



Fig. 1. Mapped outlines of slump and debris avalanche aprons. The outline in the bottom right is that of a possible lava flow from the collapsed margin dome on Sapas Mons.

On Earth, landslides on volcanic edifices can be triggered by a number of different processes, including those occurring as a result of aseismical crustal deformation, such as oversteepening of slopes due to deformation (possibly resulting from dyke emplacement of magma rise), overloading of the slope (by lavas), excess weight at the top of the slope (due to a large cone or a large area of summit lava), removal of support by explosions on the flanks, and caldera collapse. Failure occurring coseismically can result from structural alteration of the constituent parts of the slope leading to failure, dislodgement of otherwise stable slopes, and fault movement resulting in an increased slope angle [4]. Seismic pumping may also be a major control on slope stability during an earthquake [5].

On Venus, similar processes may operate. The high ambient temperatures may result in development of a weak carapace, which in turn may allow relatively rapid dome growth to occur. If the effusion rates are high, as suggested by the size of the features, then oversteepening would be a likely consequence resulting in failure and collapse. Landslide scars may be modified by continued dome growth. The existence of fractures around the base of some of the collapsed domes and of debris aprons cut by fractures suggests that

there has been seismic activity and surface deformation occurring during the period of modification of the dome.

References: [1] Moore J. G. et al. (1989) *JGR*, 94, 17465–17484. [2] Holister C. D. et al. (1978) *EPSL*, 41, 405–418. [3] Siebert L. (1984) *J. Volcanol. Geotherm. Res.*, 22, 163–197. [4] Guest J. E. et al. (1984) *J. Volcanol. Geotherm. Res.*, 21, 1–23. [5] Murphy B. (1992) Ph.D. thesis, University of London.

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MIXED-VALENCE IRON MINERALS ON VENUS: Fe²⁺-Fe³⁺-OXIDES AND OXY-SILICATES FORMED BY SURFACE-ATMOSPHERE INTERACTIONS. Roger G. Burns and D'Arcy W. Straub, Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge MA 02139, USA.

Background: The oxidation state and mineralogy of iron on the hot surface of Venus are poorly understood [1–3], despite qualitative *in situ* measurements of oxygen fugacity during the Venera 13/14 missions [4], some reflectance spectral data derived from the

Venera 9/10 and Venera 13/14 missions [5], numerous equilibrium thermodynamic calculations [e.g., 3,6,7], and several experimental studies of rock and mineral assemblages heated in air [5,8-10,11] or CO₂-dominated atmospheres [2,12]. Inferences from these investigations are that Fe³⁺-bearing minerals such as hematite (α-Fe₂O₃), magnesian ferrite [MgFe³⁺₂O₄], acmite [NaFe³⁺Si₂O₆], and epidote [Ca₂Fe³⁺Al₂Si₃O₁₂(OH)] are thermodynamically unstable [3], and that magnetite [Fe₃O₄ or Fe²⁺Fe³⁺₂O₄] is the predominant mixed-valence iron oxide mineral on Venus. Recently, the Fe²⁺-Fe³⁺ silicate mineral laihunite (with ideal endmember composition Fe²⁺Fe³⁺₂Si₂O₈), earlier suggested to occur on oxidized surfaces of terrestrial planets [10], was proposed to be a reaction product of olivine with the venusian atmosphere [3]. This possibility is discussed further here. We suggest that other mixed-valence Fe²⁺-Fe³⁺-O₂-OH⁻ silicates could also result from surface-atmosphere interactions on Venus. Such minerals include oxy-amphiboles (e.g. Ca₂(Mg,Fe²⁺,Fe³⁺)₅Si₈O₂₂(OH,OH)₂), ilvaite [CaFe²⁺₂Fe³⁺Si₂O₈(OH)], oxy-biotites (e.g. K(Mg,Fe²⁺,Fe³⁺)₂₋₃(Si₃AlO₁₀)(OH,OH)₂), and babingtonite [Ca₂Fe²⁺Fe³⁺Si₃O₁₄(OH)].

Conversion of Hematite to Magnetite: Spectroscopic observations, which correlated spectrophotometric data of the Venera 9/10 and 13/14 missions with laboratory reflectance spectra of hematite, suggested that this ferric oxide phase might exist metastably on the venusian surface [5]. However, a kinetic study of the conversion of hematite to magnetite [12] showed that, although coatings of magnetite on hematite crystallites may partially inhibit the decomposition of Fe₂O₃, complete conversion of hematite to magnetite is sufficiently rapid at venusian surface temperatures so as to dismiss the existence of metastable hematite on Venus. Thus, magnetite appears to be the dominant oxide mineral accommodating ferric iron on Venus. However, a number of other opaque mixed-valence silicate minerals also host Fe³⁺ ions and could exist on the surface of Venus.

