative to the single fast eruption is a series of slow eruptions through thinner cracks that form thick dikes by multiple eruptions.

As the magma ascends upwards through the lithosphere in magma driven fractures, the magma will lose heat to the lithosphere and some magma will solidify. Assuming that magma erupts at a temperature T_m and that the rock through which it flows has a temperature T(y), the heat lost from the magma in the crack to the country rock during an eruption is [Turcotte and Schubert, 1982; p. 161]

$$Q = 4\,k\,\ell\,(T_m - T)\left(\frac{\tau_f}{\pi\,\kappa}\right)^{\frac{1}{2}} \tag{8}$$

per unit depth where k and κ are the thermal conductivity and thermal diffusivity of the country rock respectively. The mean heat loss to the lithosphere of Venus per unit depth is given by

$$q = \frac{Q}{\tau_i} = \frac{4\,k\,\ell\,(T_m - T)}{\tau_i}\left(\frac{\tau_f}{\pi\,\kappa}\right)^{\frac{1}{2}} \tag{9}$$

This is the amount of heat added to the lithosphere from the ascending magma per unit depth.

We now make a one-dimensional approximation for the variation of T through the lithosphere and relate the heat conduction to the heat transferred from the magma using the equation

$$0 = k\,\frac{d^2T}{dy^2} + \frac{q}{4\pi q^2} \tag{10}$$

where a is the radius of Venus. Implicit in writing (10) is the assumption that adjacent magma fractures are closer than the thickness of the lithosphere ($\simeq$ 150 km) on a thermal time scale for the lithosphere ($\simeq$ 700 Myr). Substitution of (9) into (10) gives

$$\frac{d^2T}{dy^2} = - \frac{\ell}{\pi a^2 \tau_i} \left(\frac{\tau_f}{\pi \kappa} \right)^{1/2} (T_m - T) \tag{11}$$

With the boundary conditions $T = T_0$ at $y = 0$ and $T \rightarrow T_m$ as $y \rightarrow \infty$, (11) is integrated to give

$$\frac{T_m - T}{T_m - T_0} = \exp \left\{ - \left[\frac{\ell}{\pi a^2 \tau_i} \left(\frac{\tau_f}{\pi \kappa} \right)^{1/2} \right]^{1/2} y \right\} \tag{12}$$

Taking $(T_m - T)/(T_m - T_0) = 0.1$ at the base of the lithosphere, we find that the thickness of the lithosphere is given by

$$d_L = 2.30 \left[\frac{\pi a^2 \tau_i}{\ell} \left(\frac{\pi \kappa}{\tau_f} \right)^{1/2} \right]^{1/2} \tag{13}$$

Based on the examples given above, we take $\tau_i = 0.58$ yr, $\tau_f = 1$ day, $\ell =$ 100 km, a = 6050 km, and $\kappa = 10^{-6}$ m^2/s, from (13) we find $d_L = 356$ km. This is the maximum thickness of the lithosphere due to the loss of heat from magma ascent. The lithosphere can be thinner, either due to heat input to its base, or due to crustal delamination. The spacing condition for the applicability of the one-dimensional model is satisfied.

Topography and Gravity Anomalies

The tectonic processes responsible for topography and gravity anomalies on Venus must be associated with the processes of heat transport to the surface of the planet. Topography can be associated with three basic tectonic processes:

(1) Variations in the thickness of the crust. The elevation of the continents on the earth relative to the oceans is attributed to variations in crustal thickness. A large fraction of the topography on the earth can be directly related to this mechanism. The maximum relative elevation difference due to this effect on the earth is about 14 km. Anderson [1981], Bowin [1983], and Bowin et al., [1985] have suggested its applicability to Venus.

The topographic elevation w_c is related to increases in crustal thickness Δh_c by the isostatic relation

$$w_c = \Delta h_c \left[1 - \frac{\rho_c}{\rho_m} \right] \tag{14}$$

where ρ_c is the crustal density and ρ_m the mantle density. Taking ρ_c = 2,800 kg/m^3 and ρ_m = 3,300 kg/m^3, 12 km of topography on Venus requires an increase of crustal thickness Δh_c = 80 km. Such a thick crust would be incompatible with a thin (40 km) conductive lithosphere since variations in crustal thickness must be embedded in a rigid lithosphere in order to survive.

(2) Variations in the thickness of the lithosphere. This mechanism is responsible for the elevation of the mid-ocean ridge system relative to the ocean basins on the earth. It is also responsible for the elevation of oceanic and continental swells that are associated with intraplate hot spots. The maximum elevation associated with lithosphere thinning on the earth is about 2.5 km.

The topographic elevation w_L is related to the thickness of the lithosphere y_L by the isostatic relation

$$w_L = -\frac{1}{\rho_m}\int_o^{y_L}(\rho-\rho_m)\,dy = -\alpha\int_o^{y_L}(T_m-T)\,dy$$

$$= -\alpha\,(T_m-T_o)\int_o^{y_L}\frac{y\,dy}{y_L} = -\frac{1}{2}\,\alpha\,(T_m-T_o)\,y_L \qquad (15)$$

where a linear temperature is assumed in the lithosphere, α is the volumetric coefficient of thermal expansion, T_m the mantle temperature, and T_o the surface temperature. Taking $\alpha = 3 \times 10^{-5}$ °K^{-1} and $T_m - T_o =$ 825°K, the elevation associated with a conductive lithospheric thickness of 40 km is 0.5 km, an insignificant fraction of the observed topography.

