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**A DYNAMIC MODEL OF VENUS'S GRAVITY FIELD**

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**ABSTRACT**

Unlike Earth, long-wavelength gravity anomalies and topography correlate well on Venus. Venus's admittance curve from spherical harmonic degree 2 to 18 is inconsistent with either Airy or Pratt isostasy but is consistent with dynamic support from mantle convection. A model using whole mantle flow and a high viscosity near-surface layer overlying a constant viscosity mantle reproduces this admittance curve. On Earth, the effective viscosity deduced from geoid modeling increases by a factor of 300 from the asthenosphere to the lower mantle. These viscosity estimates may be biased by the neglect of lateral variations in mantle viscosity associated with hot plumes and cold subducted slabs. The different effective viscosity profiles for Earth and Venus may reflect their convective styles, with tectonism and mantle heat transport dominated by hot plumes on Venus and by subducted slabs on Earth. Convection at degree 2 appears much stronger on Earth than on Venus. A degree 2 convective structure may be unstable on Venus, but may have been stabilized on Earth by the insulating effects of the Pangean supercontinental assemblage.

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An important difference between Venus and Earth is the strong positive correlation that exists between long-wavelength gravity and topography on Venus (Sjogren et al., 1980). This relationship is evident in comparisons of spatial domain maps (Mottinger et al., 1985, Fig. 5) and can be quantified using recent eighteenth degree and order spherical harmonic models for Venusian topography (Bills and Kobrick, 1985) and gravity (Bills and Kiefer, 1985). Figure 1 illustrates the correlation between the long-wavelength components of gravity and topography as a function of spherical harmonic degree for Venus. For comparison, the correlation between the Earth's gravity field (Lerch et al., 1982), referred to its hydrostatic figure (Nakiboglu, 1982), and "equivalent rock" topography (NOAA, 1980) is also shown. For degrees 3 through 18, the correlation is positive and statistically significant for Venus. By contrast, the correlation is generally poor for harmonic degrees less than 13 on Earth.

One way to understand the relationship between gravity and topography is to model the observed gravity as the sum of two components: a coherent component, in which topography and compensating masses are related by some type of isostasy, and an incoherent part, due to density variations which are uncorrelated with topography. Assuming a linear relationship between the gravity harmonics ( $G_{lm}$ ) and topography harmonics ( $H_{lm}$ ) (Dorman and Lewis, 1970), this can be expressed as

$$G_{lm} = F_l H_{lm} + I_{lm} \quad (1)$$

where  $F_l$  is a spectral admittance and the  $I_{lm}$  are the residual gravity coefficients. Bills and Kiefer (1985) determined the best fitting admittance curve for the Venus gravity and topography harmonics. This is shown in Figure 2, along with one standard deviation uncertainties. The degree 2 admittance is poorly constrained because of the low correlation between gravity and topography at degree 2. By comparing the observed Venusian admittance curve with theoretical admittance curves for several possible compensation mechanisms, we can gain a better understanding of the relationship between long-wavelength gravity and topography on Venus.

Bowin (1983) and Bowin et al. (1985) have argued that the high correlation between gravity and topography on Venus is best explained by variations in crustal thickness. The admittance for Airy compensation of topography at degree  $l$  on a rigid planet is

$$F_l = \frac{3\rho_c}{\rho_{av}} \frac{1}{2l+1} \left( 1 - \left( 1 - \frac{D}{R} \right)^l \right) \quad (2)$$

where  $\rho_c$  is the crustal density,  $\rho_{av}$  is the average density of Venus ( 5.24 Mg/m<sup>3</sup> ),  $R$  is the planetary radius (6051.5 km), and  $D$  is the isostatic compensation depth ( Phillips and Lambeck, 1980, equation 20). Viscous adjustment due to self-gravitation, which is important at long wavelengths, is not included in the Phillips and Lambeck model. It can be included by multiplying the expression in equation 2 by  $h_l$ , the isostatic Love number for degree  $l$ . For the Earth's radial density profile,  $h_l$  is given approximately by

$$h_l \approx \frac{l+0.6}{l} \quad (3)$$

(Hager, 1983). The similarity in mass and radius between Venus and Earth suggests that the two planets have similar radial density profiles and Love numbers. Thus, the Airy compensation admittance function becomes

$$F_l \approx \frac{3\rho_c}{\rho_{av}} \frac{l+0.6}{l(2l+1)} \left( 1 - \left( 1 - \frac{D}{R} \right)^l \right) \quad (4)$$

