Geologic interpretation of the near-infrared images of the surface taken by the Venus Monitoring Camera, Venus Express

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Abstract

We analyze night-time near-infrared (NIR) thermal emission images of the Venus surface obtained with the 1-µm channel of the Venus Monitoring Camera onboard Venus Express. Comparison with the results of the Magellan radar survey and the model NIR images of the Beta-Phoebe region show that the night-time VMC images provide reliable information on spatial variations of the NIR surface emission. In this paper we consider if tessera terrain has the different NIR emissivity (and thus mineralogic composition) in comparison to the surrounding basaltic plains. This is done through the study of an area SW of Beta Regio where there is a massif of tessera terrain, Chimon-mana Tessera, surrounded by supposedly basaltic plains. Our analysis showed that 1-µm emissivity of tessera surface material is by 15–35% lower than that of relatively fresh supposedly basaltic lavas of plains and volcanic edifices. This is consistent with hypothesis that the tessera material is not basaltic, maybe felsic, that is in agreement with the results of analyses of VEX VIRTIS and Galileo NIMS data. If the felsic nature of venusian tesserae will be confirmed in further studies this may have important implications on geochemical environments in early history of Venus. We have found that the surface materials of plains in the study area are very variegated in their 1-µm emissivity, which probably reflects variability of degree of their chemical weathering. We have also found a possible decrease of the calculated emissivity at the top of Tuulikki Mons volcano which, if real, may be due to different (more felsic?) composition of volcanic products on the volcano summit.

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1. Introduction

The Venus Monitoring Camera (VMC) is a part of the Venus Express payload. It takes images in four spectral channels; one of them centered at 1.01 µm registers the night-side thermal emission from the planet’s surface (Baines et al., 2006; Markiewicz et al., 2007, 2008). Formal spatial resolution of these images taken from the working distances (2000–8000 km) is 1–5 km per pixel, but because the surface radiation on its way to the camera passes through the dense scattering atmosphere and cloud layer, the actual spatial resolution at the surface is about 50 km. The radiation intensity depends on the surface temperature, thus giving a hope to register any ongoing volcanic eruptions. This issue, however, is not considered in this paper and will be discussed elsewhere. Also the radiation intensity depends on the emissivity of the surface material, which is a function of a number of parameters including surface texture at micron to millimeter scale and mineralogical composition. The latter is interesting for a search for geological features and terrains whose chemical/mineralogical composition may be different from that of dominating basalts. On Venus with its very massive atmosphere, the surface temperature has practically no diurnal, seasonal and latitudinal variations and is a function of surface elevation (see summary in Crisp et al. (1997)). So in searches for volcanic eruptions and in attempts to find mineralogical differences, it is necessary to take into account the elevation of...
the given place and build model images showing elevation-dependent and emissivity-dependent thermal emission of the surface. Temporal variations in the cloud layer density also should be taken into account when possible.

The goal of this paper is to investigate feasibility of the geologic analysis of the VMC near-infrared images of the surface and present the preliminary results. We study a few types of terrains and features in the area southwest of Beta Regio, which are well covered by the night-time VMC images and allows the possibility to look for potential differences in chemical/mineralogical compositions of some terrains. We undertook a geologic analysis of the study area involving Magellan synthetic aperture radar (SAR) images and other Magellan mission results available from the Planetary Data System (PDS) and also accessible from http://www.mapaplanet.org/. First considerations of the possibility of using the VMC images for the geological analysis have been given by Basilevsky et al. (2008, 2010). Our study is complementary to the study based on analysis of the Venus surface thermal emission measured by Visual and Infrared Thermal Imaging Spectrometer (VIRTIS), another instrument of Venus Express (Helbert et al., 2008; Arnold et al., 2008; Mueller et al., 2008; Smrekar et al., 2010; Gilmore et al., 2011a).

2. Geological and mineralogical background

2.1. Venus primary bedrock geology and rock compositions

The Venus surface is dominated by volcanic plains, often called regional plains (e.g., Basilevsky and Head, 1998, 2000; Basilevsky and McGill, 2007), which have been interpreted to be formed by emplacement of mafic (basaltic) lavas. This inference follows from the results of the in situ geochemical measurements by the Venera and Vega landers (see summary by Surkov (1997)) in six sites located on these plains (Abdrakhimov, 2001a,b,c,d,f) and is supported by observations of plains morphology on high-resolution radar imagery (e.g., Barsukov et al., 1986; Head et al., 1992). Due to their wide areal distribution regional plains represent a reference surface for the analysis of the VMC night-side images. In places amid the regional plains there are areas of geologically younger morphologically distinctive volcanic plains, so-called lobate and smooth plains, which also seem to be basaltic based on morphology.