Topography on Venus due to changes in crustal and lithospheric thicknesses are given in Figure 5. It is seen from this figure that variations in lithospheric thickness can only make relatively small contributions to topography, about one kilometer. Variations in crustal thickness can certainly provide the observed topography on Venus, but variations as large as 75 km are required to produce the observed 12 km in Ishtar Terra. Similar results have been obtained by Morgan and Phillips [1983].

An explanation for the large topographic variations on Venus should also be able to explain the correlated gravity anomalies. The largest and most striking topography associated gravity anomaly is that of Beta Regio. The maximum topography is about 5 km and the maximum gravity anomaly is about 135-150 mgal [Reasonberg et al., 1982; Esposito et al., 1983]; this represents about 75% compensation. The associated geoid anomaly for Beta Regio is about 80 m. The geoid anomaly ΔN associated with Airy compensation is given by [Turcotte and Schubert, 1982, p. 225].

$$\Delta N = \frac{\pi G}{g} \rho_c \left[2 h_{co} w + \frac{\rho_m w^2}{(\rho_m - \rho_c)} \right] \tag{16}$$

where G is the gravitational constant and h_{co} is the reference crustal thickness corresponding to the reference geoid. The geoid anomalies and crustal thicknesses are given as a function of the fully compensated crustal elevation in Figure 6, the corresponding thicknesses are also given. It is seen that for an initial crustal thickness of 100 km, 4 km of topography requires a total crustal thickness of 125 km. The corresponding geoid anomaly is 60 m. This result is in reasonably good agreement with the 135 km depth of compensation given by Reasonberg et al. [1982] and Esposito [1983]. Thus a relatively thick crust and lithosphere on Venus can explain the topography and associated gravity anomalies. The heat-pipe mechanism for heat transfer is one way to obtain a thick lithosphere.

An alternative hypothesis to explain the topography on Venus is dynamic support. This mechanism has been considered by Kiefer et al [1986]. It is certainly possible to generate topography through pressure gradients associated with mantle convection. However, the critical quantity is the mantle viscosity. Significant dynamic topography requires high pressures that are associated with high viscosities. Essentially no dynamic topography with associated gravity anomalies are observed on the earth. Essentially all topography on the earth can be directly associated with either changes in crustal thickness (the continents) or changes in lithospheric thickness (mid-ocean ridges and hot-spot swells) [Turcotte and Schubert, 1983, pp. 225-229]. An exception is aseismic ocean ridges, but these are associated with density variations within the mantle lithosphere [Angevine and Turcotte, 1980, 1983]. Significant dynamic topography on Venus implies a higher mantle viscosity and thus a lower mantle temperature. This is inconsistent with the higher surface temperature and the lack of plate tectonics.

Conclusions

The favored explanation for most of the large topographic and correlated gravity anomalies on Venus is Airy isostacy associated with a thickened crust. This explanation is incompatible with a thin conductive lithosphere on the planet.

One method to increase the thickness of a planetary lithosphere is to transport heat by volcanism. Melting at depth and solidification at or near the surface is an efficient means of heat transport known as a heat pipe. For a 150 km thick lithosphere on Venus the required volcanic flux is 200 km^3/yr; this is compared with a flux of 17 km^3/yr associated with the formation of the oceanic crust on the earth. The high elevations on Venus are attributed to this volcanism. The principal observation that is not consistent with massive surface volcanic flows is the mean age of the surface inferred from crater statistics. If these craters are attributed to a standard flux of meteorite bombardment the minimum mean age of the surface is about 130 Myr; this implies an upper limit of 2 km^3/yr for the surface volcanic flux.

In order to maintain volcanism it is necessary to recycle basalt into the interior of the planet. A thick basaltic crust on Venus is expected to transform to eclogite at a depth of 60 to 80 km; the dense eclogite would contribute to lithospheric delamination that would return crust to the interior of the planet.

The heat pipe mechanism has been proposed [O'Reilly and Davies, 1981] for Io in order to provide a thick lithosphere to support the observed topography on that body. Isotopic studies of diamonds by Richardson et al. [1984] have been interpreted to require a lithospheric thickness on earth of at least 175 km 3 Gyr before present. Since the heat output from radioactive isotopes at that time was twice the present value it is difficult to explain such a thick lithosphere, even with plate tectonics. A large heat flux due to extensive volcanism in the Archean which transported heat by the heat-pipe mechanism would help explain this observation.

ACKNOWLEDGEMENTS

The research has been supported by Grant NAGW 797 from the National Aeronautics and Space Administration. This is contribution 838 of the Department of Geological Sciences, Cornell University.

References

Anderson, D. L., Plate tectonics on Venus, *Geophys. Res. Let.*, *8*, 309-311, 1981.

Angevine, C.L. and D.L. Turcotte, On the Compensation mechanism of the Walvis Ridge, *Geophys. Res. Let. 7*, 477-479, 1980.

Angevine, C.L. and D.L. Turcotte, Correlation of geoid and depth anomalies over the Agulhus Plateau, *Tectonophys. 100*, 43-52, 1983.

Arthur, M. A., W. E. Dean, and S. L. Schlanger, Variations in the global carbon cycle during the Cretaceous related to climate, volcanism, and changes in atmospheric CO_2, in *The Carbon Cycle and Atmospheric* CO_2*: Natural Variations Archean to Present*, E. T. Sundquist and W. S. Broecker, eds., pp. 504-529, Geophys. Mono. 32, American Geophysical Union, Washington, D.C., 1985.