At the long wavelengths considered here, the admittance for a Pratt model with compensation depth  $2D$  is also given by equation 4.

Figure 2 compares Venus's observed admittance curve with admittances for Airy compensation at 100 km and 200 km depth (or Pratt compensation at 200 km and 400 km). A crustal density of 2.8 Mg/m<sup>3</sup> is assumed. The two compensation curves broadly constrain the observed admittances, indicating that topography on Venus is deeply compensated, with a best fitting compensation depth of approximately 150 km (Airy model) or 300 km (Pratt model). However, the model curves do a poor job of reproducing the shape of the observed admittance curve, so we conclude that crustal thickness

thickness variations, whether modeled as Airy or Pratt isostasy, do not satisfactorily explain the long-wavelength relationship between gravity and topography on Venus.

Phillips et al. (1981) modeled western Aphrodite Terra using a combination of regional elastic flexure and local isostatic compensation and found a best fitting compensation depth of 115 km. They argued that compensation at such a large depth could not be passively maintained for geologically long periods of time and that the gravity and topography of this region is best understood in terms of dynamic compensation from mantle convection. Similarly, apparent isostatic compensation depths in excess of 300 km for the Beta Regio area are also best understood in terms of mantle convection (Phillips and Malin, 1983). Although neither Phillips et al. nor Phillips and Malin attempted to model Venus's gravity field using a mantle convection model, their results, together with the failure of the Airy and Pratt compensation models, suggest that Venus's admittance curve may be matched using a dynamic compensation model.

In a convecting planet, mantle flow deforms both the upper surface and the core-mantle boundary. The total gravity anomaly in such a convecting system is the sum of contributions from the density contrasts driving the flow plus the density contrasts arising from the flow-induced boundary deformations. Furthermore, the upper surface deformation produces dynamic topography relative to an undeformed spheroid. Thus, in a dynamic compensation model, both gravity and topography are manifestations of mantle convection. Richards and Hager (1984) described a dynamic compensation model in which the mantle is assumed to be self-gravitating and incompressible, with a spherically symmetric Newtonian viscosity structure. Variables are expanded azimuthally in terms of spherical harmonics, and propagator matrices are used to solve the radial dependence of the equations of motion, with free-slip boundary conditions applied at both the upper surface and at the core-mantle boundary. The gravitational potential at harmonic degree  $l$  is given by

$$\delta U_{lm} = \frac{4\pi GR}{2l+1} \int_{R_{core}}^R K_l(r) \delta \rho_{lm}(r) dr \quad (5)$$

The core radius is assumed to be 3240 km (Basaltic Volcanism Study Project, 1981, pp.682-685).  $K_l(r)$  is a dynamic response function which gives the effect of a unit density contrast at harmonic degree  $l$  and radius  $r$  on the gravitational potential. Examples of  $K_l(r)$  for a variety of radial mantle viscosity profiles are given by Hager (1984, Figure 4).  $\delta\rho_{lm}(r)$  is the component of density contrast which drives the flow at degree  $l$  and order  $m$ . Similarly, the upper surface deformation is given by

$$\delta r_{lm} = \int_{R_{core}}^R D_l(r) \delta\rho_{lm}(r) dr \quad (6)$$

where  $D_l(r)$  is a surface deformation response function. For spherically symmetric viscosity,  $K$  and  $D$  do not depend on order  $m$ . The admittance is computed as

$$F_l = \frac{\sum_m \delta U_{lm} \delta r_{lm}}{g \sum_m (\delta r_{lm})^2} \quad (7)$$

