is completely unlike any size distribution on the terrestrial planets. Therefore, the computer simulations strongly indicate that the case 1 equilibrium resurfacing model is not a valid explanation for the resurfacing history of Venus. The observed nonrandom distribution of volcanic features [3] and the noncorrelation of the density of impact craters and volcanic features in equal areas [2] are further arguments against the equilibrium resurfacing model.

Case 2 of the equilibrium resurfacing model (≥10% resurfacing areas) simply will not work, except at the 100% (global) resurfacing level. Figure 1d is a Monte Carlo simulation for constant 10% resurfacing areas with a constant 50-m.y. time interval between events. Clearly the crater distribution is completely nonrandom and totally different from the observed distribution. We have done simulations for 25% and 50% resurfacing areas with similar results. Therefore, the equilibrium resurfacing model is not a valid model for an explanation of the observed crater population characteristics or Venus' resurfacing history.

The global resurfacing model is the most likely explanation for the characteristics of Venus' cratering record. The amount of resurfacing since that event, some 500 m.y. ago, can be estimated by a different type of Monte Carlo simulation. In this simulation the cratering record begins 500 m.y. ago with the observed crater size distribution. Our first simulation randomly selects craters from this size distribution and resurfaces areas with volcanos randomly selected from the observed volcano size distribution. The time interval between volcanic events is chosen so that only 4% of the craters are flooded at the end of 500 m.y. To date, our initial simulation has only considered the easiest case to implement. In this case the volcanic events are randomly distributed across the entire planet and, therefore, contrary to observation, the flooded craters are also randomly distributed across the planet. This simulation results in a maximum resurfaced area of about 10% of the planet since the global event, and an obliteration of about 4% of the craters. Future simulations will take into account the observed nonrandom distribution of flooded craters and, therefore, the nonrandom distribution of volcanic events. These simulations will probably result in a lower percentage of planet resurfacing because volcanism will be concentrated in smaller areas.

References: [1] Phillips R. J. et al. (1992) *LPSC XXIII*, 1065–1066. [2] Schaber G. G. et al. (1992) *JGR*, special Magellan issue, in press. [3] Head J. W. et al. (1992) *JGR*, special Magellan issue, in press.

N93-14312

METHANE MEASUREMENT BY THE PIONEER VENUS LARGE PROBE NEUTRAL MASS SPECTROMETER. T. M. Donahue[1] and R. R. Hodges Jr.[2], [1]University of Michigan, Ann Arbor MI 48109, USA, [2]University of Texas at Dallas, Richardson TX, 75083, USA.

The Pioneer Venus Large Probe Mass Spectrometer detected a large quantity of methane as it descended below 20 km in the atmosphere of Venus. Terrestrial methane and ^{136}Xe, both originating in the same container and flowing through the same plumbing, were deliberately released inside the mass spectrometer for instrumental reasons. However, the ^{136}Xe did not exhibit behavior similar to methane during Venus entry, nor did CH_4 in laboratory simulations. The CH_4 was deuterium poor compared to Venus water and hydrogen. While the inlet to the mass spectrometer was clogged with sulfuric acid droplets, significant deuteration of CH_4 and its H_2 progeny was observed. Since the only source of deuterium identifiable was water from sulfuric acid, we have concluded that we should correct the HDO/H_2O ratio in Venus water from 3.2×10^{-2} to $(5 \pm 0.7) \times 10^{-2}$.

When the probe was in the lower atmosphere, transfer of deuterium from Venus HDO and HD to CH_4 can account quantitatively for the deficiencies recorded in HDO and HD below 10 km, and consequently, the mysterious gradients in water vapor and hydrogen mixing ratios we have reported. The revision in the D/H ratio reduces the mixing ratio of water vapor (and H_2) reported previously by a factor of 3.2/5.

We are not yet able to say whether the methane detected was atmospheric or an instrumental artifact. If it was atmospheric, its release must have been episodic and highly localized. Otherwise, the large D/H ratio in Venus water and hydrogen could not be maintained.

N93-14313

VISCOELASTIC RELAXATION OF VENUSIAN CORONAE AND MOUNTAIN BELTS: CONSTRAINTS ON GLOBAL HEAT FLOW AND TECTONISM. I. Duncan and A. Leith, Department of Earth and Planetary Science, Washington University, St. Louis MO 63130, USA.

Venus differs from Earth in that water is essentially absent and its surface temperatures are about 470 K higher. The competing effects of high surface temperature and dry lithologies on the long-term history of surface topography have been studied using the finite-element method (Tecton) [1].

The relaxation history of surface topographic features, such as coronae and mountain belts, is a function of thermal gradient, crustal thickness and lithology, regional stresses, and basal tractions applied to the lithosphere. In this study we have examined the relative effects of these factors over a period of 500 Ma (presumed to be the mean age of the venusian surface) [2].

We assume that the venusian crust is composed of various combinations of diabase, gabbro, komatiite, and refractory lithologies such as anorthosite and websterite. Using appropriate thermal conductivities and surface heat fluxes scaled from Earth values (with and without a secular cooling contribution from the core) [3,4], thermal gradients ranging from about 20 K km^{-1} to 60 K km^{-1} are computed. We further assume that the thickness of a diabase crust is limited by the dry solidus. The models are dynamically isostatically balanced, using an elastic foundation.

Preliminary results of the study are shown in Fig. 1, in which a 2-km-high volcanic plateau has been instantaneously emplaced on the surface. For this model the crustal thermal gradient was 28 K km^{-1}. After the elastic response (essentially representing initial isostatic balance) the topography relaxes until the plateau is about 230 m above the surrounding region, and the slope from ridge crest to moat has been reduced from an initial 6° to about 2.5°. The values we obtain for our model plateau heights and slopes are in the observed range for venusian coronae. Thus we argue that coronae on Venus can be modeled as the product of elastoviscous relaxation of volcanic plateau. Although our starting models are oversimplifications, they do show all the critical morphological characteristics of venusian coronae. Matching the observed spectrum of corona morphology by varying the size, initial slope, and rheology of model plateaus enables constraints to be placed on plausible values for the venusian heat flux. We argue that the mean global heat flux must be significantly lower (~1/3) to be consistent with the observed spectrum of coronae topography.

We are presently examining models similar to those described above to investigate venusian mountain belts. Our models differ

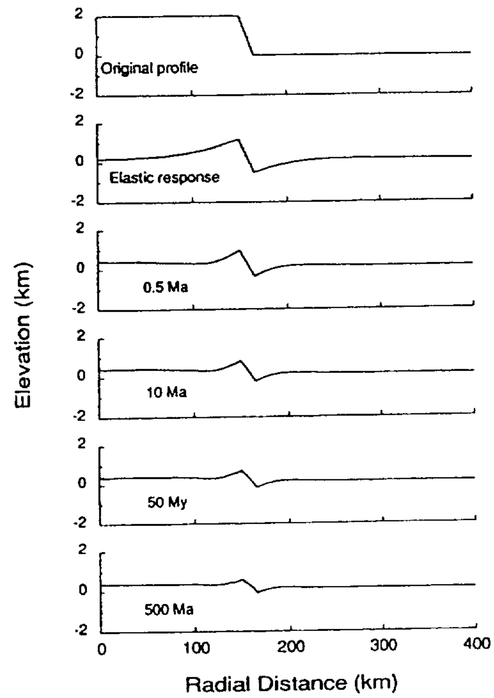

Fig. 1. Radial profiles across an initially 2-km-high and 150-km-radius plateau. The plateau and upper 2.5 km of the crust are of diabase composition and the lower 2.5 km of crust is of gabbro and overlies a peridotite mantle. The lithosphere is 33 km thick. The thermal gradient through the upper crust is 28 K km^{-1}. Topographic profiles are shown for the original plateau, the instantaneous elastic response to loading (isostatic balance), and for 0.5, 10, 50, and 500 Ma. Vertical exaggeration is ×21.

from those of other workers in that (1) our thermal models are distinctive and we believe more realistic and (2) our models are dynamically isostatically balanced with both hydrostatic restoring forces and dynamic support by tectonic stress and basal tractions being modeled. The implications of our models' stress evolution for surface deformation can be tested against Magellan imagery.

References: [1] Melosh H. J. and Raefsky A. (1980) *Geophys. J. R. Astron. Soc., 60,* 333–354. [2] Phillips R. J. et al. (1992) *LPSC XXIII,* 1065–1066. [3] Solomon S. C. and Head J. W. (1984) *JGR, 87,* 9236–9246. [4] Phillips R. J. and Malin M. C. (1984) *Annu. Rev. Earth Planet. Sci., 12,* 411–443.

N93-14314

FLEXURAL ANALYSIS OF UPLIFTED RIFT FLANKS ON VENUS. Susan A. Evans, Mark Simons, and Sean C. Solomon, Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge MA 02139, USA.

Introduction: Knowledge of the thermal structure of a planet is vital to a thorough understanding of its general scheme of tectonics. Since no direct measurements of heat flow or thermal gradient are available for Venus, most estimates have been derived from theoretical considerations or by analogy with the Earth [1]. The flexural response of the lithosphere to applied loads is sensitive to regional thermal structure. Under the assumption that the yield strength as a function of depth can be specified, the temperature gradient can be inferred from the effective elastic plate thickness [2]. Previous estimates of the effective elastic plate thickness on Venus range from 11–18 km for the foredeep north of Uorsar Rupes [3] to 30–60 km for the annular troughs around several coronae [4,5]. Thermal gradients inferred for these regions are 14–23 K km^{-1} and 4–9 K km^{-1} respectively [3,4]. In this study, we apply the same techniques to investigate the uplifted flanks of an extensional rift. Hypotheses for the origin of uplifted rift flanks on Earth include lateral transport of heat from the center of the rift, vertical transport of heat by small-scale convection, differential thinning of the lithosphere, dynamical uplift, and isostatic response to mechanical unloading of the lithosphere [6]. The last hypothesis is considered the dominant contributor to terrestrial rift flanks lacking evidence for volcanic activity, particularly for rift structures that are no longer active [6]. In this study, we model the uplifted flanks of a venusian rift as the flexural response to a vertical end load.

Tectonic Environment: We examine a linear rift system centered at 33°S, 92°E, in an area to the east of Aino Planitia. The feature appears as a linear ridge in Pioneer Venus altimetry and thus has been named Juno Dorsum. However, the increased resolution of Magellan images and topography has established that this feature is actually a linear rift with pronounced flanking highs. The rift is 100 km wide and 450 km long; it has a central depression 1–2 km deep and flanks elevated by as much as 1 km (Figs. 1 and 2). The rift connects a 1.5-km-high volcano on its western edge to two coronae, Tai Shan and Gefjun [7], to the east. Despite the presence of the volcano and coronae, both the center of the rift and the adjacent flanks appear to be free of volcanic flows. Juno Dorsum is at the end of a nearly continuous chain of coronae, rifts, and linear fractures that extends eastward and appears to terminate at the northwestern edge of Artemis Corona. The rift appears in Magellan images as numerous east-northeast-trending lineaments, which we interpret to be normal faults. To the north and south are smooth, radar-dark plains that stand very close to mean planetary elevation.

We use Magellan altimetric profiles from orbits 965–968 and 970–972 (Figs. 1 and 2). While topographic profiles perpendicular to the strike of the rift are preferable, the orbit tracks cross the rift

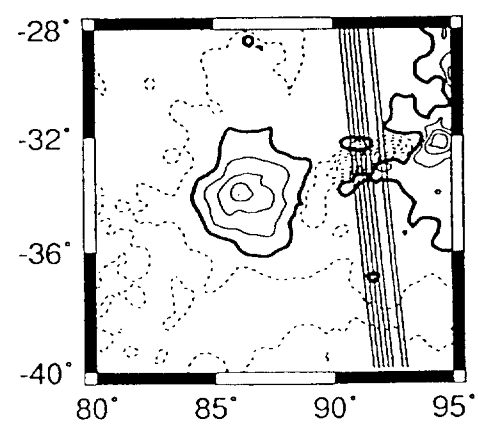

Fig. 1. Topographic contour map of the Juno Dorsum region. Thin solid lines denote positive elevation contours, thick solid lines zero elevation, and dashed lines negative elevations. The contour interval is 0.3 km, and the datum is mean planetary radius, 6051.9 km. The rift is located in the center right of the figure, with an unnamed volcano to its west. The north-south-trending lines indicate the tracks of orbits 965–968 and 971–972 (numbers increasing eastward). The orbit tracks make an angle of approximately 75° with the strike of the rift.