Barsukov, V. L., et al., The geology and geomorphology of the Venus surface as revealed by the radar images obtained by Veneras 15 and 16, *J. Geophys.*, *91*, D378-D398, 1986.

Bird, P., Initiation of intracontinental subduction in the Himalaya, *J. Geophys. Res.*, *83*, 4975-4987, 1978.

Bird, P., Continental delamination and the Colorado Plateau, *J. Geophys. Res.*, *84*, 7561-7571, 1979.

Bird, P., and J. Baumgardner, Steady propagation of delamination events, *J. Geophys. Res.*, *86*, 4891-4903, 1981.

Bowin, C., Gravity, topography, and crustal evolution of Venus, *Icarus*, *56*, 345-371, 1983.

Bowin, C., G. Abers, and L. Shure, Gravity field of Venus at constant altitude and comparison with Earth, *J. Geophys. Res.*, *90*, C757-C770, 1985.

Brass, G. W., and C. G. A. Harrison, On the possibility of plate tectonics on Venus, *Icarus*, *49*, 86-96, 1982.

Campbell, D.B., and B. A. Burns, Earth-based radar imagery of Venus, *J. Geophys. Res.*, *85*, 8271-8281, 1980.

Carslaw, H.S., and J.C. Jaeger, *Conduction of Heat in Solids*, 2nd ed., 510 pp., Oxford University Press, Oxford, 1959.

Crumpler, L. S., and J. W. Head, Bilateral topographic symmetry patterns across Aphrodite Terra, Venus, *J. Geophys. Res.*, *93*, 301-312, 1988.

Emerman, S. H., D. L. Turcotte, and D. A. Spence, Transport of magma and hydrothermal solutions by laminar and turbulent fluid fracture, *Phys. Earth Planet. Int.*, *41*, 249-259, 1986.

Esposito, L.W., Sulfur dioxide: Episodic injection shows evidence for active Venus volcanism, *Science*, *223*, 1072-1074, 1984.

Esposito, P. B., W. L. Sjogren, N. A. Mottinger, B. G. Bills, and E. Abbott, Venus gravity: Analysis of Beta Regio, *Icarus*, *51*, 448-459, 1983.

Green, D. H., and A. E. Ringwood, A comparison of recent experimental data on the Gabbro-garnet granulite-eclogite transition, *J. Geology*, *80*, 277-288, 1972.

Grimm, R. E., and S. C. Solomon. Limits on modes of lithospheric heat transport on Venus from impact crater density, *Geophys. Res. Let.*, *14*, 538-541, 1987.

Head, J. W., Ishtar Terra, Venus: A simple model of large-scale tectonic convergence, crustal thickening; and possible delamination, *Lunar Planet. Sci. XVII*, 323-324, 1986.

Head, J. W., and L. S. Crumpler, Evidence for divergent plate-boundary characterization and crustal spreading on Venus, *Science, 238*, 1380-1385, 1987.

Head, J. W., and L. Wilson, Volcanic processes and landforms on Venus: Theory, predictions, and observations, *J. Geophys. Res., 91*, 9407-9446, 1986.

Kaula, W. M., Tectonic contrasts between Venus and the earth, *Geophys. Res. Let., 11*, 35-37, 1984.

Kaula, W. M., and R. J. Phillips, Quantitative tests for plate tectonics on Venus, *Geophys. Res. Let., 8*, 1187-1190, 1981.

Kiefer, W. S., M. A. Richards, and B. H. Hager, A dynamic model of Venus' gravity field, *Geophys. Res. Let., 13*, 14-17, 1986.

Ksanfomaliti, I.V., Signs of possible volcanism on Venus, *Solar System Res., 18*, 200-207, 1985.

McKenzie, D., and R. K. O'Nions, Mantle reservoirs and ocean island basalts, *Nature, 301*, 229-231, 1983.

McLean, D. M., Mantle degassing induced death ocean in the Cretaceous-Tertiary transition, in *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present*, E. T. Sundquist and W. S. Broecker,eds., pp. 493-503, Geophys. Mono. 32, Am. Geophys. Un., Washington, DC,1985.

Morgan, P., and R. J. Phillips, Hot spot heat transfer: Its application to Venus and implications to Venus and Earth, *J. Geophys. Res., 88*, 8305-8317, 1983.

Mottinger, N. A., W. L. Sjogren, and B. G. Bills, Venus gravity: A harmonic analysis and geophysical implications, *J. Geophys. Res., 90,* C739-756, 1985.

O'Reilly, T. C., and G. F. Davies, Magma transport of heat on Io: A mechanism allowing a thick lithosphere, *Geophys. Res. Let., 8,* 313-316, 1981.

Pettingill, G. H., E. Eliason, P. G. Ford, G. B. Loriot, H. Masursky, and G. E. McGill, Pioneer Venus radar results: Altimetry and surface properties, *J. Geophys. Res., 85,* 8261-8270, 1980.

Phillips, R. J., and M. C. Malin, The interior of Venus and tectonic implications, in *Venus*, D. M. Hunten, L. Colin, T. M. Donahue, and V. I. Moroz, eds., pp. 159-214, University of Arizona Press, Tucson, 1983.