On the Earth, the density contrast  $\delta\rho_{lm}(r)$  can be estimated from seismic tomography and thermal models of subducted slabs (Hager, 1984; Hager et al., 1985). For Venus, such observational evidence is not available. For simplicity, we assume that  $\delta\rho_{lm}$  is constant throughout the mantle, with no variation with depth. We have also considered models in which the density contrast increases linearly from the core-mantle boundary to the surface and models in which the density contrast is specified randomly. Both models give results which are similar to those for the depth-independent density contrast model. The insensitivity of the admittance curve to the exact form of the radial variation of  $\delta\rho_{lm}(r)$  suggests that our results are not severely biased by the assumed density contrast model.

By comparing model admittance curves for various mantle viscosity profiles, we can determine whether or not mantle convection can explain the observed long-wavelength gravity and topography on Venus. The modeling process is sensitive only to relative viscosity variations between the layers. It is not possible to determine absolute

viscosity profiles from the modeling technique used here (Richards and Hager, 1984). Figure 3 presents admittance curves for two different mantle viscosity models, assuming mantle-wide flow. The dashed line is for a mantle with uniform viscosity. While this model has the right general form, it slightly underestimates the magnitude of the observed admittance curve. A better fit can be obtained using models in which viscosity in the lower mantle is a factor of 3 to 10 less than in the upper mantle. However, the effect of increasing pressure with depth is expected to cause an increase, not a decrease, in viscosity with increasing depth, so models whose lower mantle viscosity is less than their upper mantle viscosity seem rheologically implausible. Alternatively, we consider the effect of a high viscosity near-surface layer overlying a uniform viscosity mantle. The solid line in Figure 3 is the admittance curve for a model with a 100 km thick surface layer whose viscosity is 100 times the mantle viscosity. This model satisfactorily reproduces both the shape and the magnitude of the observed admittance curve.

## Discussion

While crustal thickness variations do not successfully account for the long-wavelength relationship between gravity and topography on Venus, simple dynamic compensation models, involving whole mantle flow and only two viscosity layers, can successfully explain the observed admittance curve. Isostatic models fail because the observed admittance curve requires a different isostatic compensation depth at each harmonic degree. In a dynamic model, the peak sensitivity of the geoid to density contrasts occurs at a different depth for each harmonic degree. Low degree harmonics are more sensitive to deep density contrasts, while higher degree harmonics are sensitive to relatively shallow density contrasts. This varying depth sensitivity enables the dynamic compensation model to reproduce the observed shape of Venus's admittance curve.

Figure 4 compares Venus's observed geoid at degrees 2 through 18 (Bills and Kiefer, 1985) with the geoid predicted by the best fitting dynamic model (the solid line

of Figure 3). The visual correlation between the two maps further emphasizes the quality of fit between the dynamic model and the observations. The quality of fit between the dynamic model and the observations is considered in greater detail in Figure 5, which shows the amount of variance at each harmonic degree in the residual (non-dynamic) geoid. The results of Hager and Richards (1985) for the Earth are shown for comparison. At degree 3, the residual variance is only 14% of the observed variance in Venus's geoid. There is a general increasing trend in the residual variance with increasing harmonic degree. Beyond degree 14, the residual variance is more than 80% of the data variance. This is not surprising, because other compensation mechanisms, such as elastic flexure and local isostasy, are expected to become more important at shorter wavelengths. However, the reduction in variance due to the dynamic model is still significant even at the higher harmonic degrees.

One interesting contrast between Earth and Venus is their difference in behavior at degree 2. Relative to the components at  $l \geq 3$ , Earth has much more power at degree 2 in both the gravity field and topography than Venus does (Mottlinger et al., 1985, Fig. 9; Bills and Kobrick, 1985, Figure 2). We suggest that the poor correlation between gravity and topography at degree 2 on Venus is due to relatively small temperature or density contrasts in the mantle at this harmonic degree; the noise term therefore dominates in equation 1. In contrast, the Earth is observed to have substantial heterogeneity at degree 2 (Hager et al., 1985; Masters et al., 1982). Figure 5 shows this difference clearly: the residual variance is only 3% of the data variance at degree 2 on Earth, whereas the residual variance exceeds the data variance at degree 2 on Venus.

