be at least partly of volcanic origin. On the moon, for instance, volcanic layering has been observed in the maria at scales of meters [*Howard et al.*, 1972] to hundreds of meters [*Peeples et al.*, 1978; *Sharpton and Head*, 1982]. Photographs at the Venera 9, 13, and 14 landing sites show evidence for centimeter-scale layering in the surface rock units [*Florensky et al.*, 1977, 1983; *Barsukov et al.*, 1982; *Garvin et al.*, 1984], although the thicknesses of these layers are much less than the values implied by the models given above. From recent radar images of Beta Regio, features interpreted to be lava flows may be traced for distances of several hundred kilometers [*Campbell et al.*, 1984]. *Warner* [1983] has proposed that the entire crustal column for many areas of Venus consists of a layered sequence of volcanic rock units separated by lenses of sediment (probably wind-transported volcanic material or weathering products). Such a crust may respond to compressive stress much as a layered elastic plate, with the volcanic rock units acting as elastic layers and the sediments acting as zones of weakness between them. It is unlikely, however, that the assumption of zero friction between individual layers will be strictly valid. In such a case the indicated thickness of the elastic unit and the compressive stress required to produce folding at the dominant wavelength λ will be intermediate between the predictions of (14) and (18) and (11) and (17), respectively [*Biot*, 1961].

Uniform viscous layer. As an alternative to one or more elastic layers, consider an incompressible layer of uniform Newtonian viscosity η and thickness h overlying an inviscid and incompressible fluid of density ρ, both in a uniform gravitational field of acceleration g (Figure 2, with viscosity η replacing elastic constants E and ν). This model has previously been used by us to treat the problem of viscous relaxation of topographic relief on Venus [*Solomon et al.*, 1982]. The solution for folding of such a layer in response to a uniform compressive stress σ has been solved by *Biot* [1959] as a special case of a model involving a viscous layer over a substrate of lower viscosity.

As in the elastic layer model there is a particular wavelength λ at which growth of folds occurs most rapidly [*Biot*, 1959]:

$$\lambda = \pi h\left(\frac{2\sigma}{\rho g h}\right)^{1/2} \tag{19}$$

Note that this expression holds also for the case of folding of a uniform elastic layer, as can be seen by combining (10), (11), and (14). For the viscous layer case, however, there is no formal minimum value of σ as in the elastic layer model [*Biot*, 1959], so that the expression for h as a function of λ analogous to (14) cannot be written without explicitly including the compressive load, i.e.,

$$h = \frac{\rho g}{\sigma}\left(\frac{\lambda}{2\pi}\right)^2 \tag{20}$$

There is, nonetheless, a practical lower limit to σ [*Biot*, 1959] because the amplitude of folding increases exponentially as $(\sigma/\rho g h)$. *Biot* [1959] uses the somewhat arbitrary criterion that amplification of fold amplitudes must exceed 10^2 to be deemed significant for an accumulated horizontal compressive strain of 25%; with this criterion he obtains $(\sigma/\rho g h) \gtrsim 7$ for significant folding. If this inequality is substituted into (19), we obtain

$$h \leq 0.085\ \lambda \tag{21}$$

with equality corresponding to the least value of σ satisfying the Biot criterion and therefore preferred. If folding of a viscous layer is adopted as a model for the formation of banded terrain on Venus, then λ = 15–20 km corresponds to a layer thickness h = 1.3 to 1.7 km. The compressive stress σ consistent with these values is 2.4–3.2 kbar. Equations (20) and (21), the latter at equality, are both shown in Figure 5; for the former relation a similar value for σ is used. As noted earlier, values of compressive stress in the vicinity of several kilobars are somewhat higher than we might expect to be supported without failure in the near-surface regions of the Venus crust. The required stresses can be reduced if *Biot*'s [1959] criterion is relaxed somewhat, but such a reduction would be less than a factor of 2. (The layer thickness is increased by the same factor for a fixed λ.) Thus kilobar-level compressive stresses are still necessary for the uniform layer model. As with the elastic cases, however, the compressive stress required to produce folding at a given wavelength can be lowered if the high-viscosity region consists of layers of high-viscosity material separated by thin layers of substantially lower viscosity [*Biot*, 1961]. In such a situation, σ decreases as n increases for a fixed value of λ.

Continuously decreasing viscosity with depth. Since viscosity is a strong function of temperature, a viscous model more representative of the rheology of planetary crusts or lithospheres may be one in which viscosity decreases continuously with depth rather than abruptly at the base of a layer. *Biot* [1960] has shown that folding in such a situation can be modeled by solving the instability problem for the surface of the medium.

The specific problem treated by *Biot* [1960] is that of a viscous incompressible half space in which the viscosity varies with depth z as

$$\eta = \eta_0 e^{-z/s} \tag{22}$$

where η_0 and s are constants. If such a medium is compressed horizontally at a uniform rate, the surface is unstable and folding develops. There is a dominant wavelength of folding given approximately by [*Biot*, 1960, 1961]

$$\lambda = \frac{2\pi}{2.2}\, s^{2/5}\left(\frac{\sigma_0}{\rho g}\right)^{3/5} \tag{23}$$

where σ_0 is the magnitude of the horizontal compressive stress at the surface. We may rearrange (23) to obtain s, the characteristic skin depth for a $1/e$ decrease in viscosity, in terms of λ:

$$s = 7.2\left(\frac{\rho g}{\sigma_0}\right)^{3/2}\left(\frac{\lambda}{2\pi}\right)^{5/2} \tag{24}$$

As with the case of the thickness of a uniform viscous layer, this measure of effective thickness of a surficial high viscosity region dependes explicitly on the magnitude of the compressive stress as well as on λ.

We may test (24) for Venus in two ways. We may postulate a value for σ_0 and solve for s. Alternatively, we may determine s independently from measurements of the activation energy for viscous creep in crustal materials and from estimates of the near-surface thermal gradient on Venus. Using the first approach, with σ_0 = 2 kbar and other parameters as before, λ = 15–20 km corresponds to s = 3–6 km (Figure 5).

For the second approach we need an independent estimate for s. We note that, from (22),

$$\frac{1}{s} = -\frac{1}{\eta}\frac{d\eta}{dz} \tag{25}$$