PARALLEL FRACTURE PATTERNS ON THE PLAINS OF VENUS

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Plains of presumably volcanic origin constitute the most extensive single geological unit on Venus, covering over 80% of its surface. These volcanic plains contain a wide variety of structural forms and patterns [1]. Large-scale structures include linear belts of intense deformation, such as rifts and ridge belts, and quasi-circular patterns associated with coronae. However, in addition to these examples of concentrated strain, there is also much evidence for distributed deformation in the form of nearly ubiquitous fractures and ridges. These smaller-scale features reflect a rich variety of deformation styles. In the present study we concentrate on a particular class of features, which we term "parallel fracture sets", that occur as groups of parallel, narrow lineations with average spacings within a given set on the order of a kilometer.

These features were first noted as the fainter of two sets of nearly orthogonal lineations observed in early Magellan images of Guinevere Planitia [1]. Subsequently, we have identified many more of these parallel patterns in Magellan radar images, all with strikingly similar characteristics. The trend of the lineations is very nearly straight, or broadly arcuate. Their spacing is remarkably uniform, and is on the order of a few kilometers. The lineations themselves are featureless at the resolution of the Magellan images (75 m/pixel), and appear to be no more than a single radar resolution cell in width. The sets cover relatively broad areas, and are often obscured by local deposits around small volcanic domes. They typically occur in flat plains with low slopes, and their orientations display no apparent correlation with topographic gradient. Another commonly observed characteristic is the occurrence of less well-developed cross-cutting sets of similar lineations with an intersection angle of ~45°.

We have chosen a number of examples of parallel fracture sets, and measured the spacing between lineations along profiles orthogonal to the lineations. Digital images were analyzed by recording the pixel line and sample coordinates of each fracture crossing along the profile. Average spacings for fracture sets were found to range between 1 and 2.5 km.

A number of mechanisms have been proposed to account for the regular spacing of some features observed in pre-Magellan images of Venus, including elastic flexure, instabilities in viscous or plastic layers, and elastic-plastic flexure and buckling [2-4]. Whereas these mechanisms have a clear application to the observed tectonic periodicities of 10-20 km and 100-300 km, they do not appear suitable for explaining the 1-3 km spacing of the parallel fracture patterns. These theories all require that the thickness of the mechanical layer (which must be underlain by a layer of negligible, or at least considerably lower, strength) to be less than about 400 m [1]. Assuming a basaltic composition for the plains, an unreasonably high thermal gradient of nearly 100°/km would be required in order for a layer of this thickness to be possible under conditions of conductive thermal equilibrium. Such a thin layer will form during the cooling of a molten lava flow, but this will be a transient condition, and it is difficult to see how the numerous fault sets described above could form consistently within this short interval (on the order of a few thousand years [5]).

Another problem is the nature of the deformation implied by the assumptions in these models. For the extensional flexure-controlled models, the elastic surface layer fails by normal faulting, and vertical displacement along these dipping faults causes the layer to bend at the characteristic wavelength determined by its thickness and elastic constants. This results in either horst and graben formation, with alternating raised and lowered blocks, or in imbricate normal faults, with block rotations about a horizontal axis. Evidence for either of these situations should be visible in SAR images of the parallel fracture sets at Magellan's resolution, but none has been observed. Compressional buckling generally produces broad, complex structures, more reminiscent of wrinkle ridges, rather than the sharp lineations observed. Similarly, viscous or plastic necking would be expected to result in evenly spaced bands of more distributed deformation. The lineations we have observed appear to have the same appearance from any azimuth, implying that the reflection is due to enhanced roughness rather than from topographic relief of fault scarps. This observation is most consistent with tension cracks or joints. This implies the layer must have been relatively intact and unfractured at the time the parallel fracture sets formed, because frictional slip on pre-existing fractures is favored over tensile failure of intact rock [6,7].