We speculate that the relatively weak convection we infer for Venus at degree 2 is the norm for one-plate planets, where flow occurs beneath a rigid lithosphere. Some support for this view is given by results for the onset of convection at marginal stability with a no-slip boundary condition at the surface (Chandrashekar, 1961). Earth might differ from Venus at degree 2 because it has continents which periodically assemble into



a supercontinent. Subduction would occur preferentially at the edge of this supercontinent, causing preferential cooling in this region of the mantle, with heating of the mantle by internal heat sources beneath the continent and antipodal to the continent (Anderson, 1982; Chase and Sprowl, 1983). The edge of a Pangea-like continental assemblage is predominantly a degree 2 feature, leading to an anomalously strong degree 2 component of mantle heterogeneity.

Substantial debate currently exists concerning the nature of tectonic processes on Venus (for a review, see Phillips and Malin, 1984). Gravity observations provide an important constraint on any tectonic model. The best fitting dynamic compensation models for the long-wavelength gravity fields of Earth (Hager and Richards, 1985) and Venus differ substantially in their radial viscosity profiles. Both planets appear to have high viscosity near-surface layers, as expected for the upper boundary layer of a convecting system with temperature dependent viscosity. Based on our geoid modeling, on Venus the mantle viscosity appears to remain approximately constant with depth, whereas on Earth the viscosity increases by a factor of about 300 from the asthenosphere to the lower mantle. It should be emphasized that the dynamic geoid models for both planets assume laterally symmetric viscosity. This is not true for real planets, where lateral temperature variations lead to lateral viscosity variations. Richards and Hager (in preparation) have calculated the effects of high viscosity subducted slabs and low viscosity hot plumes on the dynamic response function  $K_l(r)$ . Their numerical models show that low viscosity plumes have low degree geoid responses that cause the lower mantle/upper mantle viscosity ratio to appear smaller than that inferred for subducting slabs. The inferred apparent difference in the viscosity profiles of Venus and Earth and the dominance of the Venusian highland areas in both the gravity field and topography are consistent with tectonism and mantle heat transport being dominated by hot plumes on Venus. This is in contrast to the dominant role of plate tectonics and subduction on Earth. Although our results are suggestive of important differences in heat transport on

Earth and Venus, the approaches used to constrain the viscosity structures are different for the two planets. For Earth, we used seismic observations interpreted in terms of interior density contrasts to match the geoid, while for Venus we assume a density structure and match the admittance. Thus, we cannot at present make a definitive judgement as to whether or not the Venusian lithosphere exhibits a plate-like style or if subduction of the lithosphere is involved in convective downwelling and return flow. Answers to these questions must await future high resolution radar mapping of Venus.

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### Figure Captions

Figure 1. Correlation coefficients between gravity and topography. The solid line is Venus, the short dashed line is Earth, and the long dashed lines are 95% statistical confidence limits.

Figure 2. Gravity-topography spectral admittances. The open circles are observed values with one standard deviation uncertainties. The solid line is for Airy compensation at 100 km depth or Pratt compensation at 200 km depth. The dashed line is for Airy compensation at 200 km or Pratt compensation at 400 km.

Figure 3. Observed spectral admittance, as in Fig. 2. The dashed line is for dynamic compensation using whole-mantle convection and constant viscosity. The solid line is for whole-mantle convection with a high viscosity surface layer overlying a constant viscosity mantle.

Figure 4. Comparison of Venus's observed and dynamically predicted geoids. Fig. 4a is the observed geoid at degrees 2-18 and Fig. 4b is the dynamically predicted geoid using the solid line model of Fig. 3. Cylindrical equidistant projection. The contour interval is 20 m; lows are shaded.