Phillips, R. J., and M. C. Malin, Tectonics of Venus, *An. Rev. Earth Planet. Sci., 12,* 411-443, 1984.

Reasenberg, R. D., Z. M. Goldberg, and I. I. Shapiro, Venus: Comparison of gravity and topography in the vicinity of Beta Regio, *Geophys. Res. Let., 9,* 637-640, 1982.

Richardson, S. H., J. J. Gurney, A. J. Erlank, and J. W. Harris, Origin of diamonds in old enriched mantle, *Nature, 310,* 198-202, 1984.

Schaber, G. G., Venus: Limited extension and volcanism along zones of lithospheric weakness, *Geophys. Res. Let., 9,* 499-502, 1982.

Sjogren, W. L., B. G. Bills, P. W. Birkeland, P. B. Esposito, A. R. Konopliv, N. A. Mottinger, S. J. Ritke, and R. J. Phillips, Venus gravity anomalies and their correlations with topography, *J. Geophys. Res., 88,* 1119-1128, 1983.

Sjogren, W. L., B. G. Bills, and N. A. Mottinger, Venus: Ishtar gravity anomaly, *Geophys. Res. Let.*, *11*, 489-491, 1984.

Solomon, S. C., and J. W. Head, Mechanisms for lithospheric heat transport on Venus: Implications for tectonic style and volcanism, *J. Geophys. Res.*, *87*, 9236-9246, 1982.

Spence, D. A., P. W. Sharp, and D. L. Turcotte, Buoyancy-driven crack propagation: A mechanism for magma migration, *J. Fluid Mech.*, *174*, 135-153, 1987.

Spence, D. A., and D. L. Turcotte, Magma-driven propagation of cracks, *J. Geophys. Res.*, *90*, 575-580, 1985.

Taylor, H.A. and P.A. Cloutier, Venus: Dead or Alive? *Science*, *234*, 1087-1093, 1986.

Turcotte, D. L., and G. Schubert, Geodynamics, John Wiley and Sons, New York, 1982.

Turcotte, D. L., Mechanisms of crustal deformation, *J. Geol. Soc. London*, *140*, 701-724, 1983.

Figure Captions

Figure 1. Illustration of the plate tectonic mechanism for heat loss from a planetary interior. Partial melting of the athenosphere (a) beneath a mid-ocean ridge (mor) froms the basaltic oceanic crust (b). The mantle lithosphere (ml) is gravitationally unstable and is subducted into the interior at an ocean trench (ot). As the pressure increases, the basaltic crust transforms to dense eclogite (e) and partial melting produces island arcs (ia).

Figure 2. Illustration of the heat-pipe mechanism for heat transport through the lithosphere. The evolution in time of a section of lithosphere is illustrated. Ascending convection in the asthenosphere (a) produces basaltic magma that ascends to the surface through the lithosphere by magma fracture. The extrusive flows solidify and thicken forming the basaltic crust (b). At a depth of about 70 km the pressure and temperature are sufficiently high that the basalt transforms to eclogite (e). The eclogitic lower crust and mantle lithosphere (ml) are gravitationally unstable and are recycled into the planetary interior by delamination.

Figure 3. The required surface volcanic flux $\dot{V}$ on Venus as a function of the mean thickness of the lithosphere $\bar{y}_L$ from (3).

Figure 4. The mean interval τ_i between global surface eruptions on Venus from (4) and the solidification time τ_c as a function of the flow thickness h_f from (5).

Figure 5. Topographic elevation w as a function of the increase in crustal thickness Δh_c for several values of the lithospheric thickness h_L from (14) and (15).

Figure 6. The required crustal thickness h_c and associated isostatic geoid anomaly ΔN as a function of topographic elevation w for several values of the initial crustal thickness h_{co} from (14) and (16).