Stability of Laihunite: The occurrence of ferrifayalite, consisting of intergrowths of fayalite (stoichiometric Fe²⁺₂SiO₄) and laihunite (nonstoichiometric Fe²⁺_xFe³⁺_ySiO₄), in terrestrial metamorphic [13], plutonic, and extrusive igneous rocks [14-17] suggested that such Fe³⁺-bearing olivines might also exist on other terrestrial planets [10], including Venus [3]. Ferrifayalites have also been produced experimentally by aerial oxidation in the temperature range 400°-800°C of Mg-Fe and Mn-Fe olivines [8-10, 16,18,19]. These natural and experimentally induced oxidation products of olivine, including the granulite-facies laihunite (intergrown with magnetite) from metamorphic rocks at the type-locality in China [13], consist of a variety of superstructure olivine polytypes [16,17,20,21]. Maghemite (γ-Fe₂O₃ or nonstoichiometric spinel) was also identified in oxidized fayalites [8-10]. However, at elevated temperatures, ferrifayalite-maghemite assemblages decompose to hematite, leading to the suggestion [9] that a laihunite-maghemite-quartz oxygen fugacity buffer more oxidizing than the (stoichiometric) fayalite-magnetite-quartz buffer could exist at temperatures below 600°C. Such a Fe³⁺-bearing assemblage would be relevant, therefore, to the surface of Venus.

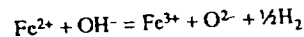
More recently, the stability of laihunite has been questioned, based on heating experiments above 300°C at various oxygen fugacities [18,19]. Laihunite synthesized from fayalite was transformed to hematite, magnetite, and amorphous silica under a CO₂ atmosphere. In a CO₂/H₂ (ratio 1000:1) atmosphere, laihunite decomposed to fayalite, magnetite, and amorphous SiO₂. Similar reaction products were obtained for laihunites synthesized from Mg-Fe and Mn-Fe olivine solid-solutions. It was concluded [19]

that at 1 bar atmospheric pressure and temperatures exceeding 300°C, laihunite has no stability field under any oxygen fugacity. These findings conflict with the numerous terrestrial occurrences of laihunite [13-17] and earlier thermodynamic calculations for the system Fe₂O₃-FeO-SiO₂ at 15 kb [22] that showed laihunite to be stable at granulite-facies temperatures (600°-700°C). Therefore, if laihunite does occur on Venus [3], it may be metastable in the venusian atmosphere at P ≤ 95 bar and T < 475°C.

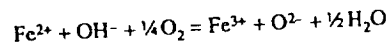
The Possible Existence of Oxy-Amphiboles and Oxy-Micas on Venus: Early thermodynamic calculations of chemical interactions between basaltic rocks and the atmosphere of Venus indicated that amphiboles such as glaucophane [Na₂Mg₃Al₂Si₈O₂₂(OH)₂] and tremolite [Ca₂Mg₅Si₈O₂₂(OH)₂], together with epidote, magnetite, and phlogopite [KMg₃Si₃AlO₁₀(OH)₂], are stable assemblages [1,6,23]. However, the suggestion that these hydroxyl-bearing minerals, as well as other amphiboles, micas, and clay silicates, are stable on Venus has been hotly debated (see [3]). The problem centers on estimated water-dissociation pressures of common OH⁻-bearing minerals, which render them unstable in the venusian atmosphere. On the other hand, fluorine-bearing amphiboles and micas, in which F⁻ ions replace OH⁻ in the crystal structures, may be stable on Venus [3,7].

Dehydroxylated amphiboles and micas formed by intracrystalline redox reactions could also be stable phases on Venus. Such dehydroxylated minerals are well known as oxy-hornblendes and oxy-biotites in terrestrial volcanic rocks, where they occur as primary magmatic phases or were formed by oxidation of extrusive amphiboles and micas at the Earth's surface.

When Fe²⁺-bearing amphiboles and micas are heated in a vacuum or inert gas, an internal oxidation-reduction reaction occurs in the mineral crystal structures, resulting in an oxy-amphibole or oxy-mica component [24,25]



In an oxidizing atmosphere, the reaction becomes



The overall result of both of these reactions is to convert the structural OH⁻ ions to O²⁻ anions, which are retained in the amphibole or mica crystal structure to charge-balance the Fe³⁺ ions produced by oxidation of Fe²⁺ cations [26,27]. Similar Fe²⁺-Fe³⁺-O²⁻-(OH⁻)-bearing amphiboles and micas would be thermally stable on the surface of Venus. Thus, oxy-hornblendes and oxy-biotites should be recognized as candidate mixed-valence iron minerals on Venus.