We propose a shear-lag mechanism to account for the close, nearly uniform spacing of the parallel fracture sets observed on the plains of Venus. This type of model has been used in the...
materials sciences to describe the cracking of a brittle film that is ductilely coupled to an elastic substrate [8]. In a shear–lag model, the average spacing of fractures \( d \) is determined by a balance between the tensile strength of a brittle plate and a traction on its base, and is given by the relation 
\[
d = \frac{3\sigma_f h}{2f},
\]
where \( \sigma_f \) is the tensile strength of the plate, \( h \) is its thickness, and \( f \) is the shear stress applied to its base. The key to applying this model is in establishing the nature of the mechanical boundary that defines the plate thickness, as this will define the mechanism responsible for the traction. The depth to this boundary must be less than about \( d/3 \) (300-700 m in this case) in order for the shear–lag mechanism to produce evenly-spaced fractures. Note that this is comparable to the depth required by the other models discussed above, but is a shear boundary only, and not necessarily a boundary between layers of differing bulk mechanical properties. One candidate for a mechanical layer of this thickness is a flood basalt deposit (from one or more eruption episodes). If this layer is not intimately bonded to the material below, it may be possible for it to detach and slip horizontally along the contact plane when the underlying lithosphere subjected to tectonic extension. The shear stress transmitted across this interface will be governed by a frictional relation 
\[
f = \mu \sigma_n = \mu \rho g h,
\]
where \( \mu \) is the component of stress normal to the interface, \( \rho \) is the density of the layer, and \( g \) is gravitational acceleration. Combining this with the shear–lag equation, we obtain 
\[
d = \frac{3\sigma_f h}{2\mu\rho g}.
\]
Thus the spacing expected for tensile cracks in an initially unfractured basalt flow in sliding contact with its substrate is independent of the flow thickness, and depends only on its density, tensile strength, and the coefficient of friction. All of these parameter values are well bounded, with the exception of \( \sigma_f \); we adopt \( \rho = 2.7 \times 10^3 \text{ kg/m}^3 \), \( g = 8.87 \text{ m/s}^2 \), and \( \mu = 0.6 \). The tensile strength of a large-scale rock mass is more problematic. Laboratory measurements for tensile strength of diabase give values around 40 MPa [9]. Using this value, a predicted spacing of about 4.2 km is obtained, some two to four times larger than the observed spacing.

The tensile strength of the layer will certainly be less than the laboratory value because of the well-known (but poorly quantified) scale dependence of the brittle yield properties of rocks due to the large number of flaws (ranging from microcracks to joints) in real rock masses [e.g., 10]. This effect is thought to decrease the strength by perhaps an order of magnitude on the Earth. However, the high temperature at Venus' surface will lessen this effect somewhat relative to terrestrial conditions because the thermal contraction stresses in a cooling lava will be considerably lower (roughly by a factor of 2) and enhanced solid–state creep at crack tips will tend to lower the stress intensity factor, limiting crack growth. Thus an apparent layer tensile strength several times lower than laboratory values is reasonable.

If this model for the parallel fracture sets is assumed to be valid, then a number of inferences can be made about the history and material properties of the surfaces on which they are found. Ideally, one could use fracture spacing to determine relative values of the physical parameters in the shear–lag equation. However, the dependence on these parameters is not strong, and it may not be practical to utilize fracture spacing in this fashion. More importantly, the requirement that the layer be relatively intact and unfractured means that the tectonic episodes responsible for the parallel fracture set must predate other deformation events recorded on that surface. Thus these structures can serve as a relative temporal marker for a deformation sequence, showing the orientation and sense of the earliest tectonic event to which the plains unit was subjected.

An example is the deformation around Heng–O Corona. In this case radial parallel fracture sets are cut by circumferential structures, most of which are extensional. This implies that the circumferential extension, due to radially directed membrane expansion of the lithosphere around the corona, came early in the tectonic history of the corona, but perhaps soon after extensive volcanic resurfacing. Subsequently, radial tensile stresses that probably resulted from flexure around the edges of a super–isostatic corona load formed the circumferential extensional features outside the rim. This sequence places important constraints on allowable models for the formation of the corona [e.g., 11].