Figure 5. Residual variance between observed and dynamically predicted geoids for Venus (solid line) and Earth (dashed line).

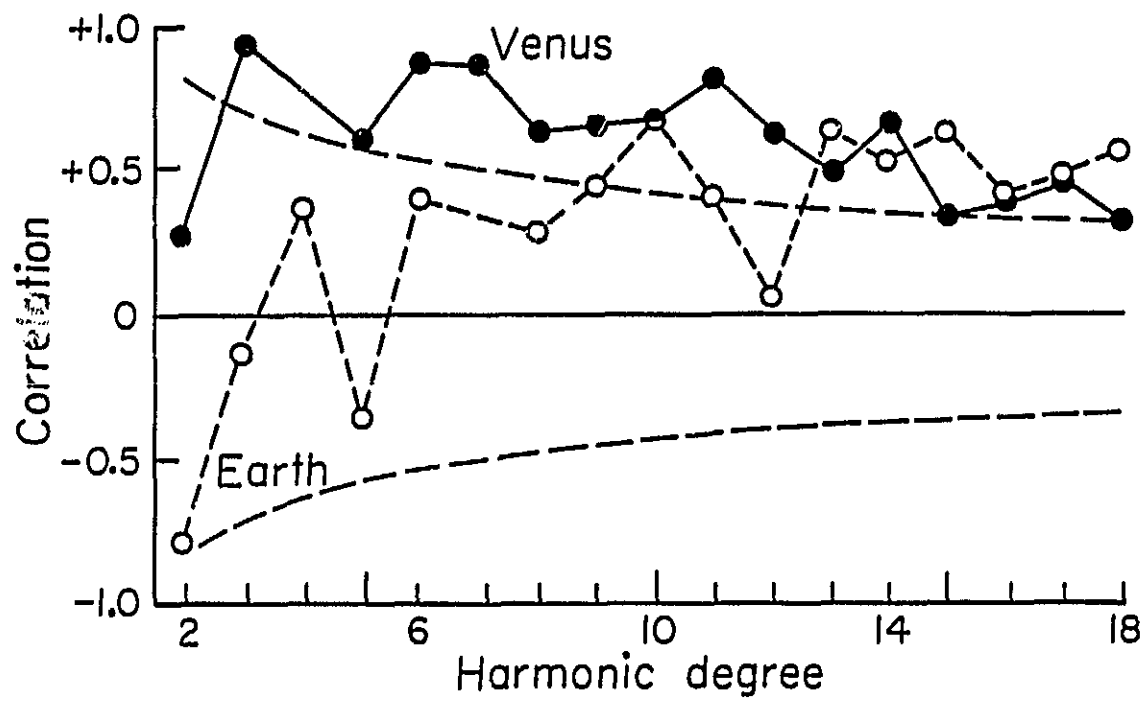
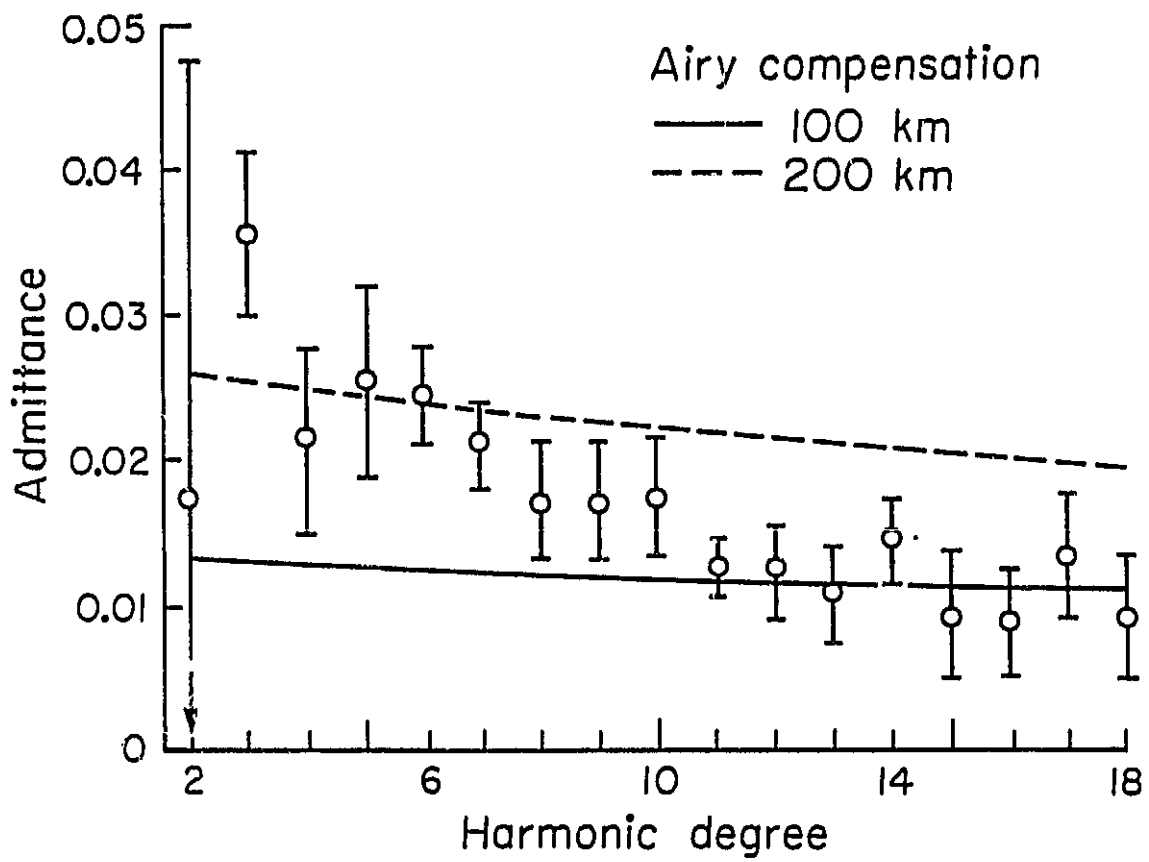
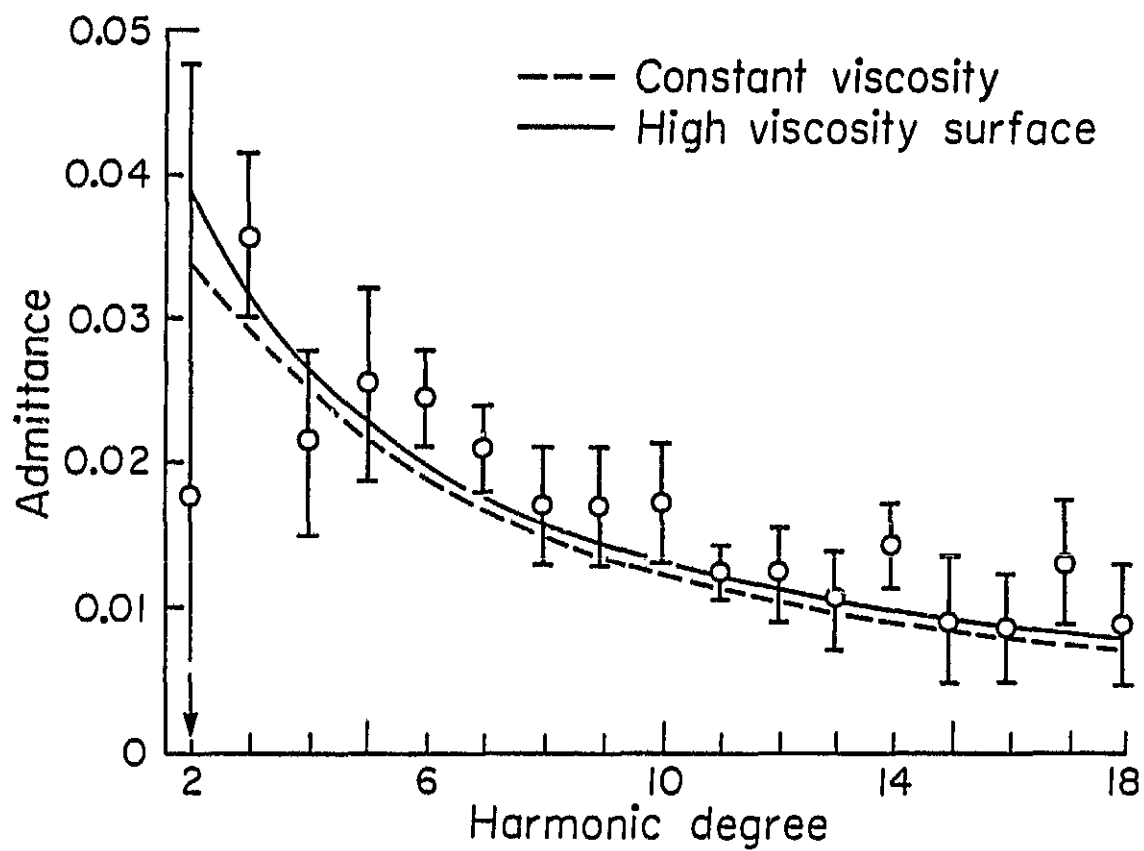