Other Mixed-Valence Fe²⁺-Fe³⁺ Silicates Likely to Exist on Venus: Calcite and magnetite are predicted to be stable over almost the entire surface of Venus [3]. On Earth, these two minerals coexist with a variety of Ca-Fe silicates in skarn deposits that formed by contact metamorphism when magma intruded ferruginous limestones or siliceous dolomites [28,29]. Minerals present in progressive decarbonation sequences away from such plutons range from high-temperature pyroxene (e.g., hedenbergite, CaFe²⁺Si₂O₆) and garnet (e.g., andradite, Ca₃Fe³⁺₂Si₃O₁₂) assemblages to ilvaite [CaFe²⁺₂Fe³⁺Si₂O₈(OH)] and babingtonite [Ca₂Fe²⁺Fe³⁺Si₃O₁₄(OH)], which formed at lower temperatures. In the CO₂-saturated environment during skarn formation, ilvaite-bearing assemblages become stable when temperatures and oxygen fugacities fall below about 500°C and 10⁻²⁵ bar respectively [30,31]. At compa-

rable temperatures, the crystallization of babingtonite requires more hydrous conditions, lower CO_2 , and slightly higher O_2 fugacities in the fluid phase than ilvaite. Since similar temperatures, CO_2 pressures, and oxygen fugacities induced within skarn deposits exist on Venus, ilvaite and perhaps babingtonite could have also formed on the surface of this planet by the interaction of the venusian atmosphere with extruded basaltic rocks. One factor that might mitigate against the formation of these calcic Fe^{2+} - Fe^{3+} silicates on Venus, however, are the high abundances of Mg and Al measured during the Venera 13/14 [32] and Vega 2 [33] missions. The Mg^{2+} and Al^{3+} cations are not accepted into the crystal structures of ilvaite and babingtonite.

Discussion: Although magnetite is generally regarded to be the predominant ferric-bearing mineral on Venus, other mixed-valence Fe^{2+} - Fe^{3+} minerals known to exist on the surface of Earth could be stable in the venusian atmosphere. Thus, in addition to laihunite (which is probably metastable) and ilvaite and babingtonite (both of which may be found in rocks depleted of Mg and Al), oxy-amphiboles and oxy-micas may also be major constituents of the venusian surface. The opacities and high electrical conductivities of such mixed-valence Fe^{2+} - Fe^{3+} silicate minerals, the properties of which resemble magnetite [34], may also contribute to high radar-reflectivity regions in the highlands of Venus [35].

References: [1] Volkov V. P. et al. (1986) *Adv. Phys. Geochem.*, 6, 136–190. [2] Gooding J. L. (1986) in *The Solar System Observations and Interpretations* (M. G. Kivelson, ed.), 208–299, Prentice-Hall. [3] Fegley B. Jr. et al. (1992) *Proc. LPSC*, Vol. 22, 3–19. [4] Florensky C. P. et al. (1983) *LPSC XIV*, 203–204. [5] Pieters C. M. et al. (1986) *Science*, 234, 1379–1383. [6] Barsukov V. L. et al. (1982) *Proc. LPSC 13th*, in *JGR*, 87, A3–A9. [7] Nozette S. and Lewis J. S. (1982) *Science*, 216, 181–183. [8] Bartels K. S. and Burns R. G. (1986) *Eos*, 44, 1270. [9] Bartels K. S. and Burns R. G. (1989) *Int. Geol. Congr.*, 28, 9293. [10] Bartels K. S. and Burns R. G. (1989) *LPSC XX*, 44–45. [11] Straub D. W. et al. (1991) *JGR*, 96, 18819–18830. [12] Straub D. W. and Burns R. G. (1992) *LPSC XXIII*, 1375–1376. [13] Zhang R. et al. (1981) *Tscher. Mineral. Petrog. Mitt.*, 28, 167–187. [14] Schaefer M. W. (1983) *Nature*, 303, 325–327. [15] Schaefer M. W. (1985) *Am. Mineral.*, 70, 729–736. [16] Kondoh S. et al. (1985) *Am. Mineral.*, 70, 737–746. [17] Banfield J. F. et al. (1990) *Contrib. Mineral. Petrol.*, 106, 110–123. [18] Iishi K. et al. (1989) *Neues Jahrb. Mineral. Mh.*, H6, 245–254. [19] Iishi K. et al. (1989) *Neues Jahrb. Mineral. Mh.*, H8, 345–356. [20] Kitamura M. et al. (1984) *Am. Mineral.*, 69, 154–160. [21] Shen B. et al. (1986) *Am. Mineral.*, 71, 1455–1460. [22] Wang S.-Y. (1980) *Geochimica (Chinese)*, 3, 31–42. [23] Khodakovskiy I. L. et al. (1979) *Icarus*, 39, 352–363. [24] Barnes V. E. (1930) *Am. Mineral.*, 15, 393–417. [25] Phillips M. W. et al. (1988) *Am. Mineral.*, 73, 500–506. [26] Phillips M. W. et al. (1989) *Am. Mineral.*, 74, 764–773. [27] Popp R. K. et al. (1990) *Am. Mineral.*, 75, 163–169. [28] Burt D. M. (1971) *Annu. Rept. Geophys. Lab. Yearb.*, 70, 189–197. [29] Burt D. M. (1971) *Soc. Mining Geol. Japan, Spec. Issue 3*, 375–380. [30] Gustafson H. J. (1967) *J. Petrol.*, 15, 455–496. [31] Gole M. J. (1981) *Can. Mineral.*, 19, 269–277. [32] Surkov Y. A. et al. (1984) *Proc. LPSC 14th*, in *JGR*, 89, B393–B402. [33] Surkov Y. A. et al. (1986) *Proc. LPSC 16th*, in *JGR*, 91, E215–E218. [34] Burns R. G. (1991) in *Mixed-Valence Systems: Applications in Chemistry, Physics and Biology* (K. Prassides, ed.), 175–200, NATO ASI-C Ser., Math. Phys. Sci., 343, Plenum. [35] Research supported by NASA grant number NAGW-2049.