ANALOG STUDIES OF NANOPHASE IRON OXIDES IN MARS SOIL

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The surface of Mars is covered with fine weathered soil material which has been studied rather intensively over the last 15 years. The mineralogical constituents of the soil have not been analyzed directly yet, and much controversy has been surrounding the issue, with numerous candidates proposed (1). Much attention was given to the iron minerals in the soil, due to their pronounced absorption features in the VIS and NIR, making them easily detectable by telescopic observations (2,3).

The typical reflectance spectrum of the bright regions of Mars in the visible/short wavelength NIR bears the strong fingerprints of oxidized iron (4,5,6). It has been reproduced more or less faithfully in laboratory measurements of a number of different iron-containing systems. These include "amorphous" iron oxides (7,8), palagonite (2,9,10), nanophase hematite deposited in silica or alumina matrix (11,12), and iron-enriched smectites (13,14). It appears that iron in the weathered component of Mars soil is mostly characterized by having clusters of poorly crystallized oxide or oxyhydroxide ferric iron, or crystalline minerals in extremely small particle size. Recent telescopic observations detected typical hematite features at 860 nm, and possibly other crystallized iron oxides (6,15). However, these features are very weak and estimates of the content of hematite and/or other crystalline phases are in the range of 1-5% of the soil (6,16). Thus the reflectance data still suggest that the bulk of the iron oxide-oxyhydroxide in the Mars soil is amorphous or short-range ordered and is characterized by extremely small particle sizes ("nanophases" (np)).

The two most thoroughly studied mineral analogs to the iron nanophases in Mars soil are np-hematite (11,12) and iron-enriched clays studied by Banin et al. (13,14,17-19). Which one of these is the "true" Mars soil analog, if any?

Analog synthesis: The production of np-hematite (11,12) follows a common procedure for hematite synthesis i.e. prolonged high temperature (550°C) calcination of ferric iron solutions pre-deposited inside silica or alumina matrix. Samples with varying iron contents were obtained by different numbers of cycles of ferric nitrate solution impregnation followed by drying and calcination. The procedure is clearly conducive to the formation of hematite, and hematite only, since this mineral is the most thermally stable iron oxide form. It does not represent a prevailing nature alteration environment where hematite is likely to form, at least not as an amorphous or long-range ordered material.

The iron nanophase oxides-oxyhydroxides in the Mars soil analog clays are deposited on the surface of the clay in a laboratory process (14,20,21) which generally simulates, but considerably enhances, the natural reaction sequences of hydrolysis-precipitation-oxidation occurring during the oxidative weathering of ferrous-iron containing primary silicate rocks on Earth. The process involves slow titration of acidic clay with a solution of Fe(II) salt while the pH is controlled by the presence of OH-ion exchange resin. The later stages of the process, during which most of the iron is deposited, occur within a narrow range of pH, in a slightly acidic to slightly basic environment (pH 6.6-7.4), and at practically constant redox (PE + pH = 11.5-12.0). The double hydroxy salt of iron, ferric oxide (Fe(II)Fe(III)2(OH)g), appears to form at this stage (19). This is typical for many natural environments on Earth in which ferrous iron in solution is oxidized and forms insoluble "green-rust" (a double hydroxy-iron phase) which then, upon further oxidation, may transform to either goethite, lepidocrocite, maghemite or ferrhydrite. Such processes may have taken place on Mars during acidic-oxidative weathering of rocks (22). Crystal growth of iron oxides in the iron-enriched clays is limited by the thermodynamic and electrostatic conditions in the precipitation system and not mechanically. Furthermore, precipitation is done from solution at low ferrous iron concentration, involving in situ oxidation at acidic to near-neutral pH - all representing potentially realistic weathering environments on Earth or on Mars. The iron phases are virtually x-ray amorphous, but selected area electron diffraction have detected some crystallization and identified lepidocrocite (γ-FeOOH) as the mineral present (19). The mineral appears to crystallize only after a relatively high load of iron was added to the clay (3-4 times the cation exchange capacity; 8-10% Fe2O3 added). It forms extremely small particles and quantitatively remains a minor phase.

