FORMATION OF TERTIARY CRUST ON VENUS BY REMELTING OF TESSERA
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Tertiary crust forms by the reprocessing of primary and secondary crusts by several possible mechanisms, according to Taylor [1] the Earth's continental crust is the only presently known example of tertiary crust. One of the most distinctive morphologic volcanic signatures of evolved crust on Earth is the presence of steep-sided domes that form from the extrusion of viscous magma [2]. Features of similar morphology have been observed on the Moon [3] and Venus [4, 5], and at least one occurrence on Venus may be linked to more evolved composition [6]. We have been analyzing several environments [7] and candidates for tertiary crust on Venus, specifically asking several questions, including: What is the evidence for post-formational evolution of tessera in terms of melting of thickened crustal roots and associated volcanism of potentially non-basaltic composition? Here we report on the characteristics, setting, stratigraphic position and preliminary assessment of the petrogenesis of the distinctive festoon deposit [8,9] lying within some of the highest-standing tessera in Ovda Regio.

GENERAL SETTING. The festoon structure (Fig. 1) lies within Ovda Regio at about 6.5S, 95.5E and is about 250 by 300 km in dimension. Ovda Regio is one of the most distinctive occurrences of tessera on Venus [10] and this part of Ovda is one of the highest-standing on the planet; the tessera at the edges of the festoon lie at about 0°56.2, -4.4 km above MRP. Stratigraphic relationships show that the festoon overlies the tessera terrain; the morphology and internal structure of the two terrains contrast distinctly, digitate and lobate projections at the deposit edges follow preexisting structural troughs and fractures in the tessera, and there are kipukas of tessera within the festoon. The deposit itself is elongated in a NE-SW direction, parallel to structural trends in adjacent tessera.

DEPOSIT CHARACTERISTICS. Radar images of the festoon show three main radar characteristics; most of the margins of the festoon (outer 15-50 km) appear relatively radar bright, the interior is intermediate, and the southeastern part (and the adjacent tessera) is very radar dark. The lowest emissivity value measured on Venus (0.26) occurs in the eastern part of the festoon, and the radar dark area to the south of the flow has a value of 0.84 [8]. Although the most distinctive flow front within the festoon deposit coincides in part with the high emissivity boundary [9], in general the boundaries do not coincide with geologic units [11]. Because of this we consider the radar bright and dark units with steep lobate margins and textured interior as a single festoon deposit, possibly composed of multiple flow units. The deposit internal structure is characterized by fingerprint-like swirls composed of ridges and troughs with typical wavelengths of 500-750 m. Map patterns of these features show that they are influenced by preexisting topography (particularly in the vicinity of the kipukas and flow margins), and that although irregular, they tend to be concave toward the central portion of the festoon deposit. Evidence for internal flow lobes is very rare with the most prominent exception being the border of the radar dark portion of the festoon deposit (arrow in Fig. 1) and is in contrast to the multiple flow units mapped in Mahuea Tholus [5]; these observations suggest that most of the flow unit (particularly the bright portion) was emplaced in a single eruptive phase [9]. The distinctive lobate boundary suggests that at least portions of the radar dark part of the festoon were emplaced separately and later. Analysis of lobate flow margins (both external and internal at kipukas) indicate flow thicknesses in the range 50-150 m [5]. These data suggest total volumes for the festoon flow deposit of the order of 5500 km³. Linear bright structures, in the form of scarps and paired features interpreted to be graben, are seen to cross the festoon deposit and cut it as evidenced by fractured festoon structures, and lack of perturbation of festoon rope orientation that would occur if the structures predated emplacement; these are parallel to structural trends in the underlying tessera (Fig. 1b) and are interpreted to be related to later stages of Phase II (latest) deformation within the tessera terrain. The topography of the deposit is not flat; there is a broad depression in the interior of the deposit that is about 1.6 km below the margins [8] and this generally coincides with the locus of festoon ridge concavity. The nature of the deposit suggests that it was emplaced on rough tessera terrain, but that the surface of the flow was initially relatively flat, or perhaps sloping slightly away from a central source region; these data thus suggest that subsidence of a magnitude considerably greater than the deposit thickness occurred after the emplacement of the festoon deposit [8,9]. Analysis of deposit densities, flow heights and ridge spacing have been used to estimate viscosity and yield strength, and these data are comparable to those for silicic terrestrial lava flows [5,9].