In the Magellan SAR images of Venus more than a hundred volcanic constructs larger than 100 km in diameter and about 300 constructs of 20–100 km in diameter are observed (Crumpler et al., 1997; Magee and Head, 2001). The youngest lavas related to these constructs are clearly superposed over regional plains. These large- and intermediate-size volcanoes are morphologically very similar to basaltic shield volcanoes on the Earth, although the latter are typically smaller than their counterparts on Venus. A basaltic composition of lavas of at least one volcanic edifice is supported by the in situ geochemical measurements by Venera-14 (Surkov, 1997), which landed within the lavas of the Panina Patera volcano (Abdrakhimov, 2001e). Volcanic constructs on Venus are often associated with rifts resembling continental rifts of Earth (e.g., Schaber, 1982; Head et al., 1992; Solomon et al., 1992; Price and Suppe, 1994; Basilevsky and Head, 2000). Sometimes venusian rifts are also sources of plains-forming lava flows.

The highest volcano on Venus, Maat Mons, stands about 9 km above the mean planetary radius of 6051.8 km. Lava flows radiating from Maat Mons cover an area about 800 km across. In their eastern extension, these lavas are superposed on 40–km-crater Uvaisy, which has an extended radar-dark parabola. Presence of the latter suggests that crater is very young, not older than a few tens of millions years (Basilevsky, 1993; Basilevsky and Head, 2002a). This is a strong indication that Venus has volcanoes active in the geologically recent time with high chances that some can be active at the present time, although the mean rate of venusian volcanism in the geologically recent time is probably by 1–2 orders of magnitude lower than the mean rate of volcanism of the Earth in the current geologic epoch (Basilevsky and Head, 2002b).

There are a few geologic features and units of the Venus surface, which could have non-basaltic, geochemically more evolved compositions. These are so-called steep-sided domes and tessera terrain. The steep-sided morphology of the domes suggests that they were formed by eruptions of viscous lavas which are often typical for geochemically evolved compositions (e.g., Pavri et al., 1992) although other suggestions on their nature have been published: low-eruption rate basaltic volcanoes (Fink and Griffiths, 1998), increased content of dissolved water and difference in crystallinity (Bridges, 1995) or foamy basaltic lavas (Pavri et al., 1992).

Sizes of these features are only a few tens of kilometers so they are too small to be resolved in the VMC images. But their presence may be a hint on different composition in the formation of VMC data.

Tessera terrain is another candidate for non-basaltic composition of its material. Nikolaeva et al. (1992) have compiled several lines of evidence that tessera can be composed of the material geochemically more differentiated than basalts, for example, essentially feldspathic materials such as silicic to intermediate rocks or anorthosites. Later, joint analysis of the gravity field and topography of Ishtar Terra allowed Kucinskas et al. (1996) to conclude that some parts of Maxwell Montes highland consisting of material structurally similar to tessera, could be composed of material less dense than basalt and possibly be silicic. Recently Gilmore et al. (2011b) have performed structural analysis of a block of tessera in Tellus region and applied a model of deformation formed the ridges in that block from the VIRTIS data analysis. They concluded that the material of this block could be a range of compositions including felsic (Gilmore et al., 2011a). At the same time, in several localities Ivanov (2001) has observed evidence that tessera was formed through tectonic deformation of some precursor plains. Suggesting that these plains have basaltic composition, he concluded that the tessera material could be also basaltic. None of the Venera or Vega geochemical probes landed on tessera terrain; thus, all information about its composition is indirect. Tessera forms blocks of different sizes, up to several hundreds to thousands kilometers across, that makes it possible to study them through the analysis of the IR emission in the atmospheric transparency windows despite of scattering in the atmosphere and clouds. Helbert et al. (2008), Mueller et al. (2008), and Gilmore et al. (2011a) retrieved the 1-µm emissivity of tessera terrain in Lada Terra and Alpha Regio form the VIRTIS data; they found that the tessera emissivity is different from the emissivity of the adjacent supposedly basaltic plains suggesting compositional difference. We will discuss their results later after description of our VMC-based results.

2.2. Chemical weathering

Surface materials on Venus most likely are involved in chemical interaction with atmospheric gases. Thermodynamic calculations (see e.g., Barsukov et al., 1980, 1982; Klose et al., 1992; Fegley, 2003; Zolotov, 2007) supported by still scarce modeling experiments (e.g., Fegley and Prinn, 1989; Johnson et al., 2003; Abbey et al., 2011) suggest several effects of chemical weathering on Venus, including: (1) oxidation and sulfurization of surface rocks through gas–solid-type reactions; (2) isochemochemical weathering of individual solid phases with respect to elements being nonvolatile at Venus’ surface temperature (e.g., Al, Si, Mg, Fe, Ca, Na); and (3) a strong altitude-dependent effect for the chemistry and physics of gas–surface interactions. Current hydration of anhydrous phases

is considered as unlikely and original hydrated phases (if any) would be dehydrated (Zolotov, 2007).