Spectral properties: The reflectance spectra of both Mars-soil iron-analogs strongly resemble the reflectance curves of the bright regions on Mars in the visible range (11,13). Both the Mars and the analog's
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spectra do not show any well defined absorption features of the crystalline iron oxides in the visible range but are characterized by monotonous and significant decrease in reflectance from the NIR towards the UV. This spectral behavior has been generally attributed to nanophasic iron oxides, and was recently specifically attributed to superparamagnetic nanophasic hematite (6,11). This assignment does not appear to be unique and singular since we find that amorphous iron oxides and nanophasic lepidocrocite also show the same reflectance spectrum in the VIS and agree quite well with the Martian spectra. Similarity to Mars spectrum has been shown to improve by mixing and adding various minor phases (either actually or by computation). However, at the present stage of the refinement of the spectral telescopic data for Mars, it is doubtful to what extent such “fine-tuning” is warranted.

Viking Biology Simulations: The chemical reactivity of Mars soil, as recorded in the Viking biology experiments, has been simulated successfully with the iron enriched clays. The analogs decomposed $^{13}C$ formate, the most labile organic component in the media used in the Labeled Release experiment, at a rate and to the extent measured on Mars (14,17) and simulated the sorption ("peak 1") and synthesis ("peak 2") activities of the Mars soil measured during the Pyrolytic Release experiments (23). These unique chemical reactivities are attributed to the combined catalytic effects of the iron oxide-oxyhydroxide and silicate phase surfaces (14,17). Information is lacking for np-hematite, but simulation studies with palagonites (18) and with pure hematite (17) have shown that they do not induce the LR reaction.

Magnetic properties: We have recently found that heating of the iron-enriched clays, in which lepidocrocite has been identified, renders them magnetic (24). The saturation magnetization ($J_s$) of an iron enriched clay (containing 9.68% total iron as Fe$_2$O$_3$) increased from $1.56x10^{-8}$ m$^3$/kg ($1.61x10^{-6}$ m$^3$/kg Fe$_2$O$_3$) for the unheated sample to $2.8x10^{-6}$ ($2.89x10^{-5}$ m$^3$/kg Fe$_2$O$_3$) for the heated ($300^\circ$C, 12h) sample, bringing it to the range of magnetic susceptibilities estimated for the Mars soil, if total iron was the same (25,26). The heat treatment converted the antiferromagnetic lepidocrocite to the strongly ferrimagnetic maghemite - a well known conversion.

The $J_s$ values for np-hematite are about 10 times higher than those of hematite (11). To account for the observations in the magnetic experiment on Mars, all the iron in the soil has to be present as np-hematite (11). Because $J_s$ of maghemite is about three orders of magnitude higher than that of hematite (and ca. two orders of magnitude higher than np-hematite), it is sufficient that a small fraction of the np-iron oxides in the soil are converted to maghemite (via lepidocrocite) to develop saturation magnetization similar to the Mars soil.

Summary: np-hematite produced in confining matrices and np-iron oxyhydroxide + np-lepidocrocite on clay surfaces, are two mineral analogs to the nanophasic iron oxides in Mars soil, which have been studied extensively in recent years. Both analogs show convincing overall spectral and magnetic similarities to the Mars soil. The np-iron oxide on clay surfaces also simulates the chemical reactivity of the soil. The evidence they supply is quite compelling leading to the conclusion that much of the iron oxides in the weathered Mars soil and in the dust, are in the nanophasic. However, the formation scenarios and detailed mineralogy of the two analogs are grossly different from one another. In order to select either one of them as the "true" analog, or reject both, better and more detailed spectral data for Mars and additional simulation studies on Earth are required.

References: