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Critical Taper Wedge Mechanics of Fold-and-Thrust Belts on Venus: Initial Results From Magellan

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Fold-and-thrust belts exist on Venus at the margins of crustal blocks such as plateaus, tesserae, and coronae and as ridge belts within the plains. These fold belts display a number of key features that are consistent with their formation by critical taper wedge mechanics, a mechanics that is well known for fold-and-thrust belts and accretionary wedges on Earth. For example, an analysis of fold geometry at the toe of the Artemis Chasma fold belt indicates fault-bend folding above a regionally extensive decollement horizon at a depth of about 1.5 km. Near-surface deformation on Venus is interpreted to be brittle and is anticipated to be dominated by cohesive strength in the upper 1-2 km. Critical taper wedge mechanics under anticipated Venus conditions suggests that brittle wedges should have maximum surface slopes in the range 10-20°, which is similar to some estimated slopes in the steep parts of the fold belts. The low taper toes of fold belts may be cohesion-dominated toes on either brittle or plastic decollement horizons. Once the base of the wedge undergoes the brittle-plastic transition, the surface slope is expected to flatten to near horizontal, in qualitative agreement with many topographic profiles of fold-and-thrust belts on Venus. The estimated depth of the brittle-plastic transition is uncertain based on rock mechanics data but is expected to be close enough to the surface to be affected by the atmospheric-temperature gradient. The relief of fold belts (measured between the toe of the wedge and the flat crest) displays a remarkable linear dependence on absolute elevation (Figure 15), ranging from 6 km for Maxwell Montes at an elevation of +10 km to a few hundred meters at the lowest planetary elevations (0 to –2 km). This remarkable phenomenon appears to reflect an absolute elevation dependence of the depth of the brittle-plastic transition, possibly controlled by an isostatic coupling of elevation, lithospheric thickness, and geothermal gradient.

INTRODUCTION

Fold-and-thrust belts and accretionary wedges at the margins of compressive mountain belts on Earth display critical taper wedge mechanics over a very wide range of geologic and mechanical conditions [e.g., *Dahlen and Suppe*, 1988]. This style of deformation, which is essentially the mechanics of a wedge of soil pushed by a bulldozer, is characterized by significant compressive shortening above a basal zone of detachment or decollement, below which deformation is minimal (Figure 1). Wedge theories exist for material displaying a variety of brittle, plastic, and viscous rheologies [*Chapple*, 1978; *Davis et al.*, 1983; *Emmerman and Turcotte*, 1983; *Stockmal*, 1983; *Dahlen*, 1990]; however, brittle behavior is by far the most significant on Earth in the upper crust, except where plastic salt is present [*Davis et al.*, 1983; *Davis and Engelder*, 1985]. It is important to realize that whereas critical taper fold-and-thrust belts are a response to large compression—normally supplied by plate motion—plate tectonics is not required. For example, fold-and-thrust belts exist at the base of sliding continental margins and deltas (e.g., offshore Texas [*Worrall and Snelson*, 1989; *Mount*, 1989; *Mount et al.*, 1990]) and on the flanks of expanding volcanoes on Earth and Mars [*Borgia et al.*, 1990].

Because the requirements for critical taper wedge mechanics are quite general, we must take seriously the possibility of this style of compressive deformation under the different tectonic and mechanical conditions of Venus. A viscous wedge theory has already been applied in preliminary fashion to Venus [*Vorder Bruegge and Fletcher*, 1990]. Our agenda here is to begin the sort of in-depth evaluation of possible wedge mechanics on Venus that is just now becoming possible with the high-resolution imagery and altimetry of the extended Magellan Mission.

First, we briefly introduce examples of fold-and-thrust belts from a variety of tectonic settings on Venus. Next, we outline what might be predicted for the mechanics of fold-and-thrust belts on Venus based on wedge theory, rock mechanics data, and currently known conditions on Venus. Finally, we present an initial comparison of these predictions with the new Magellan data.

STRUCTURAL INTERPRETATION OF RADAR IMAGES

Before we turn to the synthetic aperture radar images of intensely deformed fold-and-thrust belts we should briefly note some criteria for distinguishing folds from faults, as well as some difficulties in interpreting radar images. Illumination is from the west (left) in the case of first-cycle Magellan images. Because of the lack of extensive erosion, faults have topographic scarps with quite sharply defined upper edges. For example, a set of normal faults strike toward the northwest in Figure 2. The scarps facing the southwest are illuminated and bright, whereas the scarps facing the northeast are dark. In both cases the top of the scarp is a sharply defined line of tonal contrast. In contrast, the northeast striking folds in Figure 2 display a more gradual tonal transition between the fold limbs and the crest. The northwest facing limbs are bright, and the southeast facing limbs are dark. Thus the sharpness of the tonal transition is helpful in distinguishing faults from folds. However, some folds on Earth show much more angular hinges than those in Figure 2 and very steep fold limbs; if such steep fold limbs exist on Venus, they may be more difficult to distinguish from fault scarps. Furthermore, in areas of complex local relief energy may not be mapped back into its proper relative spatial position because of

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ambiguities in the radar method (the layover problem, see *Ford et al.* [1989]), which presents a significant challenge for interpretation of images.

FOLD-AND-THRUST BELTS ON VENUS

Compressive mountain belts on Venus were already known with substantial certainty from pre-Magellan data. Radar altimetry from Pioneer Venus [*Masursky et al.*, 1980], Earth-based radar images from Arecibo [*Campbell et al.*, 1983], and both images and altimetry from Venera 15-16 [*Basilevsky et al.*, 1986] have all shown surface features interpreted as compressional or probably compressional. For example, *Crumpler et al.* [1986] noted that the highlands surrounding Lakshmi Planum display many of the features of orogenic belts on Earth; they observed broad arches 20-50 km wide and hundreds of kilometers long on Venera and Arecibo images. These regularly spaced bright and dark bands were interpreted to be compressional features (anticlines and synclines) within areas of high topography. *Head* [1990] proposed that the Freyja Montes-Uorsar Rupes region represents a site of significant under thrusting and crustal accretion based on the style of folding and distinctive topography of the region. Furthermore, ridge belts in the low plains seen on Venera 15 and 16 radar images were interpreted to be compressional [*Barsukov et al.*, 1986; *Frank and Head*, 1990]. Analysis of synthetic aperture radar

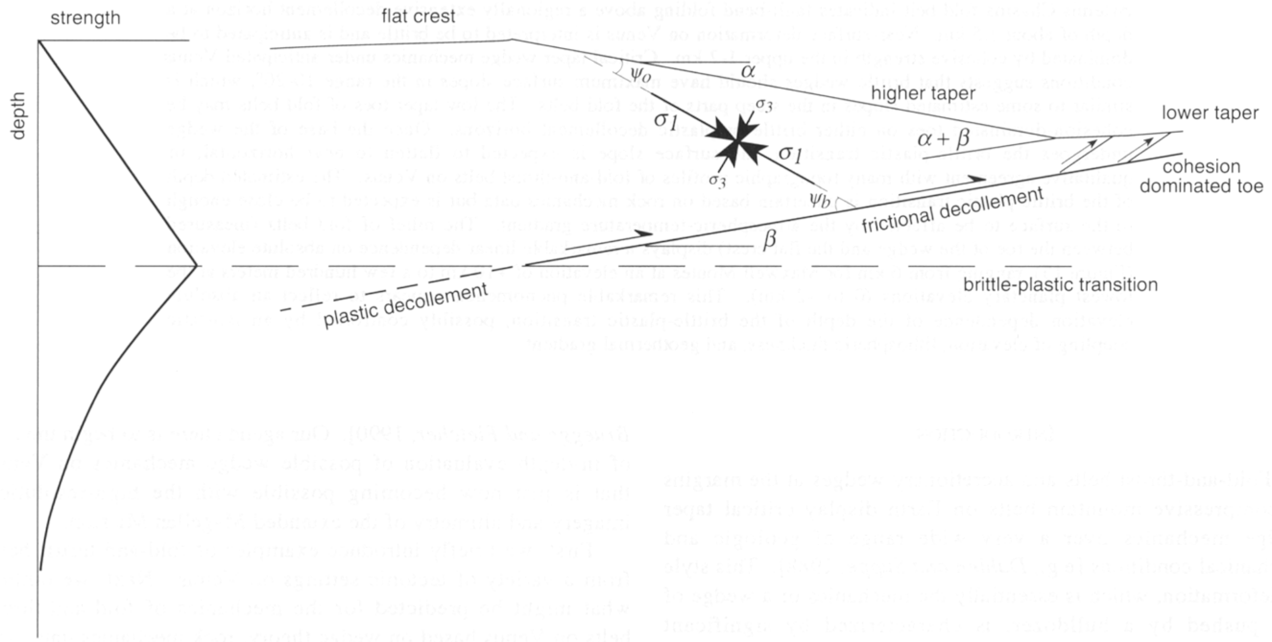

Fig. 1. Schematic cross section of a critical taper wedge showing orientations of principal compressive stresses, σ_1 and σ_3 relative to the topographic slope α and basal decollement dip β. Note that the flattening of the surface slope in the interior of the wedge is thought to be caused by the brittle-plastic transition on the decollement.

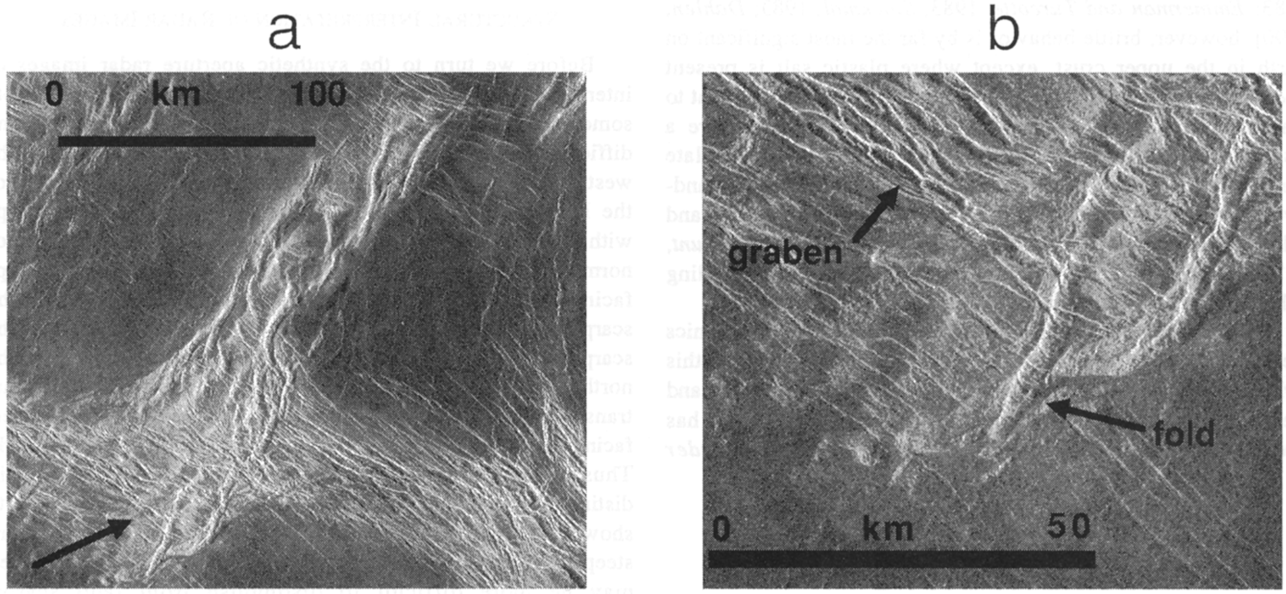

Fig. 2. Example of distinction between normal faults and folds on radar imagery. (*a*) Located at about 38°S, 348° and is 200 km across. (*b*) An enlargement of Figure 2*a*. Illumination is from the left and north is up.

TABLE 1. Elevations of Fold-and-Thrust Belts of Venus Identified From Cycle 1 Magellan Imagery

Latitude, deg N	Longitude, deg	Elevation of Toe, km	Elevation of Crest, km	Change in Elevation, km	Region
64	2	6055.5	6061.2	5.7	southern Maxwell Montes
-40	19	6051.4	6052.0	0.6	near Lada Terra
77	33	6051.4	6053.8	2.5	margin of Fortuna Tessera
60	56	6050.7	6052.5	1.8	margin of Laima Tessera
7	76	6051.6	6054.1	2.5	Semuni Dorsa
79	79	6050.3	6051.1	0.8	tessera margin
2	83	6052.3	6054.8	2.5	northern Ovda Regio
51	86	6051.2	6052.0	0.8	margin of Tellus Tessera
61	132	6050.6	6053.0	2.4	Nightingale Corona annulus
-36	144	6051.0	6052.7	1.7	eastern Artemis Corona annulus
50	150	6051.2	6052.0	0.8	Niobe Planitia
42	156	6050.6	6051.4	0.8	Niobe Planitia
39	157	6050.6	6051.6	1.0	Niobe Planitia
37	157	6051.1	6051.8	0.7	Niobe Planitia
31	158	6050.6	6051.8	1.2	Niobe Planitia
44	163	6050.7	6051.6	0.9	Niobe Planitia
51	165	6050.9	6051.2	0.3	Vinmara Planitia
51	170	6050.7	6051.3	0.6	Vinmara Planitia
16	179	6051.5	6052.9	1.4	Rusalka Planitia
11	179	6051.7	6052.9	1.2	Rusalka Planitia
46	249	6050.9	6051.7	0.8	tessera margin
68	318	6054.9	6057.7	2.8	Akna Montes
60	320	6051.8	6054.8	3.0	Vesta Rupes
73	332	6055.5	6058.2	2.8	western Freyja Montes
59	336	6052.3	6056.4	4.1	Danu Montes
77	338	6051.3	6054.0	2.7	Uorsar Rupes
73	343	6053.7	6057.7	4.0	eastern Freyja Montes
-50	350	6050.6	6051.4	0.8	Lavinia Planitia
67	356	6053.7	6058.7	5.0	northern Maxwell Montes

images from the first cycle of mapping by the Magellan spacecraft spectacularly confirm these earlier interpretations of compressive structures [*Solomon et al.*, 1991]. A more complete listing of fold belts is given in Table 1. We briefly introduce the variety of fold belts on Venus based on their general tectonic settings.

Fold Belts at the Margins of Plateaus and Tesserae

Many fold belts lie at the margins of plateaus. The highest and best known folded mountain belts appear as the lightest (highest) regions on the topographic index map (Figure 3) and surround the 4-km-high plateau of Lakshmi Planum in the northern hemisphere: Maxwell Montes on the east, Freyja Montes on the north, Akna Montes on the west, and Danu Montes on the south. Also in this category are Uorsar Rupes at the margin of the Itzpopalotl tessera plateau north of Freyja Montes and others at the margins of lower plateaus and tessera highlands, including fold belts at the northern margin of Ovda Regio in the equatorial region.

Maxwell Montes fold belt contains the highest elevations of Venus, over 11 km above mean planetary radius (6051 km), and it is also the widest fold belt, exceeding 500 km. A more typical width of fold belts on Venus is about 100 km. Figure 4 is a 200-km-wide image of the western margin of Maxwell, which is composed of a 25- to 100-km-wide toe region of less intense deformation and a more intensely deformed interior. The topographic profiles show that there is generally a flexural foredeep about 50 km wide containing the less deformed toe. Just inward of the toe is a 20- to 30-km-wide steep margin to the mountain belt, in which elevations rise about 6 km before a deformed high plateau is reached. These features of western Maxwell appear in many other fold belts: (a) a flexural foredeep, (b) a less deformed toe of low surface slope (±1°), (c) a narrow zone of steep slopes (as high as 10-20°), and (d) a relatively flat top or plateau.

The 50-km-wide Uorsar Rupes fold belt north of Lakshmi Planum (Figure 5) shows a 100-km-wide foredeep generally filled with lava, a 30-km-wide toe, a 15- to 20-km-wide steep zone, and a broad relatively flat crestal massif. The interior crestal massifs do not show the folded character displayed by Maxwell, indeed they appear more like tessera [*Head*, 1990]. A number of fold belts appear at the margins of tessera plateaus or within them. One example is Semuni Dorsa (Figure 6*b*), which is also about 100 km wide. The relief of Semuni is only 2 km compared with 6 km for Maxwell (Figure 4). Figure 6*c* is a fold belt that lies within a broader region of Laima Tessera, with tessera appearing both in the foredeep and in the crestal plateau.

A few fold belts do not show the typical well-developed flexural foredeep, narrow zone of steep slope, and flat plateau. For example, the fold belt on the east side of Freyja Montes (Figure 7) shows no foredeep flexure and no well-developed

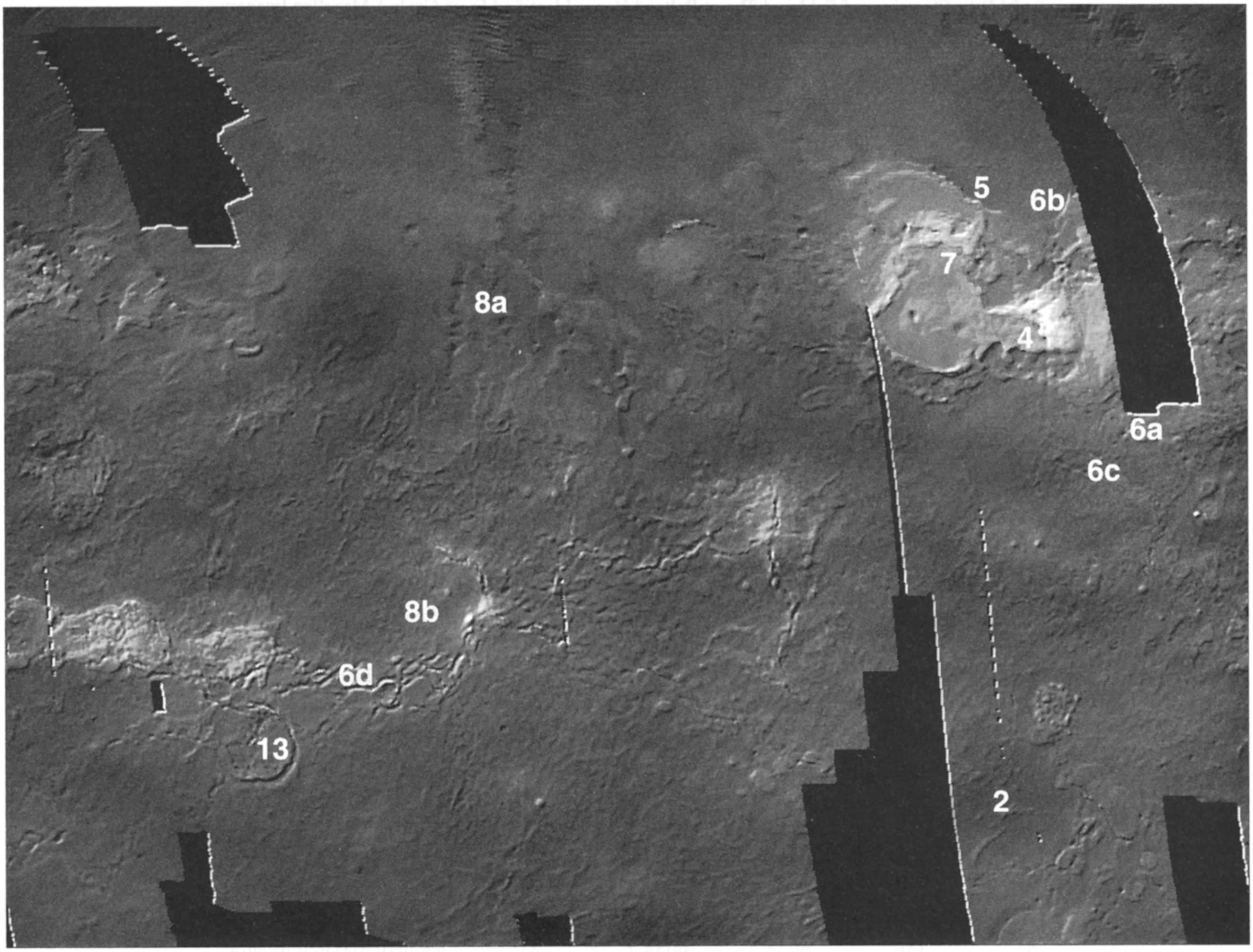

Fig. 3. Shaded relief index map of Venus based on Magellan altimetry, showing locations of key features discussed in this paper. Numbers refer to figure numbers. The highest elevations are shown in the lightest tones.

region of steep slope. The interior plateau is broken up by normal and strike-slip faulting, as discussed by *Solomon et al.* [1991]. Freyja Montes has possibly collapsed somewhat, as discussed in a later section.

Fold Belts at the Margins of Large Deformed Coronae and Chasmata

A second important setting of fold belts is at the margin of some larger deformed coronae, including Artemis Chasma (Figure 13) south of Aphrodite Terra. This 100-km-wide and 1500-km-long fold belt lies within an enormous foredeep like trough between the corona and a great outer flexural high [see *Sandwell and Schubert*, this issue]. The structure of the 50-km-wide toe of Artemis will be discussed in a later section. The inner steep zone displays clear normal faulting in some areas.

Other deformed coronae with marginal fold belts include Nightingale and various deformed coronae within the region of Diana and Dali Chasma. A detail of one 100-km-wide example is shown in Figure 6*d*.

Ridge Belts and Wrinkle Ridges of the Low Plains

The low relief ridge belts that exist within the low plains of Venus were interpreted to be the result of compression based on Venera 15 and 16 radar images [*Barsukov et al.*, 1986; *Frank and Head*, 1990]. These belts are strongly concentrated in the region between 140° and 240° longitude and from 30° to 90°N, although they are present in other low plains of Venus. Magellan radar images show these ridge belts to be composed of anastomosing folds on the average less than 10 km wide (Figure 2) within anastomosing fold belts that can be over 100 km wide and hundreds of kilometers long (Figure 8*a*). Magellan altimetry show that the ridge belts have very little relief, at most about 1 km higher than the surrounding plains (Figure 5). The ridge belts of Lavinia Planitia (Figure 2) have been described in detail from Magellan data by *Solomon et al.* [1991] and *Squyres et al.* [this issue].