At plains elevations the expected mineral assemblage of weathered basalts includes (in order of decreasing abundance): plagioclase, clinoenstatite, pyrite or magnetite, anhydrite or diopside, microcline, and a few minor phases (Barsukov et al., 1982; Klose et al., 1992). Later considerations showed that redox conditions on the plains level of Venus are most probably close to coexistence of magnetite and hematite and thus pyrite cannot be stable there (Fegley et al., 1995a,b; Zotov, 1996, 2007). The major difference of the assemblage of weathered basalts from that of unweathered basalts is the expected presence of anhydrite (CaSO4) formed due to sulfurization of diopside (CaMgSi2O6) and anorthite component of plagioclase (CaAl2Si2O8) as well as presence of hematite (Fe2O3) due to oxidation of olivines and pyroxenes containing ferrous iron (e.g., Zotov, 2007). Presence of hematite and anhydrite in the weathered surface material of Venus was assumed in a recent paper of Smrekar et al. (2010). A degree of possible chemical weathering on Venus is unknown but it affects should be most prominent for the uppermost surface layer and thus potentially could influence the NIR emission, which we see in the VMC and VIRTIS images. Appearance of anhydrite may be noticeable in the VMC observations because of its high reflectivity and thus low emissivity at 1 μm wavelength (see below).

The chemical surface modification effect is directly seen in the results of the microwave remote sensing of Venus in the areas higher than some critical altitude, typically ~4 km above the mean planetary radius. These mountain tops, with a few exceptions, show very low microwave emissivity (and correspondingly very high radar reflectivity). These microwave emissivity anomalies are a subject of controversy and have been attributed to the temperature-controlled presence of conductive, semiconductive, ferroelectric or ferrimagnetic materials (e.g., Klose et al., 1992; Pettengill et al., 1997; Shepard et al., 1994; Starukhina and Kreslavski, 2002; Wood et al., 1997). Among the proposed variety of materials, the least exotic and the most plausible from physical sky, 2002; Wood et al., 1997). Among the proposed variety of materials, the least exotic and the most plausible from physical point of view are hematite, magnetite or pyrite. The 1-μm emissivity of these minerals, however, is rather close to that of unweathered basalts so the mountain tops mineralogy is probably not a promising target for the VMC image analysis.

There are no reliable data on how fast chemical surface modification works on Venus. Only one estimate (applicable to high altitudes) is available: Klose et al. (1992) noted that the top surface of very high (~9 km) Maat Mons volcano shows a significant decrease in microwave emissivity only in some places while most part of its summit has microwave emissivity close to the values typical for the plains. The authors of this work suggested that this is because this volcano is so young that only earliest lavas of it had enough exposure time to get chemically modified, while the majority of its lavas had not. We should keep in mind that microwave signature considered by Klose et al. (1992) is relevant to upper centimeters of the surface while the IR optical properties measured by VMC are relevant to upper microns and chemical alteration of the micron layer is obviously much faster than in the centimeter-thick layer.

The hypothesis of a very young age of Maat Mons volcano was then independently supported by the above mentioned studies of Basilevsky (1993) and Basilevsky and Head (2002b). They analyzed age relations of lavas of this volcano with the crater Uvassy having associated radar-dark parabola implying that at least part of lavas of this volcano formed less than a few tens of millions years ago, while the mean surface age of Venus is estimated to be of several hundred millions of years (McKinnon et al., 1997). The fact that all other high enough mountains on Venus have low emissivity tops suggests that they are old enough for the chemical modification to be developed. Some of these highs (Maxwell, Ovda) have ages somewhat older than the mean surface age of Venus, but majority of the highs are younger. For example, the Beta Regio rise has microwave low-emissivity tops and associated rifting in Devana Chasma shows evidence of activity more recent than 0.5 of the mean surface age (Basilevsky and Head, 2002b, 2007).

Pieters et al. (1986) measured spectra of hematite heated up to temperature of 500 °C and compared them with the data taken by the Venera-9 and -10 wide-angle spectrophotometer (Ekonomov et al., 1980). They concluded that the spectra of the surface materials at the landing sites resemble the spectra of hematite and hematite-bearing weathered basalt, rather than that of magnetite. Because hematite is not typical for unaltered basalts, this suggests that surface materials in these sites are weathered (oxidized).

The Venera 9 site is in the area with tectonic steep-sloped graben (Abrakhimov, 2001f) where down-slope mass wasting and thus rather effective resurfacing is logical to expect. The Venera 10 site is in the regional plains (Abrakhimov, 2001c) with almost absent steep slopes that suggests negligible role of the down-slope mass wasting and associated resurfacing. Thus, if observations of Ekonomov et al. (1980) are interpreted as indication on the chemically weathered surface in the Venera 9 and 10 sites, then one can conclude that the chemically weathered surface material is typical not only for plains (Venera 10) but for the areas with indications on down-slope mass wasting (Venera 9).

Wood et al. (1997) calculated how much sulfur could be present in the Venera-13, -14 and Vega-2 surface materials, if they would be totally weathered (sulfurized) basalts, and compared the result with the actually measured sulfur contents; he concluded that weathering is only partial: 50% for Vega-2 and 7–20% for Venera-13 and -14. The analyzed samples, each about 1 cm³, were taken by drilling from the depth down to ~3 cm. Thus, these estimates represent the mean weathering degree in a few centimeters thick surface layer, while for the very surface seen in NIR range, the weathering degree can be essentially higher.

The Venera-13 and -14 landers also measured electrical resistivity of surface soil and found it unexpectedly low: 89 and 73 Ω-m respectively (Kemurzdjian et al., 1983). A co-author of Kemurzdjian et al. (1983), V.V. Gromov (personal communication to J. Wood) ascribed “this low resistivity to the presence of a thin film of electrically conductive material on the soil particles” (Wood et al., 1997, p. 649). Presence of high content of magnetite can also lead to low resistivity. This is evidence of the weathering of surface material.

Thus, the observations in situ and their analysis suggest that the Venus surface material, especially its thin uppermost layer, is mineralogically modified, unless the material has been exposed for rather short time due to its recent emplacement or continuing resurfacing. The case of recent emplacement is probably exemplified by the summit of Maat Mons discussed above. The cases of ongoing resurfacing are probably associated with down-slope material movement, which in scale and intensity probably is most prominent on slopes of rift zones and walls of large impact craters. These slopes however have large range of altitudes at short horizontal distances, which makes them difficult to be analyzed with the VMC data. Besides, as it was mentioned above, observations by Ekonomov et al. (1980) in the place with the down-slope material movement (Venera-9 site) provide evidence of chemically weathered rather than unweathered material.

2.3. Eolian resurfacing

Volcanic and tectonic features on Venus may be affected by eolian resurfacing caused by normal “meteorological” winds (Greeley et al., 1997) and locally by strong winds, which are thought to accompany impact cratering events (Ivanov et al., 1992; Schultz, 1992). The eolian features observed in Magellan SAR images are...
represented by radar-dark mantles, wind streaks, yardangs and dunes. The first two types of eolian features are rather common on Venus, while the features of the second two types, large enough to be seen on the Magellan images, are observed only in a few localities. Yardangs indicate effective wind erosion but their rarity (Greeley et al., 1997) suggests that wind erosion does not play a great role among surface processes on Venus at the scale of features observed on the Magellan images.

For smaller features, however, deflation, eolian transport and deposition certainly play an important role. It is seen, in particular, in localization of the surface fines (considered to be a loose material) in local lows in between slab-like outcrops of the finely-layered rocks at the Venera 10, 13 and 14 landing sites (Florensky et al., 1977; Basilevsky et al., 1985). Deflation of loose fines was directly observed in panoramas taken by the Venera 13 lander. The three panoramas taken with 20 min time interval showed that a clod of dark fines thrown at the landing upon the supporting ring of the lander was shrunk with time to much smaller size. The only reasonable explanation of this observation is deflation of this fine material by the near-surface wind (Selivanov et al., 1983).

For our analysis we have a special interest in radar-dark mantles, which represent a veneer of fine-grained material covering the local bedrock. Radar-dark mantles are commonly seen in association with impact craters, forming halos of different sizes and forms. They look dark in SAR images because their surface is smooth (Campbell et al., 1992; Bondarenko and Head, 2009). The source of the radar-dark-mantle material is fine debris formed and lifted into atmosphere by crater-forming impacts and then deposited from the air. When the fine-grained ejecta material is at high levels of the atmosphere, it is driven by strong zonal winds and travels long distances before it is deposited on the surface and covers the local material. This is how the hundreds- to thousands-km-long radar-dark parabolas associated with young impact craters form (Campbell et al., 1992; Vervack and Melosh, 1992).

With time, the deposited material is being reworked by the near-surface winds. Its surface looses its smoothness and extended radar-dark parabolas shrink into smaller non-parabolic radar-dark haloes and then the haloes disappear. But significant part of the deposited material is essentially not moved far away and we probably see it in the TV panoramas taken by the Venera landers as the mentioned above slabs of fine-layered material (Florensky et al., 1977; Basilevsky et al., 1985). Deflation of loose fines was directly observed in panoramas taken by the Venera 13 lander. The three panoramas taken with 20 min time interval showed that a clod of dark fines thrown at the landing upon the supporting ring of the lander was shrunk with time to much smaller size. The only reasonable explanation of this observation is deflation of this fine material by the near-surface wind (Selivanov et al., 1983).

Table 1

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Reflectivity of different size (μm) fractions</th>
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<tr>
<td></td>
<td>25–63</td>
</tr>
<tr>
<td>Anorthite</td>
<td>0.70</td>
</tr>
<tr>
<td>Oligoclase</td>
<td>0.78</td>
</tr>
<tr>
<td>Orthoclase</td>
<td>0.72</td>
</tr>
<tr>
<td>Ortho-pyroxene Eo85</td>
<td>0.45</td>
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<tr>
<td>Olivine Fo90</td>
<td>0.48</td>
</tr>
</tbody>
</table>

In attempt to study mineral composition of Venus surface through analysis of the VMC data we, first, collected and analyzed spectral data on materials, which are potentially present there. These materials are: (1) basalts as major candidates for rocks composing dominant part of venusian surface and (2) more silicic rocks, rhyolites and andesites, which could compose tessera terrain, as well as some minerals. Among the latter are some components of basalts, minerals, which are considered to be products of chemical weathering on Venus surface, and some minerals hypothesized to be present on the low-microwave-emissivity mountain tops. So we collected spectral data on Ca-rich plagioclases anorthite and labradorite, as well as on anhydrite, hematite, magnetite and pyrite. We used reflectance spectra from the ASTER spectral library (http://spec.lpl.nasa.gov, Baldridge et al., 2009) and Brown University Keck/NASA Relab Spectra Catalog (http://www.planetary.brown.edu/relab/) to estimate emissivity values of these materials at 1 μm. Due to the absence of laboratory data on mineral and rock emissivity (ε) in the near-infrared spectral range, we had to use 1-μm reflectivity (R) and apply Kirchhoff’s law (ε = 1 − R). For R we use all types of reflectances (bidirectional, hemispherical and biconical), which is not rigorously grounded, and our quantitative results on ε should be treated with caution, while the trends are qualitatively reliable.

Note that the material reflectivities/emissivities at the considered wavelength depend not only on the material composition and temperatures but also on the particle size. For silicates and common rocks, whose optical properties at 1 μm are controlled by volume scattering, reflectivities at 1 μm typically increase with decreasing particle size (see Table 1). The opposite trend or the lack of particle size dependence are typical of highly absorbing materials, e.g., magnetite and sulfides.

As follows from the data collected by us, a negative correlation between 1-μm reflectivity and particle size is typical for basalts, to be from 2 to 10 m s⁻¹ (see summaries in Schubert et al. (1980), Moroz, 1981, and Kerzhanovich et al., 1983). So the vertical gradient of wind velocity is probably from a few dm s⁻¹ to about 1 m s⁻¹ per kilometer of altitude and we may expect that on the surface of high-standing landforms the wind velocity is noticeably higher than at lows so the smaller particles may be suspended and blown to the lowland (Leeder, 2007).

Besides, the mentioned above strong winds, which are believed to accompany impact cratering events (Ivanov et al., 1992; Schultz, 1992) might episodically strip out and suspend in the air the loose surface material which would not be mobilized by normal “meteorologic” winds. Taking in mind that for the morphologically observed part history of Venus, these events (~1000) together with blasts of meteoroids in the lower atmosphere (responsible for formation of ”splotches”) could play a noticeable role in the mobilization and redistribution of fines on the surface of this planet.

2.4. Reflectivity/emissivity of potential surface materials at 1 μm wavelength

In attempt to study mineral composition of Venus surface through analysis of the VMC data we, first, collected and analyzed spectral data on materials, which are potentially present there. These materials are: (1) basalts as major candidates for rocks composing dominant part of venusian surface and (2) more silicic rocks, rhyolites and andesites, which could compose tessera terrain, as well as some minerals. Among the latter are some components of basalts, minerals, which are considered to be products of chemical weathering on Venus surface, and some minerals hypothesized to be present on the low-microwave-emissivity mountain tops. So we collected spectral data on Ca-rich plagioclases anorthite and labradorite, as well as on anhydrite, hematite, magnetite and pyrite. We used reflectance spectra from the ASTER spectral library (http://spec.lpl.nasa.gov, Baldridge et al., 2009) and Brown University Keck/NASA Relab Spectra Catalog (http://www.planetary.brown.edu/relab/) to estimate emissivity values of these materials at 1 μm. Due to the absence of laboratory data on mineral and rock emissivity (ε) in the near-infrared spectral range, we had to use 1-μm reflectivity (R) and apply Kirchhoff’s law (ε = 1 − R). For R we use all types of reflectances (bidirectional, hemispherical and biconical), which is not rigorously grounded, and our quantitative results on ε should be treated with caution, while the trends are qualitatively reliable.

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As follows from the data collected by us, a negative correlation between 1-μm reflectivity and particle size is typical for basalts,
rhyolites and andesites, as well as for anorthite, labradorite, and anhydrite. For example, the mean value of the 1-μm reflectivity of fine-grained basalts is ~0.2 while for the coarse-grained separates and whole rock chips it is 2–3 times lower. Hematite reflectivity at 1 μm is also higher for the finer size fractions and rather high (up to 0.4) for nanophase synthetic powders (Baldridge et al., 2009; Morris et al., 1985), meanwhile the 1-μm reflectivity of magnetite for finer size fractions is slightly lower than for the coarser ones, while pyrite shows no prominent dependence of this sort.

As mentioned above, rocks and minerals on the surface of Venus are probably chemically weathered. This weathering happens through the surface–atmosphere interaction in the oxidizing environment (e.g. Zolotov, 2007) at not very high temperature and very low partial pressure of H2O. In this environment, the major rock-forming components (Na, Si, Mg, Al, Ca, K, Fe) do not form volatile species (e.g. Wood et al., 1997). So the primary minerals and glasses decompose and recrystallize in situ with or without addition of oxygen and/or sulfur dioxide from the atmosphere into new mineral assemblages (e.g. Zolotov, 2007). So we assume that surface materials on Venus are typically fine-grained, and for the materials of our interest we consider mostly 1-μm reflectivity of fine (<25 to <5 μm) separates. Other researchers (e.g. Smrekar et al., 2010) also consider that weathering products are fine-grained. Very young lavas, however, probably have rough unweathered surfaces and thus should show lower 1-μm reflectivity. And for the high-standing landforms we may suspect eolian removal of the fine-grained fractions so the residual fractions may be coarser and thus have lower reflectivity/higher emissivity.

The acquired from the above mentioned spectral libraries mean values of 1-μm reflectivity for fresh and weathered (oxidized) basalt, as well as for rhyolite, andesite, anorthite, labradorite, anhydrite, hematite, magnetite and pyrite are given in the “room temperature” line of Table 2. The table shows that relatively dark in visual range fine-grained powders of fresh and weathered basalt, hematite, magnetite and pyrite are also dark at 1 μm, while visually brighter fine-grained powders of rhyolite, andesite, anorthite, labradorite and anhydride are also brighter at 1 μm. Here are also given mean values of calculated emissivity (ε = 1 – R).

It was found in several works that for common rock-forming minerals reflectance spectra, in general, and reflectivity at some wavelengths, in particular, may significantly change with temperature (e.g. Singer and Roush, 1985; Pieters et al., 1986; Roush and Singer, 1986; Moroz et al., 2000). These changes can be due to temperature-dependent change of amplitude of the thermal vibrations of absorbing cations about the centers of their coordination sites, resulting in widening an absorption band as the temperature increases (Burns, 1970). An increase in temperature may also change bond lengths between cations and surrounding ligands, resulting in wavelength shifts of electronic absorption bands. These shifts would affect reflectivity values at band wings. Since we are interested in the reflectance values at 1-μm, minerals with absorptions band wings at this wavelength (notably low-Ca pyroxenes) would be especially affected by temperature.

Hinrichs and Lacey (2002) showed that at the temperature increase from 80 K (~193 °C) to 400 K (127 °C), the 1 μm reflectivity changes from 0.43 to 0.2 for orthopyroxene En86, from 0.265 to 0.225 for eucrite EETB3551, and from 0.125 to 0.09 for mature basaltic lunar soil 12023. The changes are correspondingly 54%, 15% and 28%. This temperature dependence of 1-μm reflectivity is just an example and cannot be applied to all pyroxenes and basaltic rocks. Basaltic rocks may vary in composition and contents and sizes of opaque grains. 1-μm reflectivity of basalts containing low-Ca pyroxenes and/or poor in fine-grained opaques should depend on temperature more significantly compared to basalts enriched in high-Ca clinopyroxenes, olivines, and/or oxides (see e.g. Burns, 1993).

Based on the results of Hinrichs and Lucey (2002) one can suggest that for basalts the decrease of the 1 μm reflectance due to temperature increase from 80 K to 400 K (ΔT = 320 K) may be ~20%. As a very rough guess we can suggest that for basaltic materials, the temperature increase from the room temperature to the Venus surface temperature (ΔT = 500 K) could lead to the 1 μm reflectance decrease by maximum 30%. Pyrite and magnetite show electronic absorption bands centered at 1 μm. If these bands do not significantly shift at high temperatures, we suggest that their 1-μm reflectivity does not change significantly. This is not the case for hematite, which shows an absorption band at 0.85 μm, so that the 1-μm reflectivity (long wavelength wing of the band) can significantly decrease at 500 °C. If the band does not shift with temperature, based on the data of Pieters et al. (1986) for wavelengths shorter than 0.8 μm we can roughly estimate that the 1-μm reflectivity decrease by maximum 30% may be expected at 500 °C. For minerals and rocks having low iron content and stable at 500 °C the effects of Venus temperature on 1-μm reflectivity are probably minor or negligible. Here we assume that temperature dependence of 1-μm emissivity does not differ from that of 1 – R. Our estimates for 1-μm reflectivity/emissivity of materials expected on Venus surface for the room temperature and the Venus environment are summarized in Table 2.

The data and estimates given in Table 2 show that fresh and weathered basalts have rather close high temperature emissivity

<table>
<thead>
<tr>
<th>Material</th>
<th>Samples</th>
<th>The 1 μm reflectivity/emissivity</th>
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<tr>
<td>Basalt fresh</td>
<td>13 samples: Kilauea, samples H1, 2, 5, 7, 9, 10; Mariana isl., 4F; C. Massif, France</td>
<td>Room temperature: 0.2/0.8 Venus surface: 0.15/0.85</td>
</tr>
<tr>
<td>Basalt oxidized</td>
<td>1 sample: C. Massif, France</td>
<td>Room temperature: 0.3/0.7 Venus surface: 0.2/0.8</td>
</tr>
<tr>
<td>Rhyolite</td>
<td>1 sample: Custer Co</td>
<td>Room temperature: 0.5/0.5 Venus surface: 0.5/0.5</td>
</tr>
<tr>
<td>Andesite</td>
<td>2 samples: andesite1f, andesite2f</td>
<td>Room temperature: 0.3/0.7 Venus surface: 0.3/0.7</td>
</tr>
<tr>
<td>Anorthite</td>
<td>1 sample: TSO3AF</td>
<td>Room temperature: 0.7/0.3 Venus surface: 0.7/0.3</td>
</tr>
<tr>
<td>Labradorite</td>
<td>3 samples: Lake Co, Oregon, Nain, Labrador</td>
<td>Room temperature: 0.7/0.3 Venus surface: 0.7/0.3</td>
</tr>
<tr>
<td>Anhydrite</td>
<td>6 samples: Birmingham, Boise, Mesab, unknown location</td>
<td>Room temperature: 0.7/0.3 Venus surface: 0.7/0.3</td>
</tr>
<tr>
<td>Hematite</td>
<td>2 samples, Nova Scotia, Ajo</td>
<td>Room temperature: 0.2/0.8 Venus surface: 0.15/0.85</td>
</tr>
<tr>
<td>Magnetite</td>
<td>20 samples: Utah, Langesundfjord, Balmet, San Benito-2 samples, San Bernard, Hop Spring, Grangesberg, Oka, Gladhammar, Marquette, Barrio, Mother Lode, Eve Washington, Glacier MP, Nye, Iron Co, Putnam, Sussex</td>
<td>Room temperature: 0.05/0.95 Venus surface: 0.05/0.95</td>
</tr>
<tr>
<td>Pyrite</td>
<td>4 samples: Rio Mariana, Rouyn District, Austin, Alberta</td>
<td>Room temperature: 0.12/0.88 Venus surface: 0.12/0.88</td>
</tr>
</tbody>
</table>

and these are values expected for venusian plains and for majority of volcanic constructs. Materials expected to be present on the mountain tops (iron oxides and sulfides) have the high-temperature emissivity (~0.9) only slightly higher than basalts so probably they cannot be distinguished from basalts in the VMC image analysis. Materials which could compose tesserae (rhyolites, andesites as well as anorthitic and labradoritic anorthosites) all except andesites probably have high-temperature emissivity significantly lower (0.5 and less) than those of basalts and this gives hope to find in further analysis if tesserae are basaltic or not. Comparison of Tables 1 and 2 shows, however, that the emissivity dependence on grain size could be more significant that the dependence on mineralogy and this makes the analysis even more difficult (see also Helbert et al., 2008).