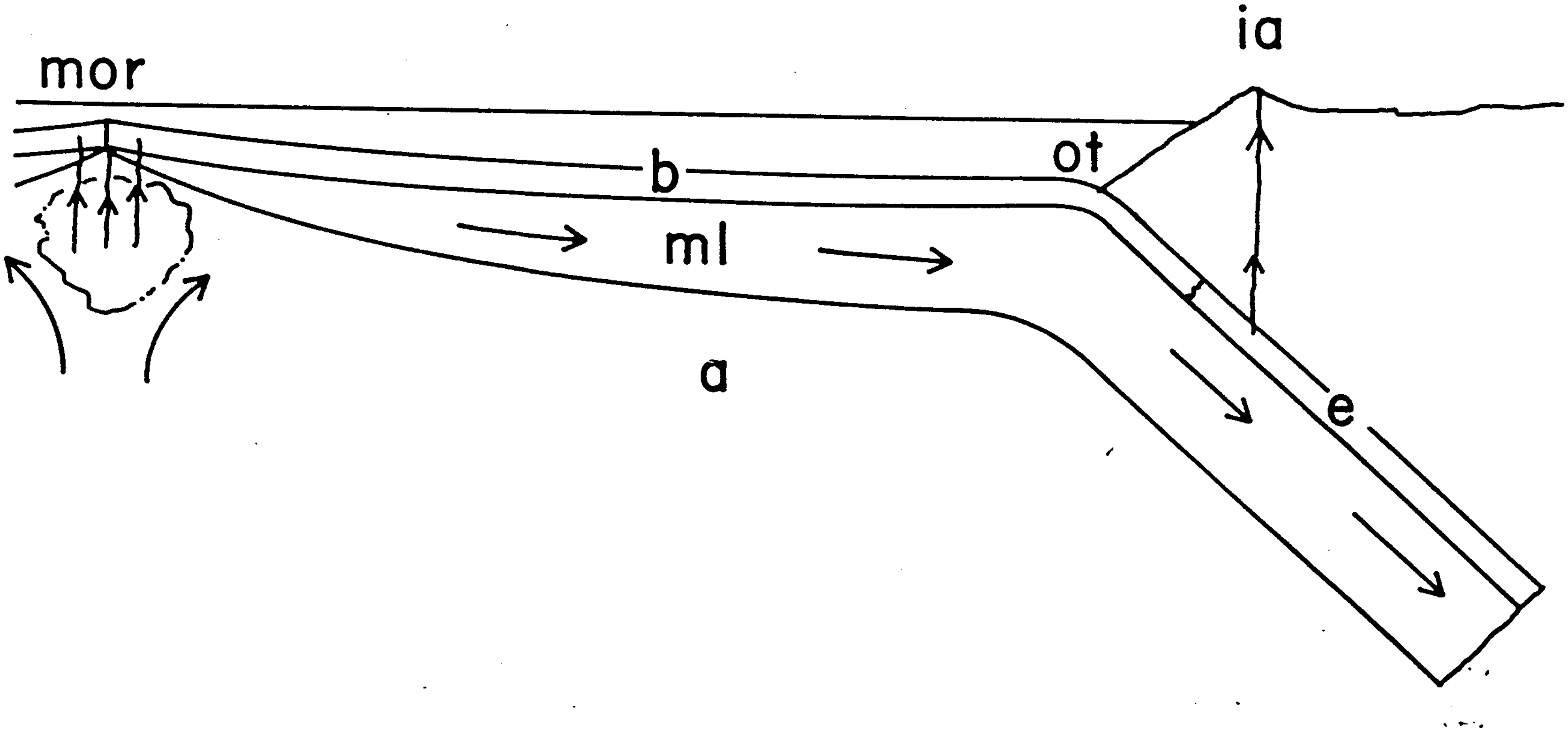

Figure 1

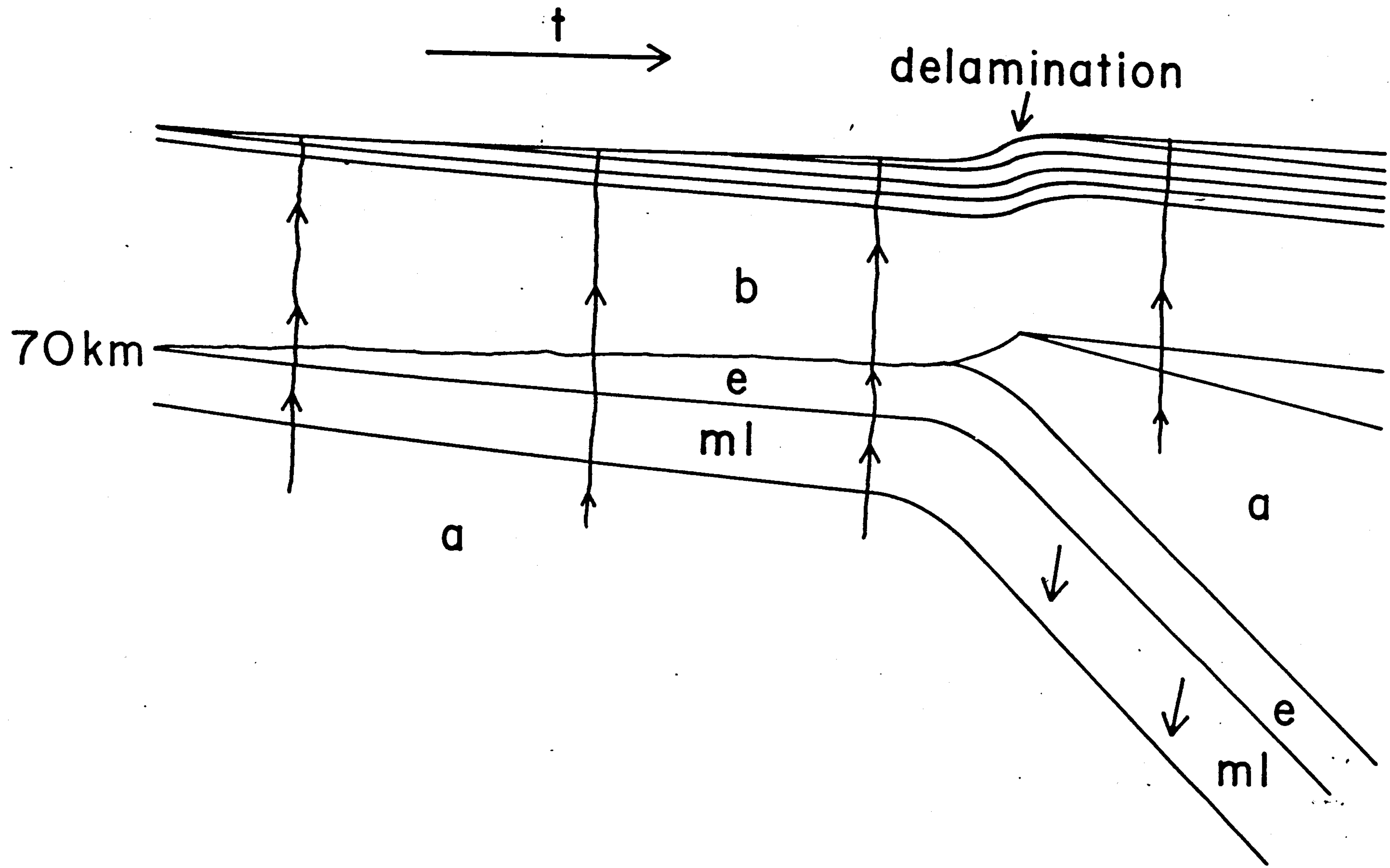

Figure 2