Fig. 1



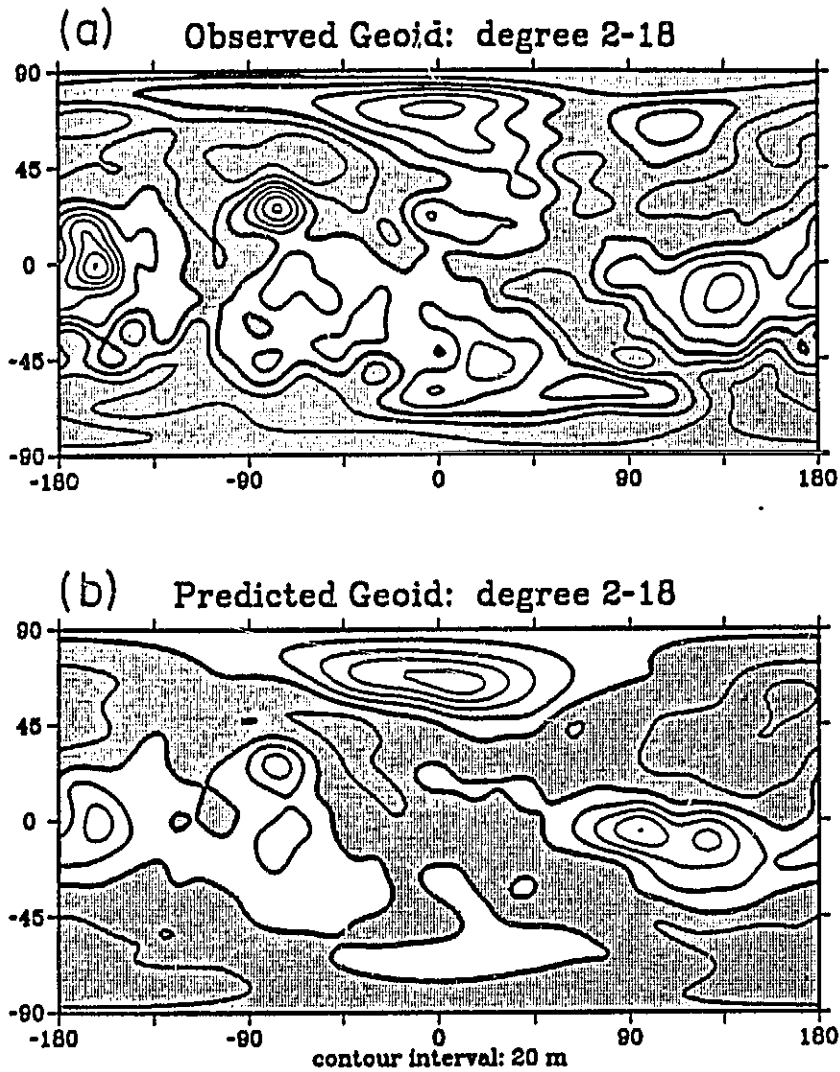
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