STRATIGRAPHIC RELATIONSHIPS. Internally, the vast majority of the bright portions of the festoon appear to have been emplaced in a single flow event, while parts of the radar dark portion of the festoon may represent a later flow unit [9]. Externally, the festoon deposit is superposed on the tessera and on early intratessera plains (Fig. 1b). Superposition on early intratessera plains, defined as those plains cut by relatively widely spaced graben of the later period of tessera Phase II extensional deformation [12], as well as cross-cutting of the festoon deposit by fractures and graben with this same structure and orientation, is interpreted to mean that the deposit was emplaced during the waning period of Phase II deformation; Phase II deformation has been interpreted to follow immediately Phase I compressional deformation [10], and to have lasted at most only a few tens of millions of years [13].

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PETROGENESIS. Interpretation of the underlying tessera as a product of downwelling and crustal thickening [10,14] provides a basis for the assessment of petrogenesis of this deposit. On the basis of evidence for lava plains as tessera precursor terrain [12] and the likelihood of a basaltic crust throughout the history of Venus, one of the main candidates for the origin of this deposit is the remelting of a basaltic crust initially derived from melting of a peridotite mantle. Venera lander analyses of presently exposed plains lead us to consider the remelting of tholeiitic basalt [7,6,15] under anhydrous conditions [7]; melting of tholeiitic basalt above 15 to 25 kb (about 53-88 km) begins at temperatures in excess of 1200°C in and the eclogite facies (or in the garnet granulite facies, depending on the bulk composition) [16]. The melt that coexists with the eclogite assemblage (garnet and clinopyroxene) typically is quartz-normative and strongly enriched in SiO2 [17]. Small degrees of melting (<20%) generate trondhjemites (SiO2 > 65%) whereas intermediate degrees of melting (20-50%) yield andesites and basaltic anesites. In contrast, the first melts obtained at lower pressures (10-15 kb; 35-53 km) appear to be relatively SiO2-poor. For example, the liquids obtained by small degrees of melting at 8 kb (28 km) of high aluminobasalts are ferrobasalts containing only 42% SiO2 and more than 20% FeO [18]. In summary, large amounts of silicic magmas are generated at high pressures and large amounts of relatively SiO2-poor, and highly fluid, ferrobasalts are obtained under more modest pressures. Thus, shallow crustal melting occurring in environments such as underthrust basaltic crust and basal melting of tessera crustal blocks less than about 50 km thick should result in the production of fluid ferrobasalts. Deeper crustal melting, such as that which might be occurring at the base of zones of very thick tessera, should produce more viscous SiO2-rich melt products such as trondhjemites, andesites, and basaltic anesites. The fact that the festoon deposit occurs at the highest elevations in Ovda Regio [8] suggests that in terms of simple Airy isostasy this is an area of some of the thickest crust on the planet [19] (and thus deepest melting), a conclusion consistent with apparent depths of compensation of 70 +/- 7 km [20] and Magellan gravity data [21].

INTERPRETATION AND CONCLUSIONS. We interpret the festoon deposit to be the product of relatively deep (in excess of ~50 km) remelting of thickened basaltic crust following shallower crustal melting producing deposits of more fluid basalts (producing early intratessera plains). This scenario is consistent with: 1) processes of downwelling, progressive crustal shortening and thickening during tessera formation [14, 10] to produce basaltic crustal roots at many tens of km depth, and 2) the timing for the formation and modification of the tessera terrain in which both stages of melting occur subsequent to shortening and prior to the time that tessera extensional deformation related to gravitational relaxation has ceased. This petrogenetic model provides an independent estimate of crustal thickness, an estimate consistent with those from gravity data [20,21].