Magellan synthetic aperture radar images indicate that although much of the low plains of Venus lack the pervasive deformation manifest in the mountainous regions, much of the surface shows well developed wrinkle ridges [*Solomon et al.*, 1991; *Bilotti and Suppe*, 1992]. These features are interpreted to be compressional in origin because of their similarity to wrinkle ridges analyzed on other terrestrial planets [*Plescia and Golombek*, 1986; *Watters*, 1988, *Golombek et al.*, 1991]. Figure 8*b* is a synthetic aperture radar image showing ubiquitous wrinkle ridge development within the crust of Rusalka Planitia. Although the amount of shortening being accommodated by these

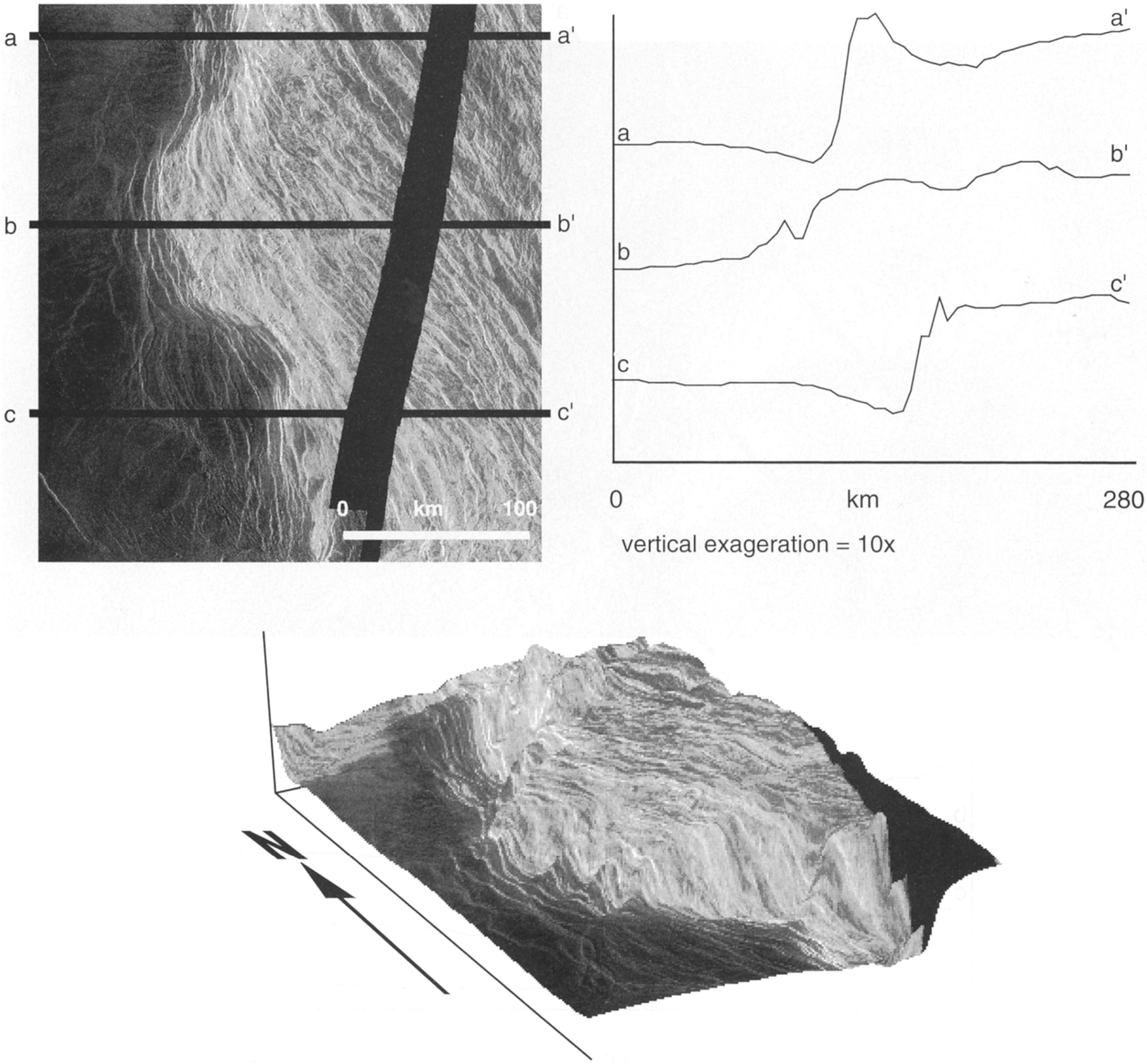

Fig. 4. Topographic profiles across Maxwell Montes (at about 64°N, 001°) showing (1) a foredeep flexure with widely spaced folds, (2) a narrow zone of steep slopes, and (3) a relatively flat crest.

structures is small, they are widely distributed over thousands of kilometers in plains regions of Venus and thus are an important expression of Venusian tectonics [*Bilotti and Suppe*, 1992].

Polydeformed Fold Belts

Fold belts from all of the tectonic settings can show later deformation under changed tectonic conditions. For example, the crest of Freyja Montes at the northern margin of Lakshmi Planum is severely deformed by a 100-km-wide zone of normal and strike-slip faulting, indicating substantial later extension and collapse. The fold belt in Figure 6*a*, which is within a broadly tesserated terrain southwest of Maxwell Montes, has been redeformed by later normal faulting, thrusting along a new trend, and an impact structure. In addition a number of fold belts have been partially flooded, particularly those at lower elevations. An example is the northern margin of Ovda Terra.

PREDICTIONS OF WEDGE THEORY FOR VENUS

Wedge theory is based on the observation that fold-and-thrust belts form by substantial horizontal compression above a basal sliding surface or zone of detachment [*Chapple*, 1978]. This deformation produces a narrowly tapered, wedge-shaped cross section above the base. The taper at any point in the wedge is simply the sum of the surface slope α and the basal dip β (Figure 1). Tapers on Earth are usually less than 10° and sometimes less than 1-2° [*Davis et al.*, 1983]. The critical wedge taper is the equilibrium shape for which the compressive force across any segment of the wedge is balanced by the resistance to translation of that part of the wedge lying in front of the segment. This balance of forces is described in detail in *Davis et al.* [1983] and *Dahlen* [1990], which are good introductions to wedge theory generally and Coulomb wedge theory specifically. Here we simply mention key points of interest to Venus.

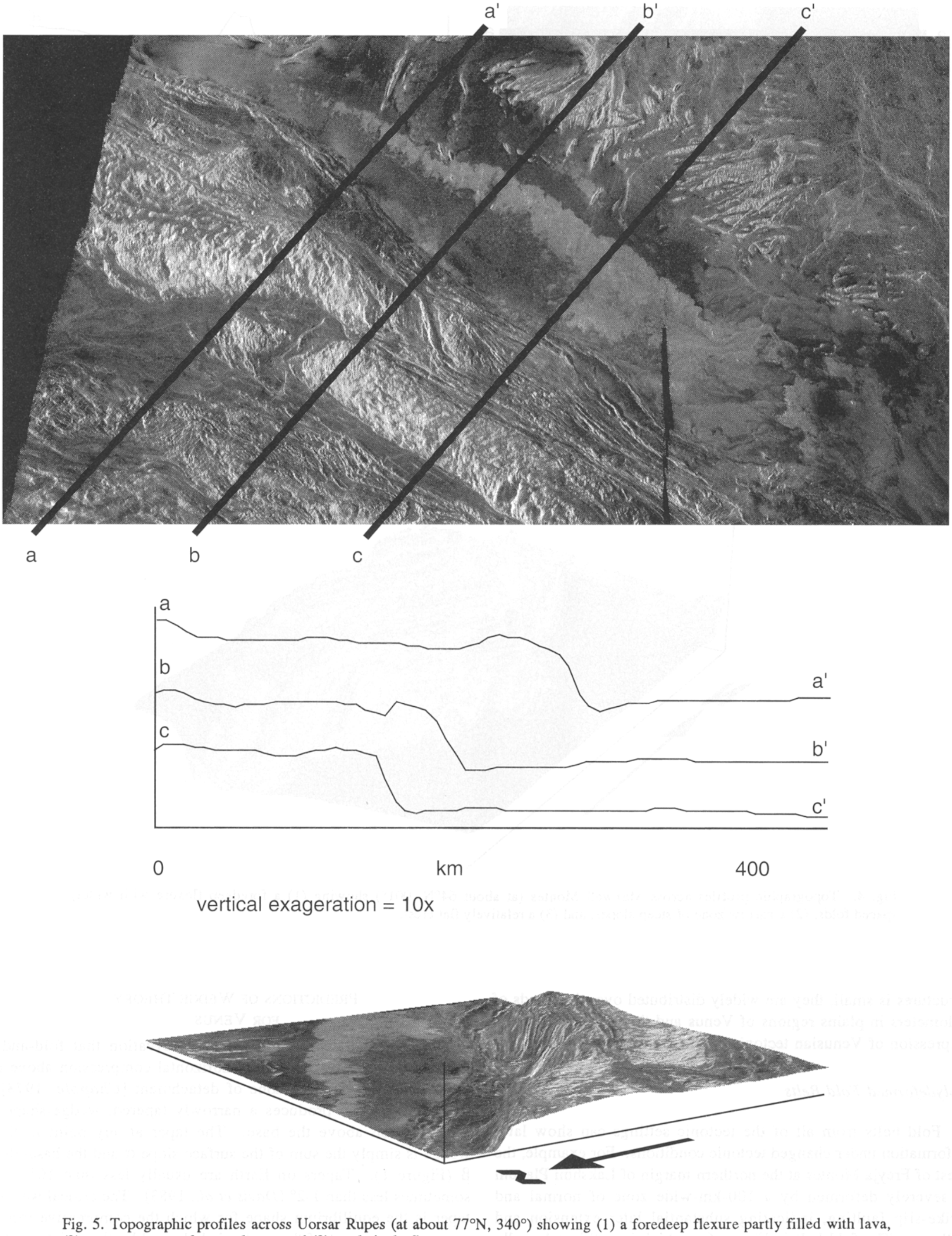

Fig. 5. Topographic profiles across Uorsar Rupes (at about 77°N, 340°) showing (1) a foredeep flexure partly filled with lava, (2) a narrow zone of steep slopes, and (3) a relatively flat crest.

An increase in the basal resistance or a weakening of the interior of the wedge produces a larger taper. For example, the two areas of lower taper on the right and left sides of Figure 1 formed at lower basal resistance or higher wedge strength or both, relative to the area of high taper in the middle. If a wedge taper were less than critical, then the wedge would deform internally until critical taper was reached, at which point the toe of the wedge could translate forward. Tapers of subaerial wedges on Earth are continually reduced by erosion causing continued deformation within the wedge; critical tapers can only be measured in subaerial wedges that are active [*Dahlen and Suppe*, 1988]. Fortunately wedges on Venus are noneroding and thus more like submarine wedges on Earth. However, noneroding wedges are more difficult to interpret because the current taper is only constrained to be greater than or equal to the current critical taper, which leads to ambiguities of interpretation. For example the steep sloped segment of the wedge in Figure 1 might have formed above a high-friction decollement but now be translating over a less resistive surface, perhaps one above which the low taper toe has formed. Alternatively the wedge could be entirely at critical taper with the low taper toe an effect of cohesive strength [*Dahlen et al.*, 1984]. In any case, wedges on Venus are expected to maintain the critical tapers at which they formed, unless they have been later deformed to some new shape or covered by lava.