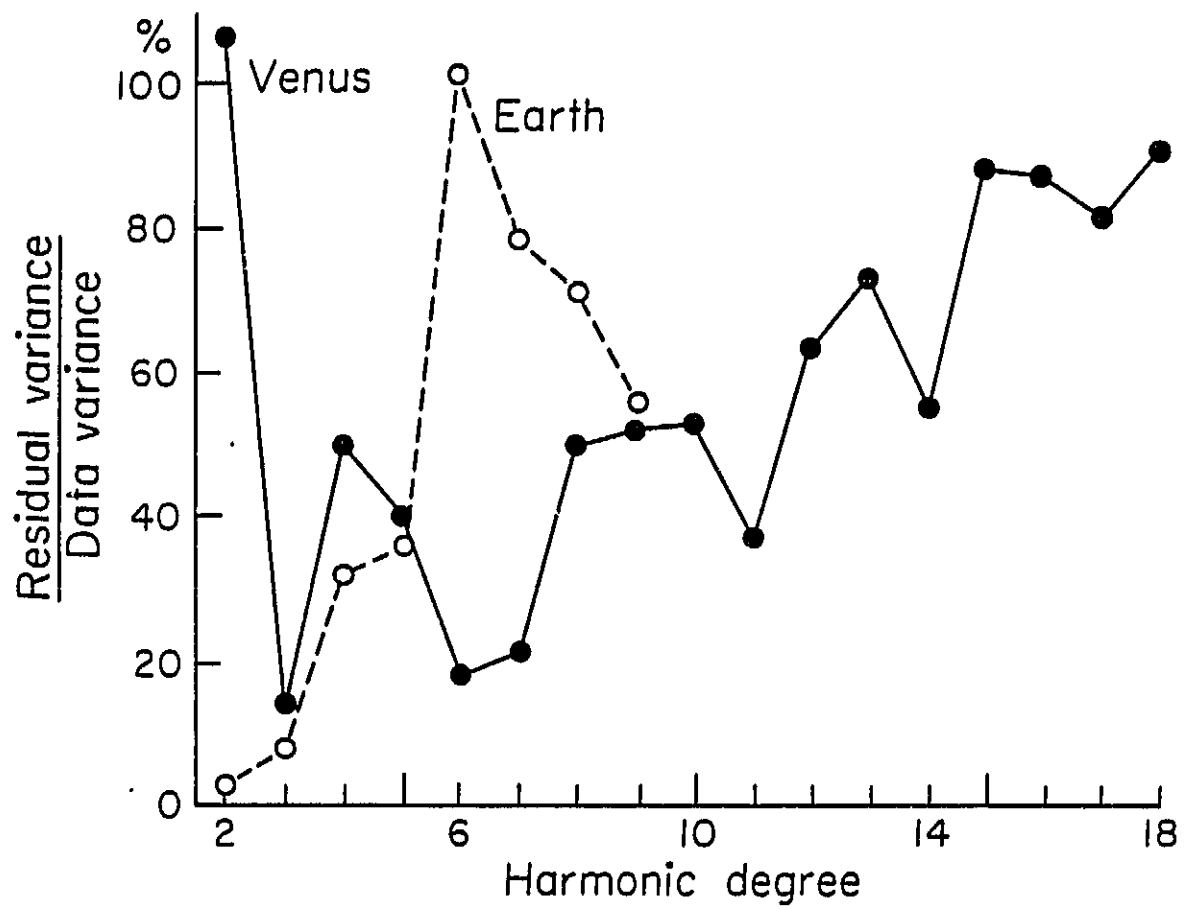
*Fig. 3*



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*Fig. 4*



*Fig. 5*