THERMAL BUOYANCY ON VENUS: PRELIMINARY RESULTS OF FINITE ELEMENT MODELING. J. D. Burt and J. W. Head, Department of Geological Sciences, Brown University, Providence RI 02912, USA.

Introduction: Enhanced surface temperatures and a thinner lithosphere on Venus relative to Earth have been cited as leading to increased lithospheric buoyancy. This would limit [1] or prevent [2] subduction on Venus and favor the construction of thickened crust through underthrusting. Underthrusting may contribute to the formation of a number of features on Venus. For example, Freyja Montes, a linear mountain belt in the northern hemisphere of Venus, has been interpreted to be an orogenic belt [3] and a zone of convergence and underthrusting of the north polar plains beneath Ishtar Terra, with consequent crustal thickening [4]. Such mountain belts lie adjacent to regions of tessera and contain evidence of volcanic activity. Tessera is also considered to consist of thickened crust [5] and crustal underthrusting is one possible mode for its formation [4]. Models for the formation of the mountain belts and associated features must then explain compressional deformation and crustal thickening, as well as melt production.

In order to evaluate the conditions distinguishing between underthrusting and subduction, we have modeled the thermal and buoyancy consequences of the subduction end member. This study considers the fate of a slab from the time it starts to subduct, but bypasses the question of subduction initiation. Thermal changes in slabs subducting into a mantle having a range of initial geotherms are used to predict density changes and thus their overall buoyancy. Finite element modeling is then applied in a first approximation of the assessment of the relative rates of subduction as compared to the buoyant rise of the slab through a viscous mantle.

Subduction Model: In the model, slabs, having a thickness set by 90% of the basalt solidus, subduct at a 45° angle into the mantle. The initial geotherms match surface thermal gradients of 10°C/km, 15°C/km, and 25°C/km [6]. Slabs heat via conduction, crustal radioactivity, phase changes, and adiabatic compression. Phase changes involving the conversion of basalt to eclogite at depths of 60 to 160 km and then enstatite to forsterite plus stishovite between 260 and 360 km generate 0.13×10^{-5} ergs/cm³s and 0.36×10^{-5} ergs/cm³s respectively [7]. The slab radiogenic heat production is 2.63×10^{-7} ergs/g s [8]. Adiabatic compression adds 0.5°C per kilometer of depth. Convergence rates ranged from 5 mm/yr to 100 mm/yr.

The thermal evolution of the slab is followed using a finite difference technique [7,9]. The model region measures 800 km horizontally by 400 km deep. Processing ends when the slab tips reach a 300-km depth, implying time intervals of 10 m.y. to 100 m.y. Slab density changes derive from the thermal results through calculation of the thermal expansion ($\alpha_v = 3 \times 10^{-5}/^\circ\text{K}$ [8]) and the effects of pressure ($b = 1 \times 10^{-3}/\text{kbar}$ [8]) on initial densities set for zero pressure and temperature. The assumed initial density structure includes a 10-km or 25-km basaltic crust (density = 3.0 g/cm³), a corresponding 25-km or 65-km-thick depleted mantle zone (density = 3.295 g/cm³), and an underlying undepleted mantle (density = 3.36 g/cm³). Density changes due to the phase transitions are also included. Results take the form of density distributions within the model region.

Finite element modeling is then employed to gauge the rate at which a slab having the density structure derived above will move through a viscous mantle. Buoyancy body forces are applied to a slab having a viscosity of 10^{21} Pa s surrounded by a mantle with a