Brittle Wedges on Venus

The wedge shape is controlled by actual rheologies. Therefore we must assess judiciously the proper rheologies for any modeling of Venus. Faulting is clearly evident over much of Venus on the high-resolution Magellan images. This observation suggests that rocks are commonly brittle at the surface under the tectonic conditions of Venus and at some depth the rock must undergo a transition to thermally activated plastic behavior. We assume that strength in the uppermost crust is controlled by brittle fracture or frictional sliding according to the Coulomb equations:

$$|\sigma_\tau| = S_o + (1-\lambda)\,\mu\,\sigma_n \qquad (1)$$

and

$$(1-\lambda)\,\sigma_1 = C_o + (1-\lambda)\,q\,\sigma_3 \qquad (2)$$

where σ_τ, σ_n, and $\sigma_1 \geq \sigma_3$ are the shear, normal, and principal compressive stresses; λ is the Hubbert-Rubey fluid-pressure ratio; and the material constants S_o and C_o describe the cohesion and μ and q describe the pressure dependence of strength [see *Jaeger and Cook,* 1979]. Fluid pressures, which are a major concern on Earth, can be ignored on Venus given the low density of the atmosphere (about 75 kg/m^3 [*Kliore et al.* 1985]) relative to the density of water. The hydrostatic fluid-pressure ratio $\lambda = \rho_f/\rho_r$ is just slightly greater than zero. The pressure dependence is in the range μ = 0.6-0.85 (q = 3.1-4.7) and insensitive to rock type [*Byerlee*, 1978; *Davis et al.*, 1983].

Cohesive strength S_o or C_o is commonly not of great importance in fold-and-thrust belts on Earth because cohesion increases with depth in sedimentary basins and therefore acts like a pressure dependence [*Zhao et al.*, 1986]. However on Venus, with its basaltic or diabasic crust, we expect cohesion to be roughly constant throughout and relatively high. Diabase in small samples shows very high cohesion (e.g., S_o = 114 MPa and C_o = 487 MPa [*Brace,* 1964]), but here we are speaking of the large-scale in situ cohesive strength of the crust which will be composed of lava flows, sills, dikes, and faults and joints in nonoptimal orientation [see *Dahlen et al.*, 1984]. We guess a cohesive strength of about S_o = 25 MPa. It should be noted that brittle rock strength parameters are relatively insensitive to temperature [*Paterson*, 1978].

The effect of cohesion is to cause brittle wedges to have an upper boundary layer in which the strength is dominated by cohesion and stresses are oriented parallel to the surface slope. The approximate thickness h of the boundary layer is of the order [*Dahlen et al.*, 1984]

$$h = S_o / \mu \rho_r g \qquad (3)$$

where g is the gravitational acceleration. We estimate the cohesive boundary layer h to be roughly 1-2 km thick on Venus (g = 8.87 m/s^2). Therefore we expect cohesion to play a significant role near the toes of fold-and-thrust belts that have a decollement depth of a few kilometers or less. The effect of high constant cohesion on brittle wedge taper is to produce a low-taper toe but a higher-taper interior (compare Figure 1) that is close to the taper of a cohesionless wedge because at large wedge thickness the pressure-dependent term of (1) or (2) dominates [*Dahlen et al.*, 1984]. Thus the maximum surface slope of a wedge can be simply computed using non cohesive wedge theory.

The exact critical taper equation for a homogeneous non cohesive Coulomb wedge [*Dahlen*, 1984] is

$$\alpha + \beta = \psi_b - \psi_o \qquad (4)$$

where α is the topographic slope, β is the decollement dip, ψ_b is the angle between the maximum compression σ_1 and the base of the wedge, and ψ_o is the angle between σ_1 and the surface (see Figure 1). Equation (4) is simply a trigonometric identity, but ψ_b and ψ_o are functions of rock properties within the wedge and on the base:

$$\psi_b = \frac{1}{2}\sin^{-1}\left[\frac{\sin\phi_b}{\sin\phi}\right] - \frac{1}{2}\phi_b \qquad (5)$$

and

$$\psi_o = \frac{1}{2}\sin^{-1}\left[\frac{\sin\alpha}{\sin\phi}\right] - \frac{1}{2}\alpha \qquad (6)$$

where $\phi = \tan^{-1}\mu$ within the wedge and $\phi_b = \tan^{-1}\mu_b$ on the basal decollement. Reasonable values of these parameters (μ = 1.1-1.2 μ_b for Earth wedges [*Davis et al.* 1983]) indicate that maximum surface slopes should be in the range 10° to 20° for brittle wedges on Venus if the basal dip β is no more than 5-10°, as shown in Figure 9.

Effect of Brittle-Plastic Transition on Wedge Shape

We have concluded above that deformation in the Venusian lithosphere, like the Earth's, is governed by brittle mechanisms near the surface. However because of the high surface temperature (470° C at mean planetary radius; *Kliore et al.*, [1985]) we expect to see a transition to thermally activated plastic mechanisms at a relatively shallow depth. These mechanisms are strongly temperature, strain-rate and material-dependent [for example, *Kirby,* 1983]; for simplicity we adopt the power-law creep constitutive equation

$$\dot{\varepsilon} = A\,[\sigma_1 - \sigma_3]^n \, \exp\left[\frac{-Q}{R\,T}\right] \qquad (7)$$

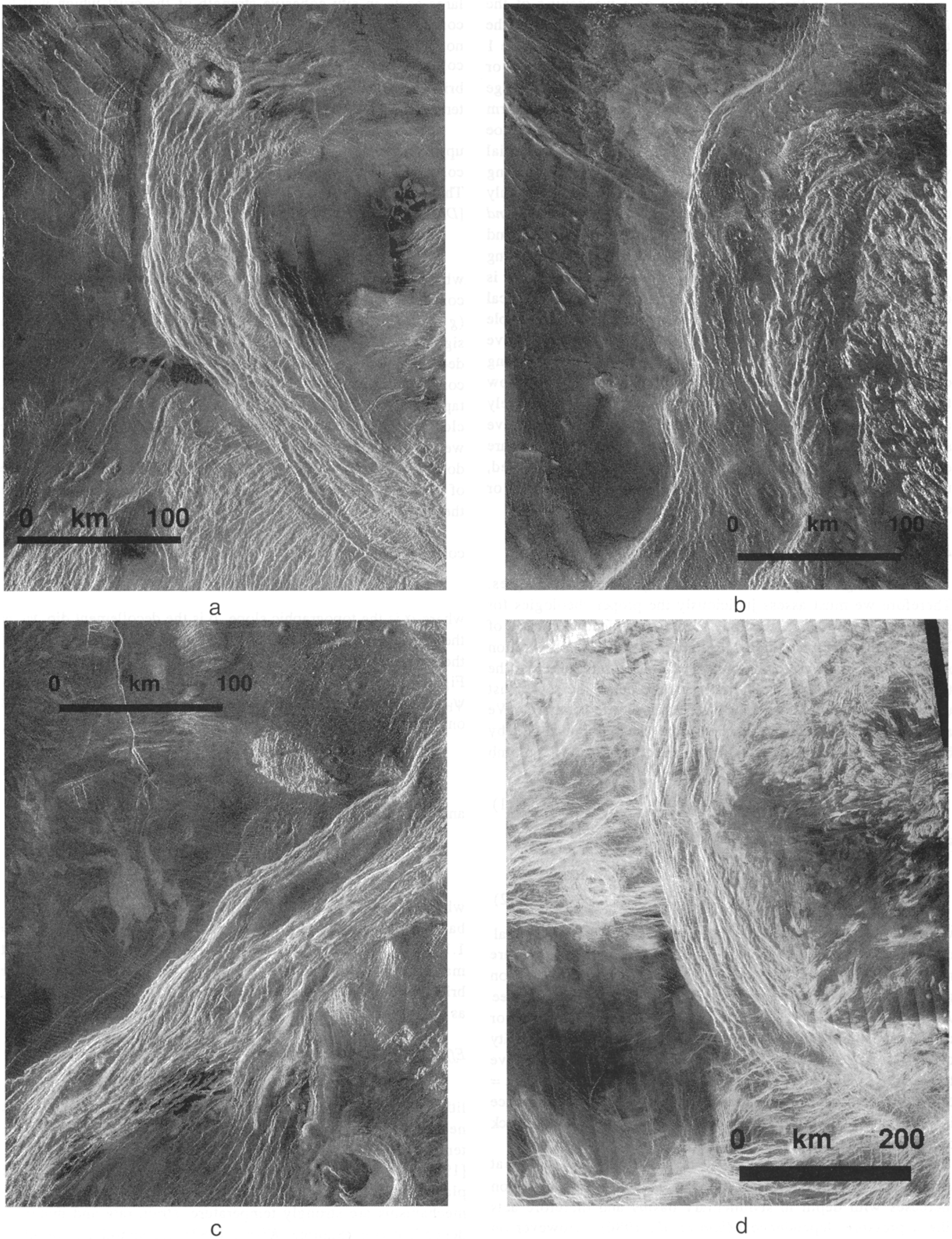

Fig. 6. Fold belts of Venus. (*a*) A polydeformed fold belt near Laima Tessera at about 58°N, 027°. (*b*) Semuni Dorsa located at about 77°N, 010°. (*c*) A fold belt near Laima Tessera at about 26°N, 048°. (*d*) A fold belt near Diana and Dali Chasma at 17°S, 156°.