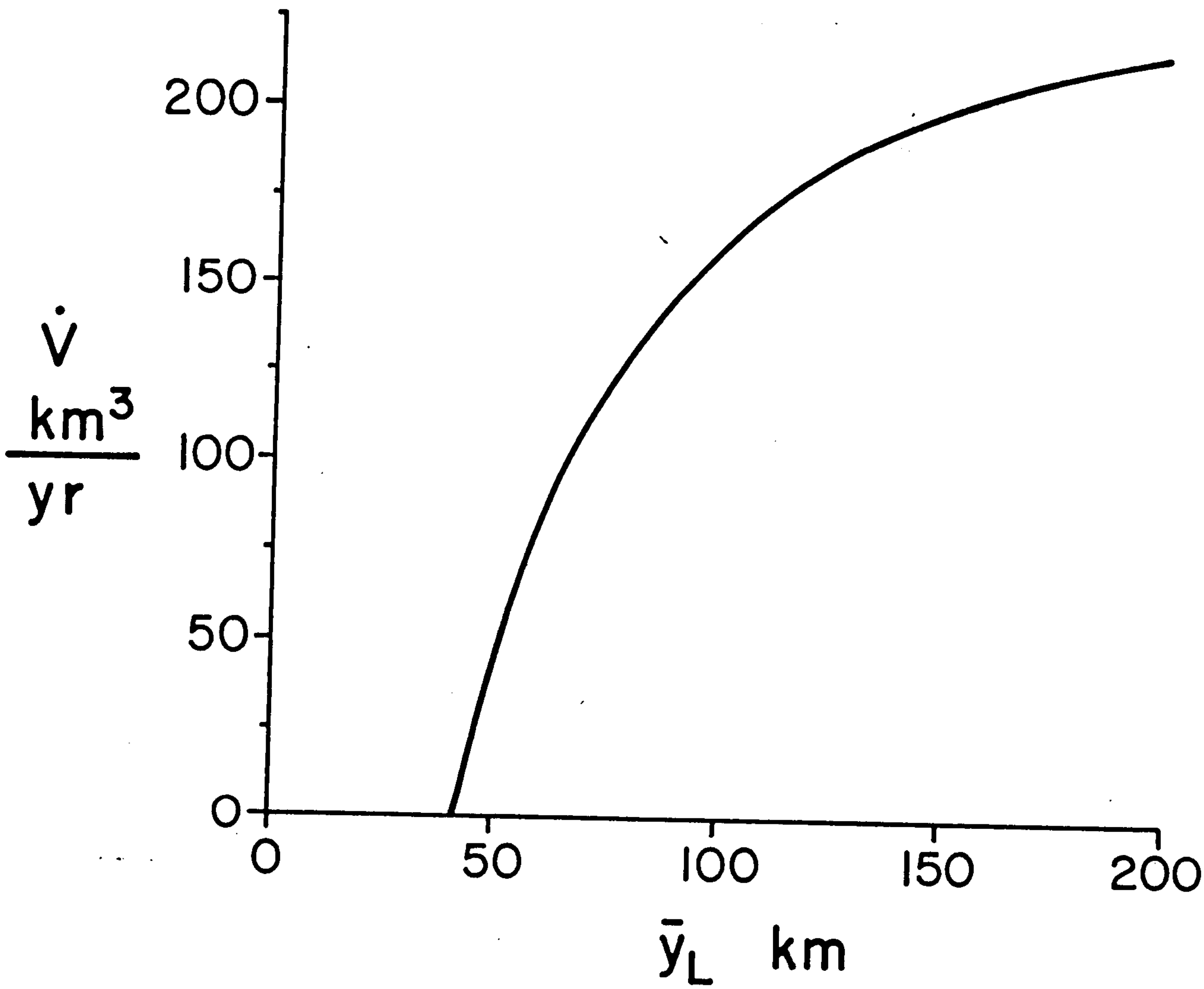

Figure 3

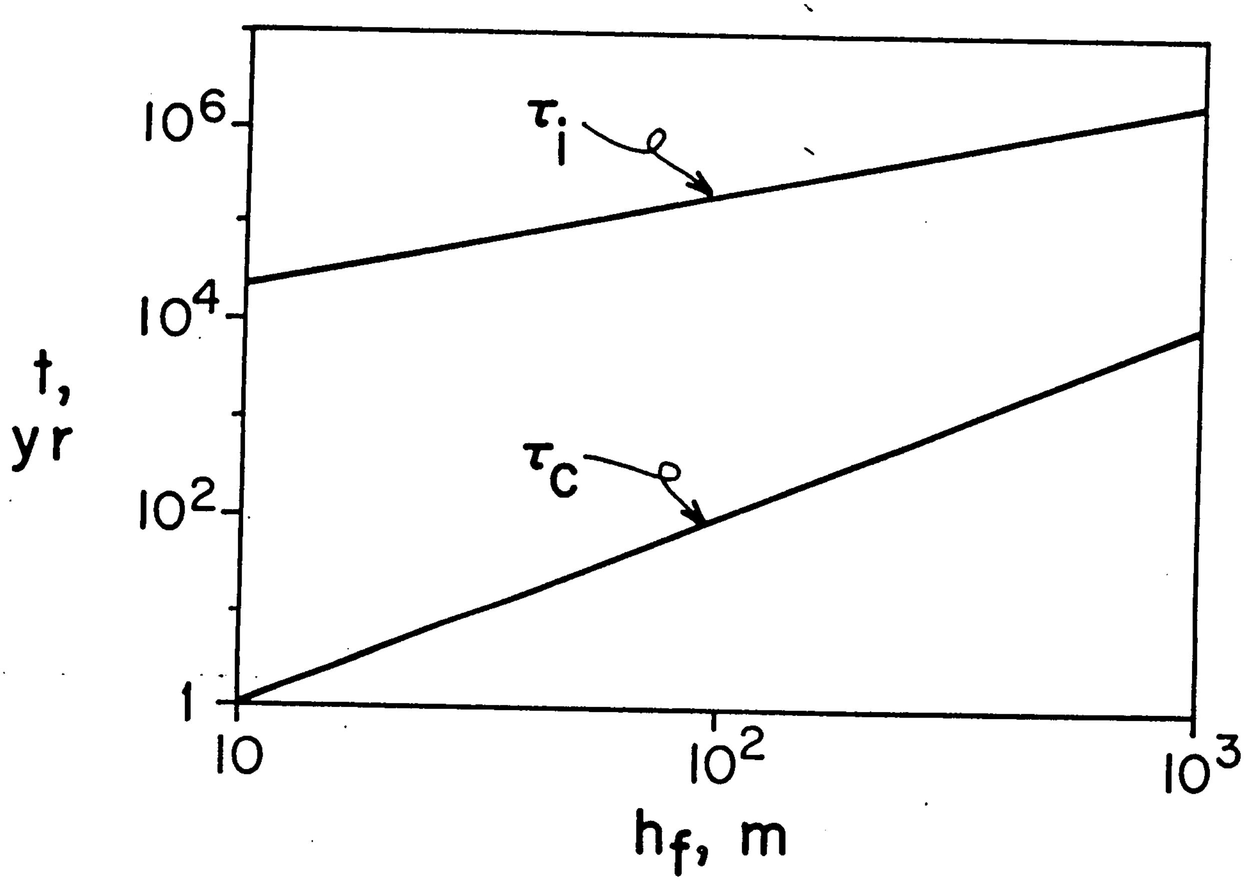

Figure 4

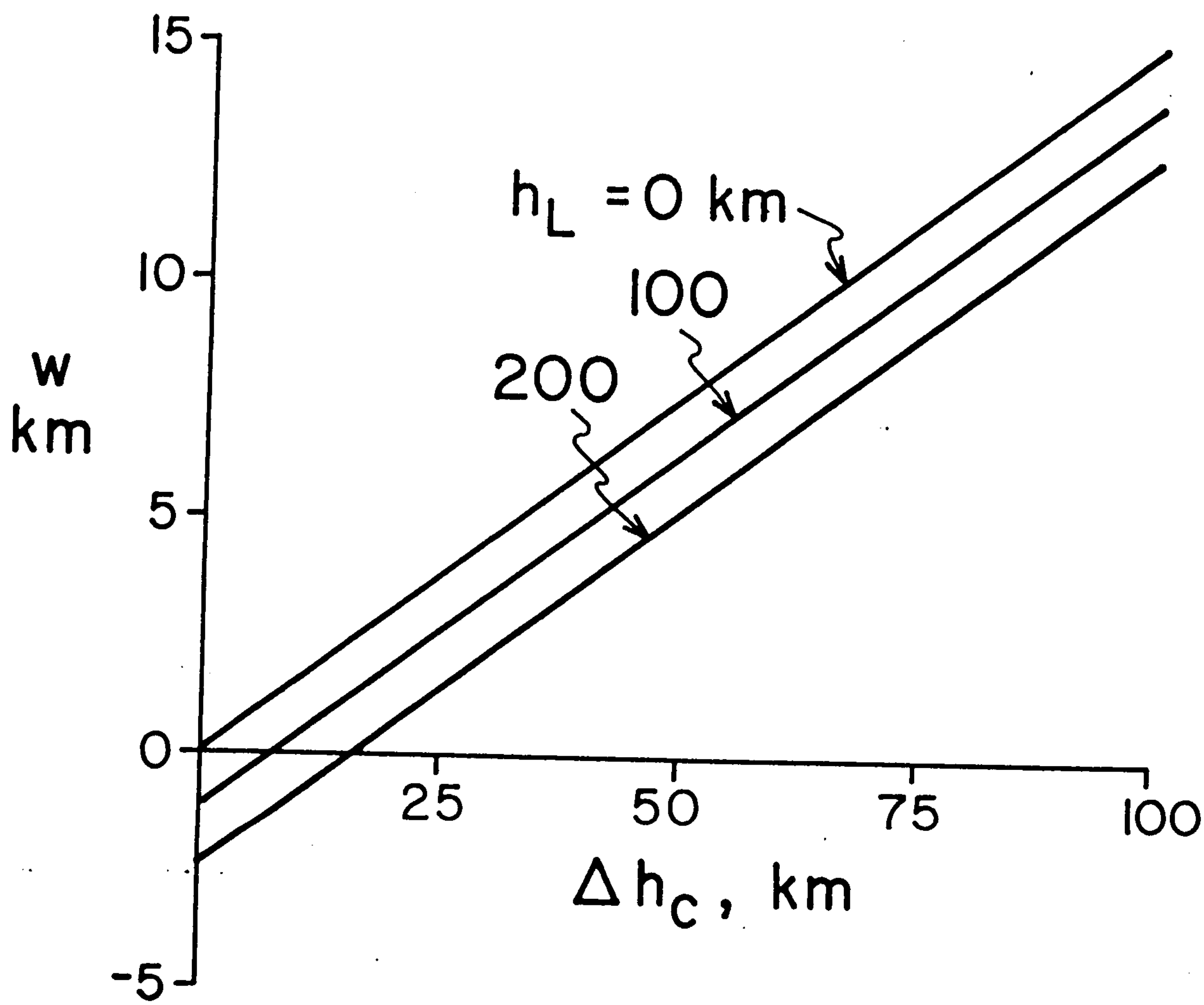

Figure 5

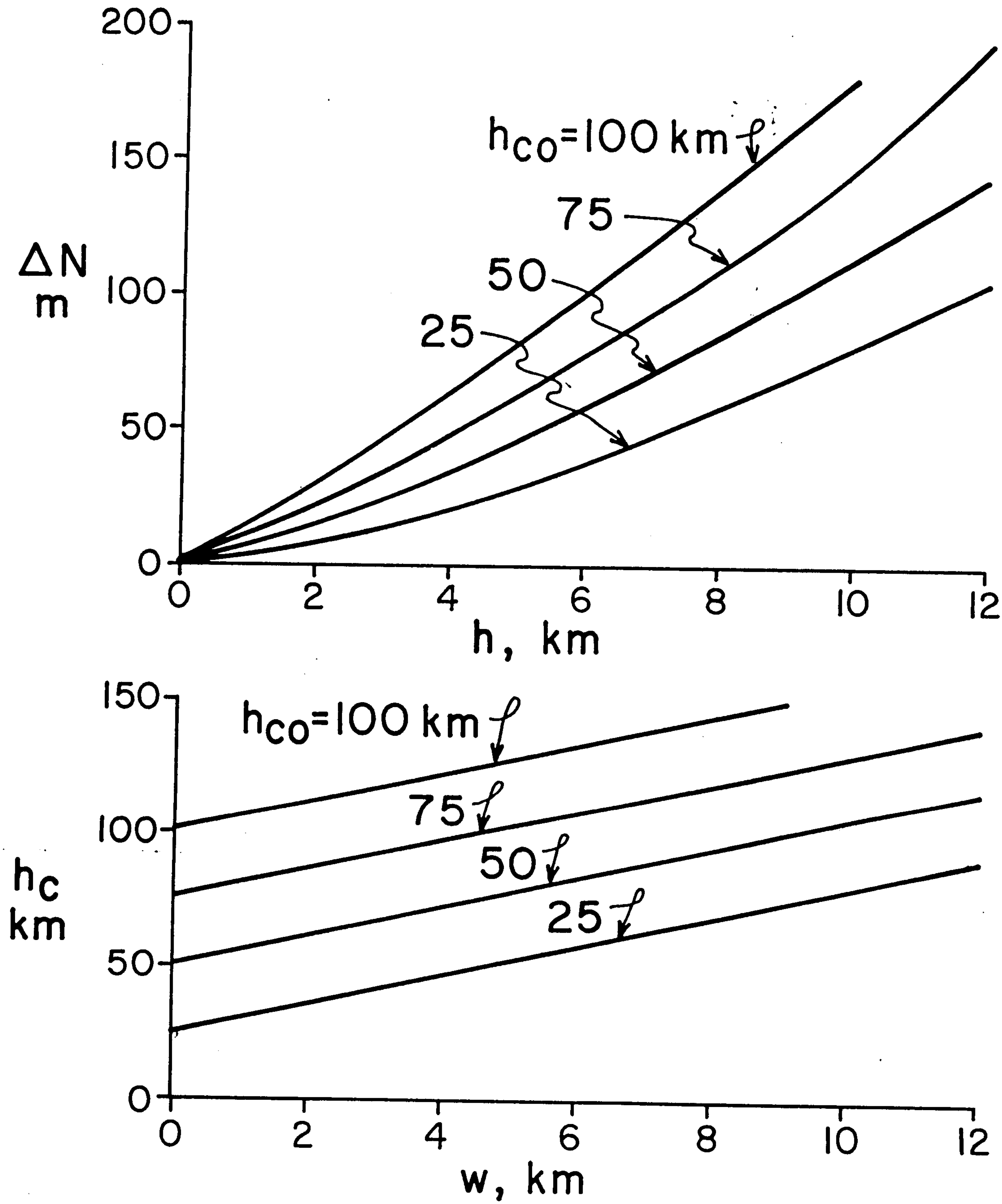
200
150
100
50
0
ΔN
m
h_{co}=100 km
75
50
25
0
2
4
6
8
10
12
h, km
150
100
50
0
h_c
km
h_{co}=100 km
75
50
25
0
2
4
6
8
10
12
w, km

Figure 6