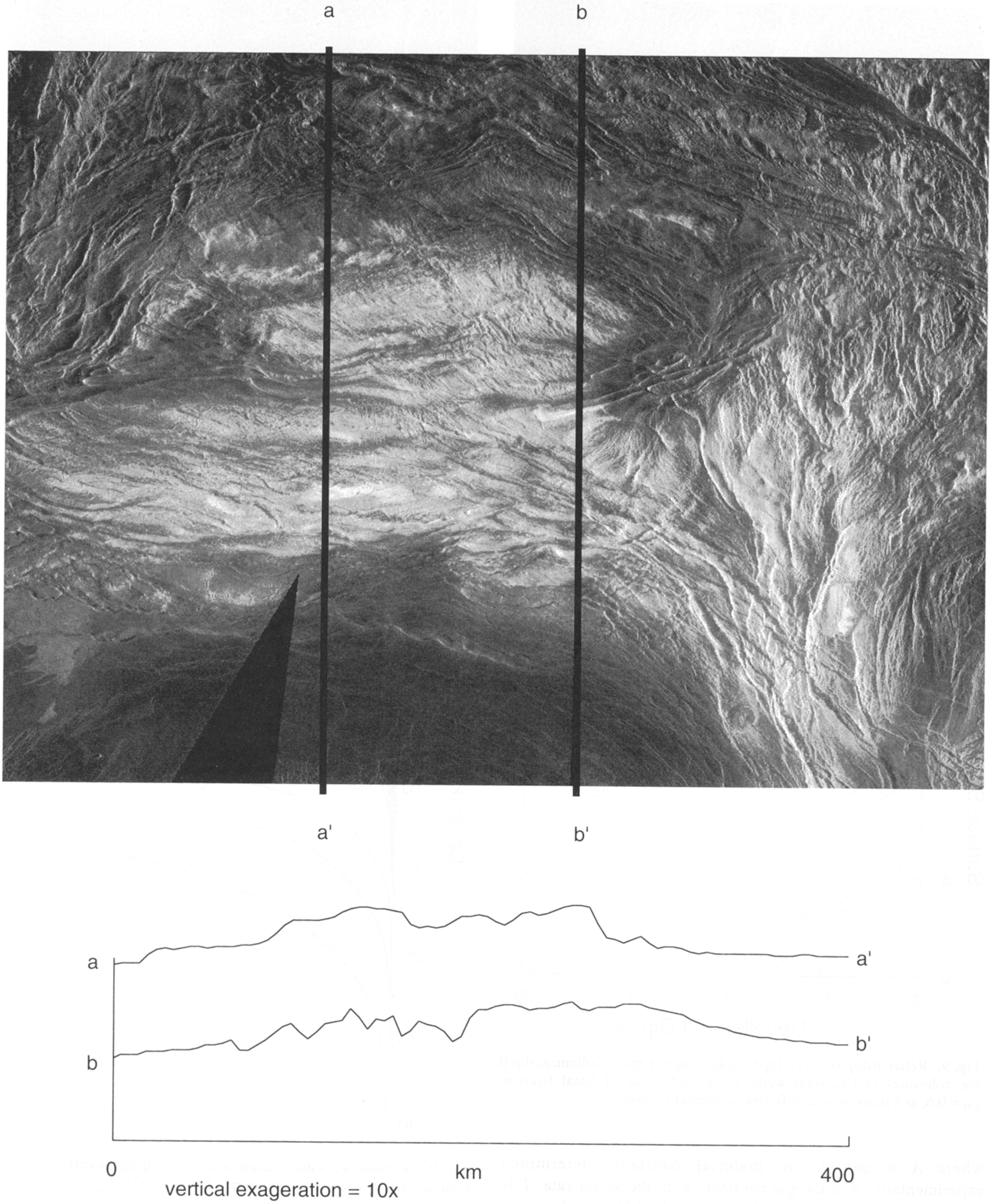

Fig. 7. Topographic profiles across Freyja Montes (at about 74°N, 330°) showing, in contrast with Figures 4 and 5, (1) a lack of a foredeep flexure, (2) a less well-developed zone of steep slope, and (3) a less well-developed high plateau with normal faulting.

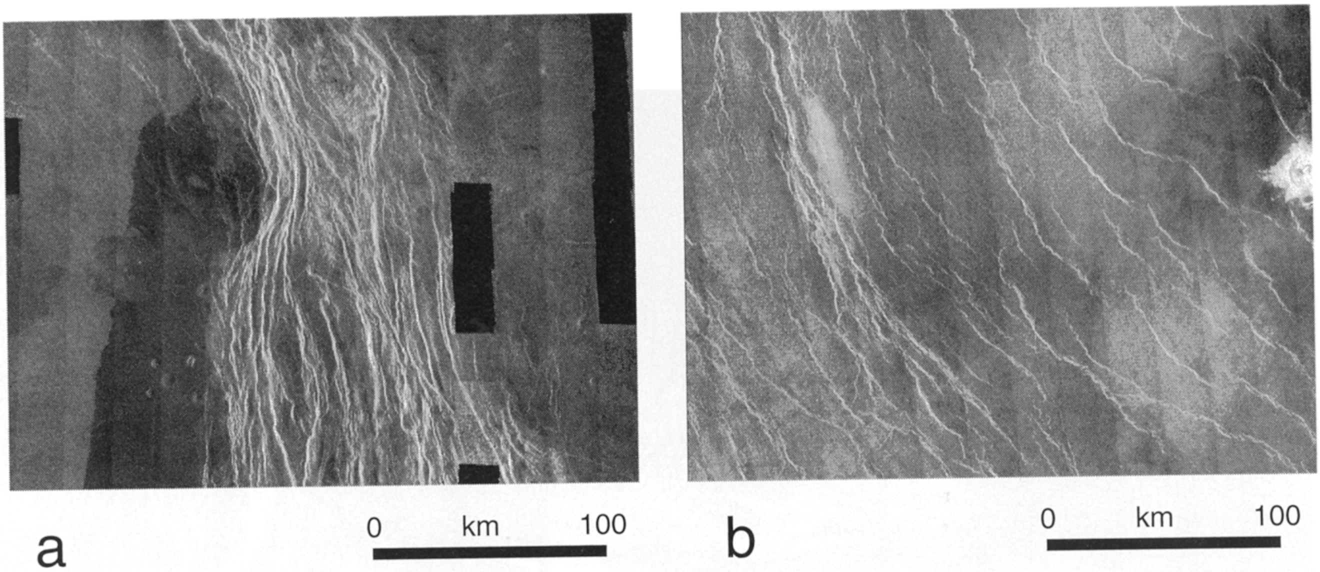

Fig. 8. Fold belts within the low plains of Venus. (*a*) A typical ridge belt located at 44°N, 164°. (*b*) Wrinkle ridge development, typical of the low plains of Venus. Located at 05°N, 177°.

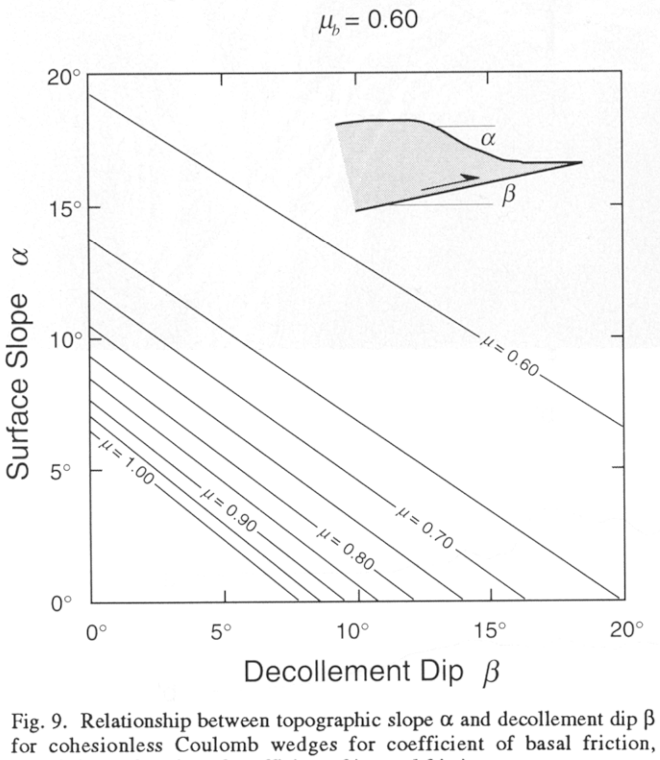

Fig. 9. Relationship between topographic slope α and decollement dip β for cohesionless Coulomb wedges for coefficient of basal friction, $\mu_b = 0.6$, as a function of coefficient of internal friction, μ.

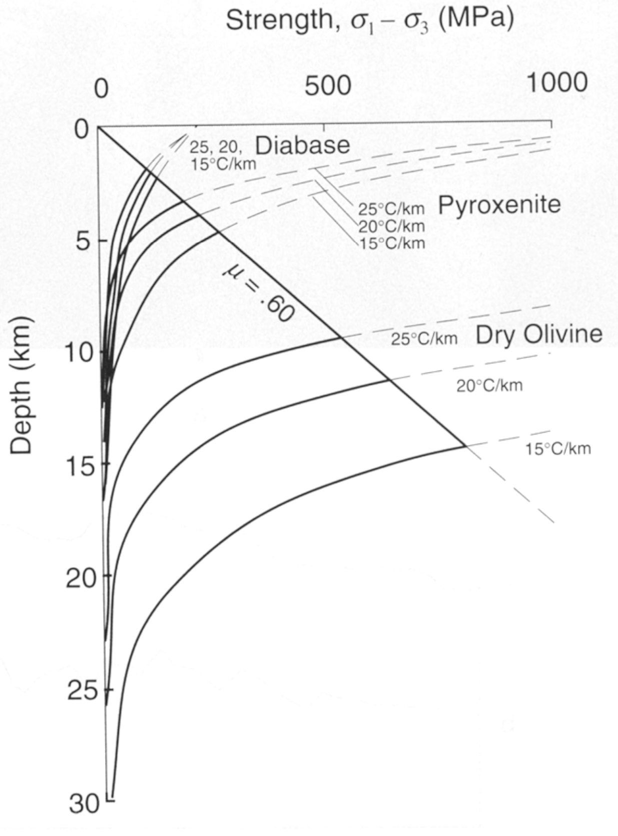

Fig. 10. Nominal laboratory plastic and brittle strengths extrapolated to presumed Venus conditions, strain rate $\dot{\varepsilon} = 10^{-14} s^{-1}$, $T_{surface} = 470°C$.

where A, n, and Q are material constants determined experimentally, R is the gas constant, $\dot{\varepsilon}$ is the strain rate, T is the absolute temperature, and $(\sigma_1 - \sigma_3)$ is differential stress. Using values of A, n, and Q from rock mechanics experiments [*Shelton and Tullis*, 1981; *Kirby*, 1983], we estimate plastic strengths as a function of depth for diabase, pyroxenite, and dunite, based on consideration of possible strain rates, geothermal gradients, and surface temperatures. We should emphasize that this extrapolation to Venusian conditions has substantial uncertainties; we pursue this simply in the spirit of obtaining an initial indication of what might be expected for mechanics of fold-and-thrust belts under conditions that are very different from Earth.

Diabase is a reasonable crustal material based on data from the Venera landers, indicating a tholeiitic composition for surface rocks at several sites [*Surkov et al.*, 1983]. Pyroxenite

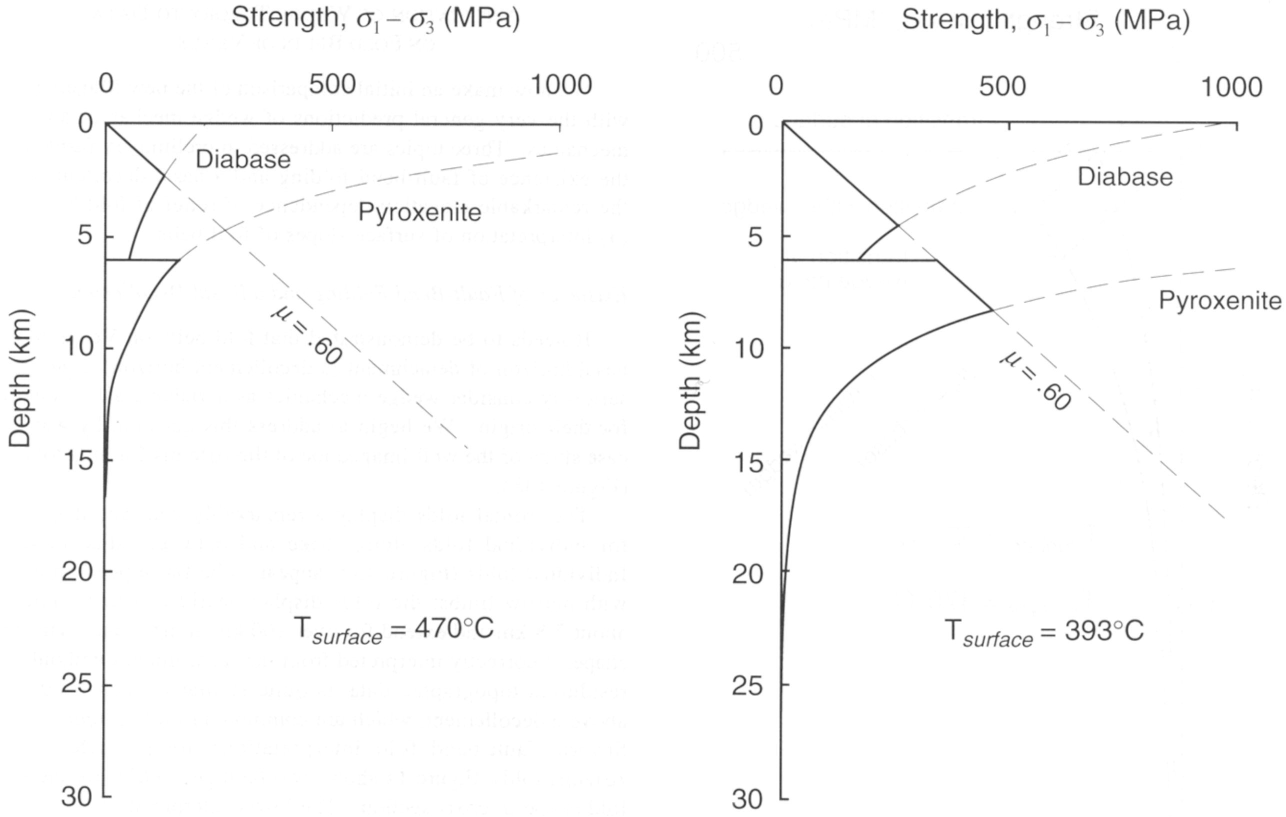

Fig. 11. An example of the effect of atmospheric temperature on depth of brittle plastic transition, strain rate $\dot{\varepsilon} = 10^{-14}s^{-1}$, geothermal gradient dT/dz = 15°C/km. The atmospheric temperature at mean planetary radius is 470°C and 393°C at +10 km.

and dunite may give reasonable estimates of mantle strength, pyroxenite being a proxy for peridotite. There is very little water vapor in the Venusian atmosphere [*Kaula*, 1990]. Therefore we should be concerned about the dryness of the laboratory measurements in light of the importance of water in diffusion in silicates; rock strengths in the experiments were reported to be strongly dependent on water content [*Shelton and Tullis*, 1981]. Therefore we consider the plastic strengths to be possibly lower bounds. We use the commonly quoted Earth value of $10^{-14}s^{-1}$ for strain rate $\dot{\varepsilon}$, recognizing that distributed strain rates on Earth lie in the range 10^{-13} to $10^{-15}s^{-1}$ [*Pfiffner and Ramsay*, 1982] and that strain rates in a basal decollement entering the brittle-plastic transition could be several orders of magnitude higher. Typical geothermal gradients away from volcanic centers and active coronae might be 15°-25°C/km; for example, *Solomon and Head* [1990] analyzed the flexure of the Venusian lithosphere in the Uorsar Rupes foredeep to determine the effective elastic thickness and estimated thermal gradients similar to terrestrial values.

The computed strengths as a function of depth for a variety of geothermal gradients are given in Figure 10 with a cohesionless brittle-strength line $\mu = 0.6$ for reference. Thus the predicted depth of a nominal brittle-plastic transition along a decollement would be 2-3 km for diabase, 4-5 km for pyroxenite, and 10-15 km for dunite. These values are similar to a number of previous estimates for Venus [e.g., *Solomon and Head*, 1984; *Zuber and Parmentier*, 1987; *Phillips*, 1990; *Fletcher*, 1989]. However, wedge mechanics on Venus require some additional considerations, including the effects of variable surface temperature and the effects of wedge cohesion.

The atmosphere of Venus is relatively stable and has a 8°C/km vertical temperature gradient [*Kliore et al.*, 1985; *Lyons*, 1991]. The surface temperature is only 393°C at an elevation of 10 km, whereas it is 470°C at mean planetary radius and 490°C at the lowest elevations. These atmospheric temperature differences affect the strength of the uppermost lithosphere as is illustrated in Figure 11, which models the uppermost lithosphere as a 6-km-thick crust of diabase overlying the pyroxenite proxy for peridotite. At mean planetary radius we predict for this model a 10- to 15-km thick "jelly sandwich" structure, analogous to the common strength-depth relations for continental crust on Earth [e.g., *Suppe*, 1985]. Note that only the diabasic crust is brittle. In contrast, at 10 km elevation the uppermost lithosphere is vastly stronger with both the diabase and pyroxenite showing an interval of brittle behavior and almost no weak middle zone of plastic diabase.

It should be emphasized that some of the above conclusions are highly model dependent. In particular there are large uncertainties in the plastic strengths and in the thickness of the diabasic layer, which probably varies substantially over the planet and within mountain belts specifically. However, the dependence of the strength and thickness of the strong uppermost lithosphere on atmospheric temperature seems fairly robust over a broad range of parameters. The key observations are that there is a 100°C change in atmospheric temperature over the 12 km range of planetary elevation, that rock is brittle at the surface and that the brittle-plastic transition is probably within 10-15 km of the surface; therefore the transition will be elevation dependent.

So far we have only considered the brittle-plastic transition for a cohesionless decollement; the brittle-plastic transition

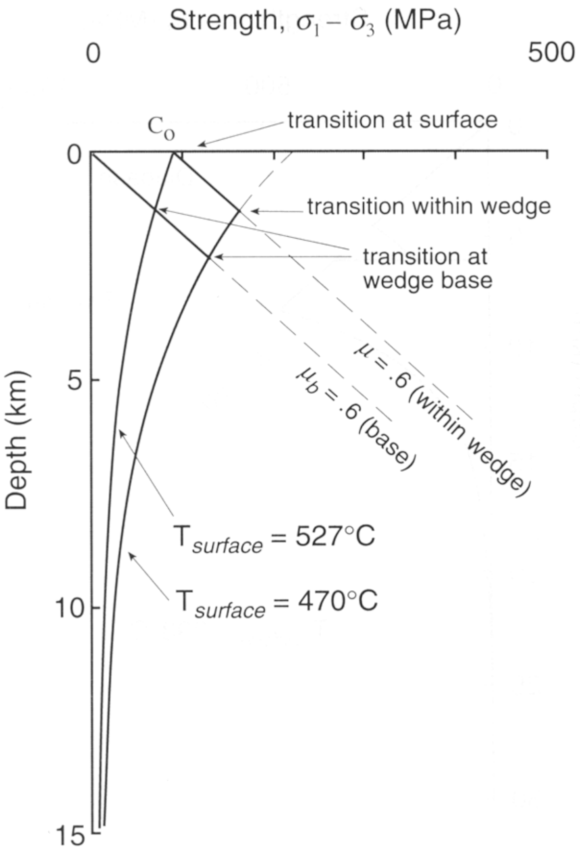

Fig. 12. Near-surface brittle-plastic transitions within a cohesive wedge and along a cohesionless decollement, strain rate $\dot{\varepsilon} = 10^{-14}s^{-1}$, geothermal gradient dT/dz = 15°C/km.

within a cohesive wedge is different as is illustrated in Figure 12. Because of finite cohesive strength, the brittle-plastic transition within the wedge will always have a shallower depth than the transition on the basal decollement. Thus even at the lowest planetary elevations there will still be a plastic lower segment to a cohesive wedge. Only when the plastic strength at the surface is less than or equal to the cohesive strength C_0 (Figure 12) is the brittle-plastic transition truly at the planetary surface.

The main effect of the brittle-plastic transition in critical taper wedge mechanics is the strength drop on the basal decollement which causes a reduction in taper ($\alpha+\beta$) and surface slope α. This prediction is verified on Earth by the low tapers (1°) of fold-and-thrust belts with plastic salt as the basal decollement [*Davis and Engelder*, 1985]. If a transition from brittle to plastic deformation takes place along a basal decollement interior to the toe of the wedge, then a relatively abrupt flattening of the surface slope of the wedge is predicted above the plastic decollement, as shown qualitatively in Figure 1. It has been suggested that the decrease in slope of accretionary wedges across the trench-slope break, the decrease in slope of the Himalayas into the Tibetan Plateau and the slope of the Andes into the Altiplano are all surface expressions of the brittle-plastic transition on the decollement, generally in the 10-15 km range on Earth in the case of quartz-rich rocks [*Davis et al.*, 1983; *Dahlen*, 1990]. If brittle wedges exist on Venus, we would expect a qualitatively similar topographic form reflecting the brittle-plastic transition on the basal decollement.

APPLICATION OF WEDGE THEORY TO DATA ON FOLD BELTS OF VENUS

We now make an initial comparison of the new Magellan data with the very general predictions of wedge mechanics and rock mechanics. Three topics are addressed in preliminary fashion: (1) the existence of fault-bend folding and a basal decollement, (2) the remarkable elevation dependence of relief of fold belts, and (3) interpretation of surface slopes of fold belts.

Existence of Fault-Bend Folding and a Basal Decollement

It needs to be demonstrated that fold belts on Venus have a basal horizon of detachment (a decollement horizon) if we are to seriously consider wedge mechanics as a viable class of theories for their origin. We begin to address this question by way of a case study of the well-imaged toe of the Artemis Chasma fold belt (Figure 13*b*).

The frontal folds display a remarkably constant shape both for individual folds along strike and between adjacent folds. Individual folds (Figure 13*c*) appear to be flat topped anticlines with narrow limbs; the folds display nearly constant widths of about 7-8 km and extend for over 100 km along strike. This fold shape, if correctly interpreted from the radar images without high resolution topographic data, is quite similar to fault-bend folds above a decollement, which are common on Earth [*Suppe*, 1983]. Several fault-bend fold interpretations are possible for the Artemis folds; Figure 14 shows two fault geometries for the same fold shape in cross section. The lower interpretation has a short fault ramp and large slip, the fault slip being the distance between the first and third axial surface. In contrast, the upper interpretation has a long fault ramp and short slip, the fault slip being roughly the limb width. These two interpretations are not easily distinguished with a topographic cross section alone, but the along-strike termination of the folds may allow these solutions to be distinguished, as shown in the three-dimensional model of Figure 14. Wherever a fold has a constant width along strike, the fault slip must be constant. Wherever the fault slip varies the fold width must vary. In particular, where the fault slip goes to zero at the lateral termination of the fold, the fold must narrow to approximately the ramp width. It appears that the frontal Artemis folds terminate with a nearly constant width, suggesting that the short-slip solution is correct. The low slip solution is also suggested mechanically because the very weak upper decollement that is required by the large-slip solution seems unlikely for Venus. However, the stereo coverage and higher resolution imagery of the extended Magellan Mission is needed to further test this interpretation.

It should also be noted on the radar images that the large folds converge southward. Where adjacent folds begin to impinge, small secondary folds begin to appear in the synclines. This is a widespread phenomenon on Earth associated with interference of adjacent folds [e.g., *Mount*, 1989; *Mount et al.*, 1990].

We can determine the depth to the decollement from the fold width (compare Figure 14). The biggest uncertainty is the dip of the ramp, which can be determined directly from topographic limb dip if it is not modified by mass wasting. Preliminary modeling of left-looking images of cycle 1 in comparison with right-looking cycle 2 images suggests a topographic back-limb dip of about 13° and a steep front limb, perhaps a scree slope. Fault step-up angles of about 13° are common on Earth, for example, in Taiwan [*Dahlen et al.*, 1984]. The steep front dip suggests the fault exits to the surface in a scree slope. The fold widths of about 7-8 km combined with the possible 13° ramp

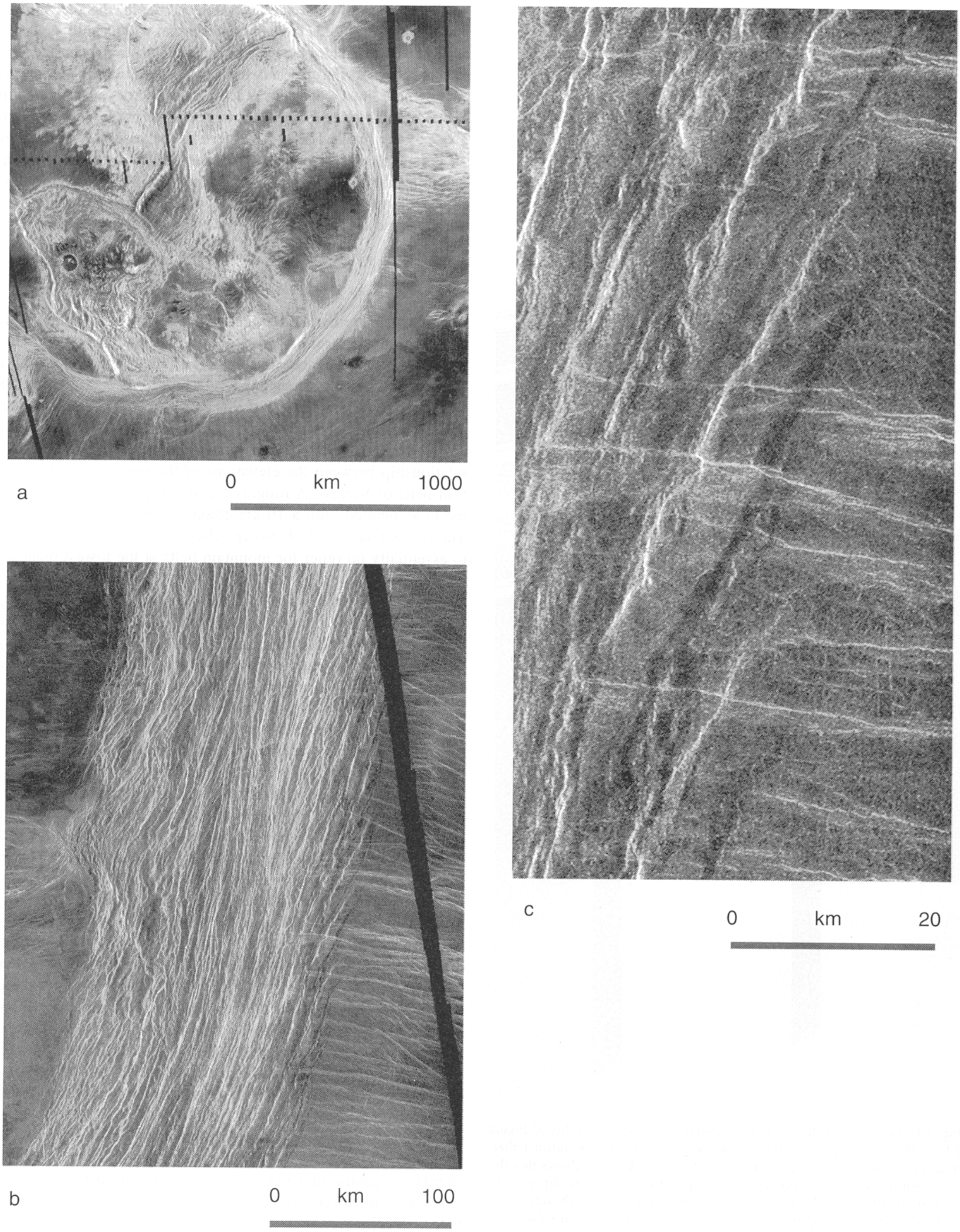

Fig. 13. Mosaic of images of the Artemis Chasma fold belt. (*a*) A view of most of the corona centered around 32°S 135°. (*b*) A detail of the eastern edge of Artemis Chasma centered around 36°S 143°. (*c*) A detail of the frontal folds centered around 36.2°S, 144.3°.

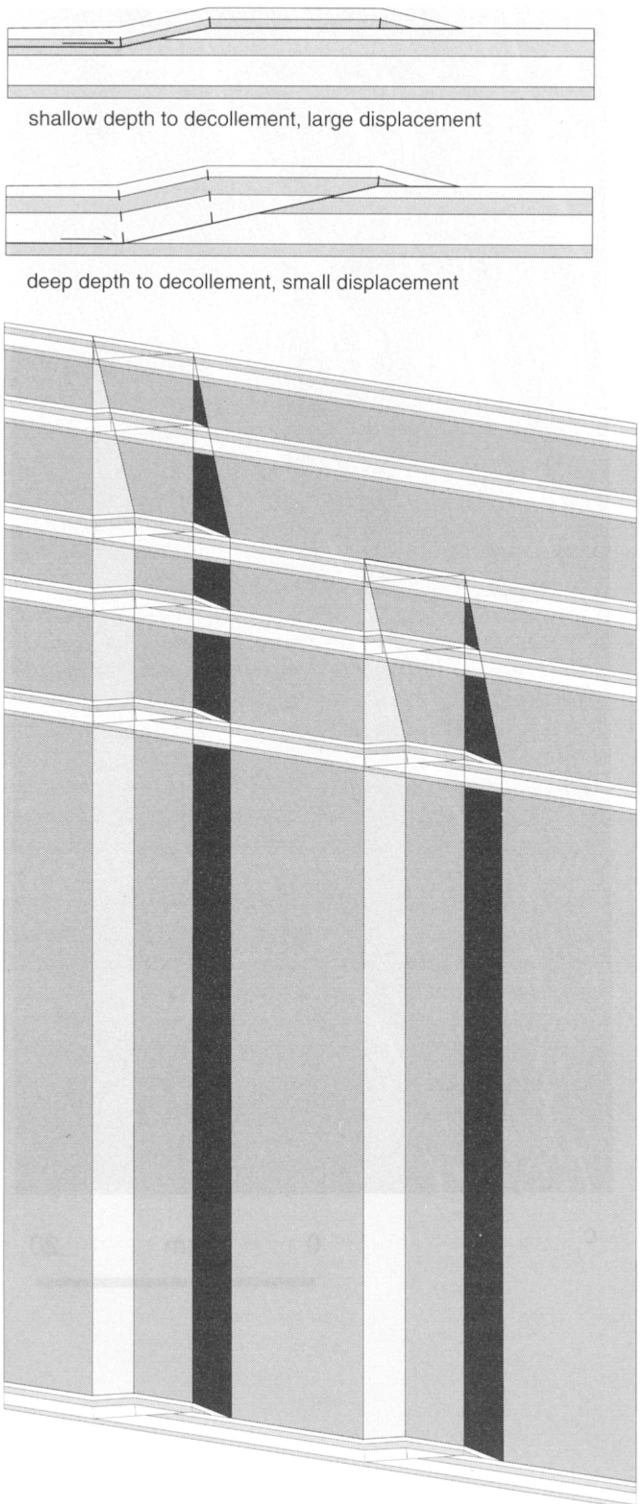

Fig. 14. Fault-bend folding models similar to the Artemis folds of Figure 13*c*. The two cross sections show two possible interpretations that produce the same topographic profile. The block diagram shows that the two solutions can be distinguished in map view, because as slip goes to zero along strike, the fold width goes to the ramp width. Therefore the fold terminations in Figure 13*c* suggest the short-displacement solution.

angles suggests a decollement depth of about 1.5 km. Refined analysis of these structures will be possible with the higher-resolution imagery and stereo topography of the extended Magellan Mission.

Regardless of the exact decollement depth, the fact that the major fault-bend folds have nearly constant widths both along and across strike indicates a nearly constant depth to the decollement. In the area of Figure 13*b* and 13*c* a regionally extensive decollement must extend for at least 50 km across strike and more than 350 km along strike. The shortening is not less than 15 km, given by the limb widths. Therefore the necessary conditions for wedge mechanics appear to be met in the toe region of the Artemis fold belt. Detailed analysis of high-resolution images and topography of folds at the toes of other fold belts is needed to determine how truly widespread decollement tectonics is on Venus.

Relief and Elevation of Fold Belts of Venus

One of the remarkable features of Venus is that the fold belts with the highest relief stand at the highest elevations. For example Maxwell Montes has a relief of about 6 km with a toe at about 4 km, Uorsar Rupes has a relief of 3 km with a toe at 500 m, and ridge belts in the low plains at or below the mean planetary radius have only about 500 m of relief. (Figures 4 and 5). Figure 15 (see also Table 1) shows that there is a nearly linear relationship between the elevations of the toes and crests of the fold belts of Venus. A roughly 5-km change in elevation of the toe is associated with a 10-km change in elevation of the crest. The linear relationship between relief and elevation extrapolates to essentially no relief for mountain belts at the lowest planetary elevations.

A few mountain belts deviate from this linear relationship, showing less relief than the linear trend. The principal mountain belt to deviate substantially from the linear relationship is Freyja Montes (shown as open squares in Figure 15). The lower relief of Freyja Montes is expected because it contains a wide zone of normal and strike-slip faulting within its crest and lacks a foredeep flexure, as discussed earlier (Figure 7). Therefore Freyja Montes seems to have collapsed from an originally higher elevation.

This remarkable linear relationship between relief and elevation is unlike anything seen on Earth but within the framework of wedge mechanics must reflect an elevation dependence to the depth to the brittle-plastic transition (compare Figure 1). A very simple and straightforward explanation seems required to produce such a strikingly linear global relationship. Two possible explanations involve (1) the atmospheric temperature gradient and (2) isostasy. (1) An elevation dependence is expected from the 8°/km atmospheric temperature gradient, as discussed above (Figure 11). However, preliminary wedge modeling suggests that atmospheric temperature gradient can account for no more than 10-20% of the elevation dependence [*Price et al.*, 1992; *Price*, 1992]. (2) A positive isostatic coupling of elevation and lithospheric thickness gives a decreasing geothermal gradient with increasing elevation (Figure 16). This requires a positively buoyant lithosphere on Venus, in contrast with negatively buoyant oceanic lithosphere on the Earth. Preliminary wedge modeling suggests that a factor of two change in geothermal gradient may be required to satisfy the observed elevation dependence of relief [*Price et al.*, 1992; *Price*, 1992].

Interpretation of Topographic Profiles of Fold Belts

Surface slope α is the one parameter of wedge theory that is directly observable on Venus and is the key observation to model. However significant uncertainties in surface slope exist,

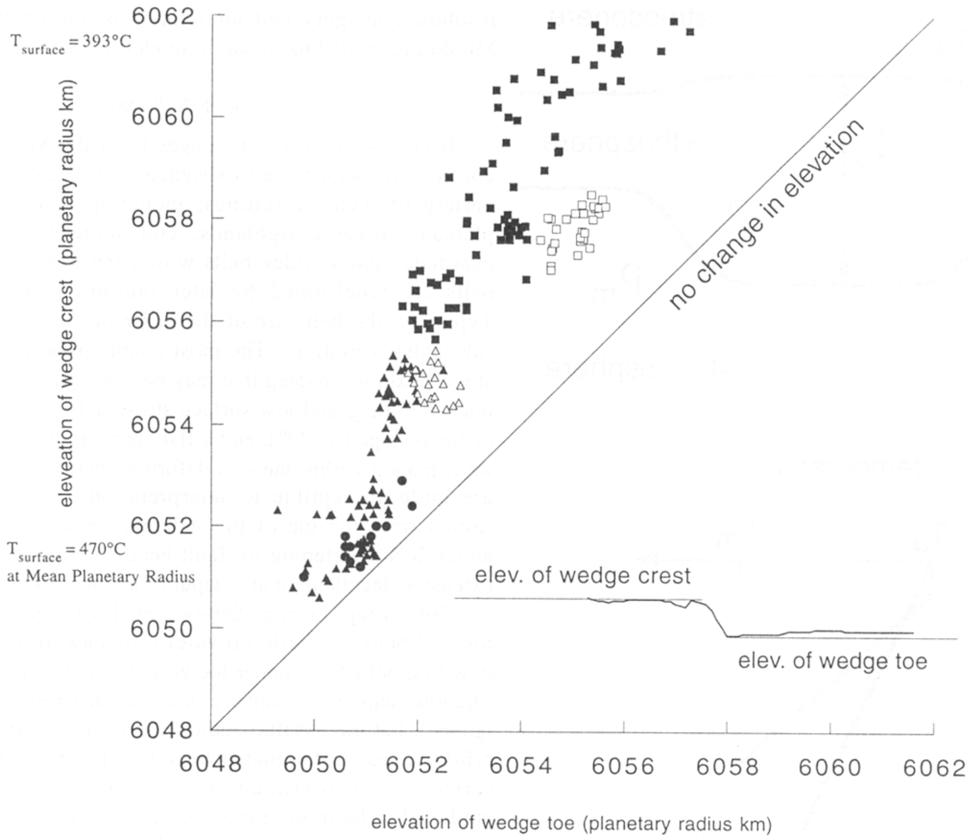

Fig. 15. Observed relationship between elevations of toes and crests of fold belts on Venus. Note that the relief of fold belts (i.e., the difference between the elevation of the toe and the crest) is a maximum at the highest elevations and decreases linearly to near zero at the lowest elevations. Most fold-and-thrust belts display a characteristic altimetric profile including a flexural foredeep and low-taper toe, a narrow zone of steep slopes and a relatively flat crest or plateau (e.g., Figures 4 and 5); these are plotted as solid symbols. Some fold-and-thrust belts do not display the characteristic flexural foredeep and have less relief than the linear trend (e.g., Freyja Montes, Figure 7); these are plotted as open symbols. Squares represent profiles (spaced at 23 km along strike) of mountain belts surrounding Lakshmi Planum. Triangles represent profiles (spaced at 46 km along strike) of fold-and-thrust belts at the margins of tessera plateaus. Circles represent average elevation of wedge crest and wedge toe for fold-and-thrust belts in the low plains.

especially in the 10- to 30 km-wide zones of steep slope (Figures 4 and 5), because of the large radar-altimetry footprint (10 to 20-km) and large spacing of the footprints in many areas. It will take considerable effort during the extended Magellan Mission to obtain the best possible data on the topography of the fold belts, particularly in the regions of steep slope. Nevertheless the 5 km gridded altimetry used to generate the topographic profiles and the oblique views of Maxwell Montes and Uorsar Rupes (Figures 4, 5, and 7) represents an enormous step forward in our knowledge of the topography of the fold belts of Venus and is sufficient for our present analysis.

Many of the fold belts exhibit a low-taper toe, indicated by the wide spacing of folds, indicating a small amount of shortening, and by the small change in surface slope at the deformation front. There are two known ways to make a low-taper toe: (1) a weak base or (2) a strong cohesive wedge, or both [*Dahlen et al.*, 1984]. In the case of the toe of the Artemis Chasma wedge (Figures 13), the low, negative taper ($\alpha = -1°$) is consistent with a cohesive wedge above a plastic base. Furthermore, the 1.5 km depth to the decollement (Figure 14) is consistent with the uncertain predicted depths to the brittle-plastic transition in diabase (Figures 10 and 11). However, it should be emphasized that a low-taper toe is also qualitatively consistent with a more resistant base and a cohesion dominated toe [*Dahlen et al.*, 1984]. The 1.5 km depth to the decollement is consistent with the 1-2 km thickness of the cohesive boundary layer, estimated above. The distinction of these two possibilities requires modeling of entire wedges, which is beyond the scope of this paper.

The narrow steep-sloped regions interior to the toe are particularly difficult to interpret because of difficulties in imaging the structures and in measuring the slopes. Based on the gridded altimetry, the steep-sloped regions may have slopes in the range of 5°-20°, which are quite similar to the maximum slopes attainable in a Coulomb wedge (Figure 9) or in the interior of a cohesion dominated wedge, as discussed above. However, steep slopes are also possible in plastic or viscous wedges, for example, *Vorder Bruegge and Fletcher* [1990] modeled the Venera altimetry of Uorsar Rupes as a viscous wedge. Preliminary modeling involving crustal strength of the form shown in Figure 12 produces an overall topography quite similar to the Venus wedges with the narrow steep-sloped region lying above the region in which the basal decollement is still brittle but the lower part of the wedge interior is plastic [*Price et al.*, 1992; *Price*, 1992].

The flattening of the interiors of most fold belts on Venus (Figures 4 and 5) is qualitatively similar to the flattening of wedges on Earth above the brittle-plastic transition. This

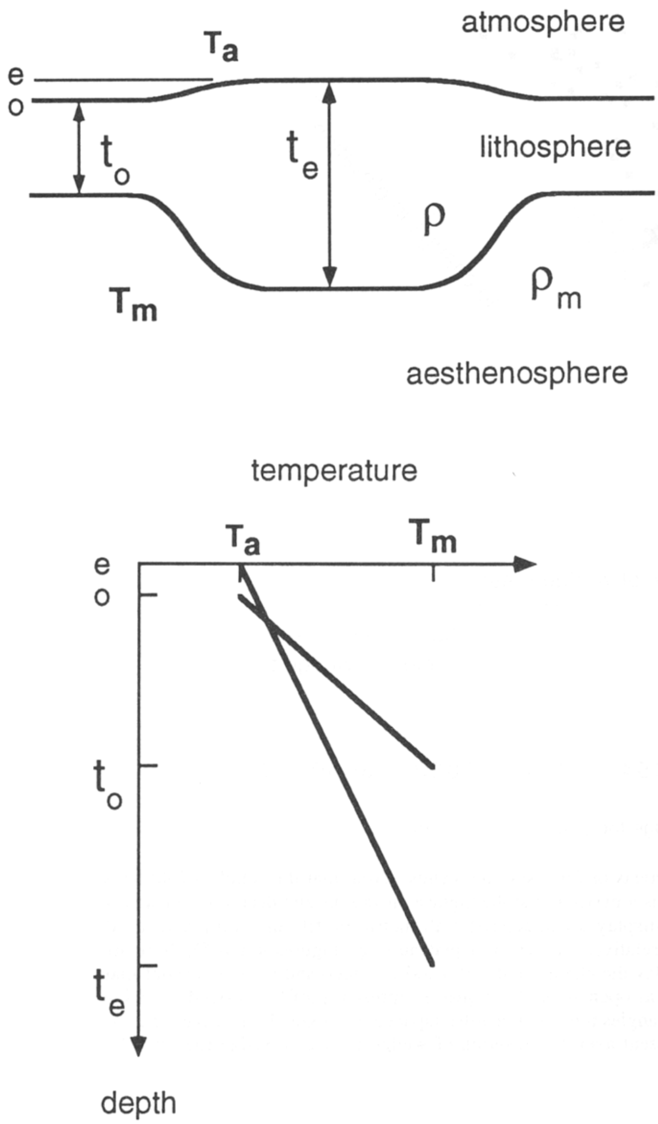

Fig. 16. Isostatic coupling of surface elevation *e* and geothermal gradient in the lithosphere with positive buoyancy $\rho < \rho_m$, where ρ and ρ_m are the densities of the lithosphere and mantle aesthenosphere. The temperatures of the atmosphere and mantle aesthenosphere are T_a and T_m.

suggests that it may be possible to model the entire toe and steep sloped regions of these wedges as having a brittle base that goes plastic under the flat crest of the mountain belt. Preliminary modeling agrees with this conclusion [*Price et al.*, 1992; *Price*, 1992]. If this proves correct, then the relationship between elevation and relief of wedges (Figure 15) is controlled by variation in the depth of the brittle-plastic transition on the decollement with elevation.

A number of structural/mechanical scenarios are conceivable for the fold belts of Venus within the framework of wedge mechanics. Some may form by gravity sliding, similar to the gravity-driven fold belts of the Gulf of Mexico [e.g., *Mount,* 1989]. This seems impossible for some belts such as Maxwell which shows folds across the steep-sloped zone and into the crest. However, the steep inner wall of Artemis Chasma shows extensive normal faulting in some areas (Figure 13*b*); gravity sliding off the flank of the coronal uplift could be responsible for both the normal faults and fold-and-thrust belt. This possibility is mentioned simply to indicate that the highest-possible resolution imagery and topography of the extended Magellan Mission is needed to constrain models of the fold belts of Venus.

CONCLUSIONS

High-resolution radar images from the Magellan spacecraft confirm the widespread existence of fold belts on Venus in a variety of tectonic settings, including along the margins of plateaus, tessera highlands, and some larger coronae and chasmata, and as ridge belts within the low plains. Some fold belts are redeformed by later faulting in new orientations. Typically, the belts are of the order of 100 km wide, but with substantial variation. The most common morphologic elements are: a flexural foredeep that may be filled with lavas, a toe of less intense folding and low surface slope, a 10- to 30-km-wide zone of steep slope (5°-20°), and a flat crest. Individual structures are well imaged within the less deformed parts of the fold belts and are quite susceptible to interpretation. For example, initial results from the toe of the Artemis Chasma fold belt indicate about 20% shortening by fault-bend folding above a regionally extensive decollement at a depth of about 1.5 km.

Critical taper wedge theory, which has been quite successful for fold belts on Earth, provides a framework for understanding how fold belts form under the very different conditions of Venus. The low-density atmosphere indicates that fluid pressures can be ignored, but the basaltic composition suggests the existence of a brittle cohesion-dominated boundary layer 1-2 km thick at the surface. The maximum surface slope of brittle wedges is predicted to be in the range 10°-15°, which is similar to current estimates of some of the steep-sloped regions of the fold belts. The low-taper toes of fold belts may be an effect of the cohesive boundary layer or alternatively of a low-strength plastic decollement. If the toe and steep-sloped regions are effects of brittle behavior, then the flat top of the fold belts is predicted to be a result of the brittle-plastic transition on the basal decollement.

A remarkable linear relationship is observed between the relief of fold belts and their absolute elevations (Figure 15). Within the framework of wedge theory, the relief is controlled by the depth to the brittle-plastic transition and thus geothermal gradient. The depth to the brittle-plastic transition is predicted to be elevation dependent because of the high atmospheric temperature (470°C at mean planetary radius) and the 100°C change in temperature over the range of planetary elevations. However, the linear relationship between relief and elevation may be dominated by isostatic effects, which give rise to a linear relationship between elevation, lithospheric thickness, and geothermal gradient (Figure 16).

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