3. VMC observations

Venus Monitoring Camera performs wide-angle observations of Venus in four narrow band filters sharing a single CCD (Markiewicz et al., 2007, 2008). On the night side of the planet, VMC maps thermal emission of the surface in the 1.01 μm spectral transparency “window”. These measurements are at the limit of instrument capability. Even at the maximum exposure of 30 s the faintness of the surface emission and low efficiency of the CCD detector (~2%) at 1.01 μm result in that the measured signal does not exceed ~200 digital units (DNs), which is ~3% of the CCD full well. The second difficulty of the surface observations results from the solar stray light. In order to cope with this problem, VMC observes the night side when the spacecraft is in eclipse. This limits the observations to low latitudes (±40°). These latitudes are, however, poorly covered by VIRTIS (Mueller et al., 2008) that makes both experiments highly complementary. Fig. 1 shows examples of the VMC night side images captured in the 1.01 μm filter.

VMC acquired thousands images of the Venus night side during the time of observation, but here we consider only results of several orbits for two areas. One is an area SW of Beta Regio, where we try to understand if there is any difference in 1-μm emissivity between the massif of tessera terrain and the surrounding plains of presumably basaltic composition. Because VMC does not have a separate spectral channel for deducing the optical thickness of clouds, we select orbits that are relatively free from clouds opacity variations. These orbits were found by forming the ratio of images from two consecutive orbits (transformed into the same projection). For images in the same projection surface features are in same places and thus all contrasts on images of ratio are due to changes in atmospheric properties. Absence of contrast means absence of clouds opacity changes from one orbit to another. Because of strong winds in the venusian atmosphere it is very likely that the absence of temporal variations (on 24 h basis that is the time span between two consecutive orbits) means also absence of spatial variations on ~1000 km basis. However, it is still possible that distribution of clouds variations would be stable for 24 h and in that case we will interpret them as emissivity variations. Estimates of errors caused by the clouds opacity variations for the particular images used in this work please see below in Section 5.1.

4. Calculation of the surface emissivity maps

To obtain maps of the surface emissivity the VMC observations must be compared to the model images. Due to unknown cloud
opacity and uncertainties in the VMC radiometric calibration we normalized the images by the value at a reference location, where the surface was assumed to be of basaltic composition.

To obtain synthetic images it is necessary to model surface emission and atmospheric effects. The dense atmosphere affects surface images in two ways. It strongly attenuates the emission through scattering back to the surface and gaseous absorption and leads to degradation of spatial resolution, or blurring. Thus, assuming that both surface emissivity and atmospheric transmittance do not strongly vary within the scale of blurring function (\(a \sim 100\) km), emission intensity at a point with horizontal coordinates \(x, y\) at the top of the atmosphere can be expressed by the formula:

\[
I(x, y) = \frac{t(x, y)e(x, y)}{1 - (r(x, y) - e(x, y))} \int \int B[T(x', y')] |F(x - x', y - y)| dx' dy'
\]

(1)

where \(t(x, y)\) is the atmospheric transmittance, \(r(x, y)\) is the atmospheric reflectance of surface radiation in backward direction (both depend on surface altitude), \(e(x, y)\) is the emissivity distribution of Lambertian surface, \(B[T]\) is the Planck function of the surface temperature \(T_s\), and \(F\) is the blurring function. In this formula we applied the two-stream approximation to a single layer atmosphere to account for attenuation; convolution with the blurring function describes smoothing contrasts. We note that in our model both atmospheric transmittance and reflectance depend on surface topography. Mueller et al. (2008) used a similar approach to analyze atmospheric transmittance and reflectance depend on surface topography. The atmospheric blurring effect was modeled by the Monte-Carlo method (Metropolis and Ulam, 1949) taking into account multiple scattering by aerosols and gases. In the Monte-Carlo method scattering and emission processes are described by their probabilities calculated for a certain model of the atmosphere. Here we used the vertical structure of clouds and their optical properties from Tomasko et al. (1985). Henyey-Greenstein phase function with asymmetry parameter \(g = 0.78\) and single scattering albedo \(\omega_0 = 0.9995\), and Rayleigh scattering coefficient in the lower atmosphere from Moroz (2002). Thermal emission of the surface is a product of the Planck function that strongly depends on surface temperature and therefore on altitude, and surface emissivity \(e\) defined by mineralogical composition and surface material grain size. Topography related variations of the atmospheric absorption were included in the transmittance. For a particular surface point we can express corrected coefficient \(t\) as \(t_0 \cdot \exp(kH)\), where \(H\) is surface altitude. To calculate gas absorption, we used approach as in Ignatiev et al. (2009) (that uses the same form-factors for line wings as in Meadows and Crisp (1996)). Radiative transfer model for this calculations is based on the DISORT code (Stamnes et al., 1988) and line parameters from the preliminary version of the Carbon Dioxide Spectroscopic Database (CDSDB) for Venus, the CO\(_2\) high-temperature database, and HITRAN. More detailed description and references are given in Ignatiev et al. (2009). These calculations gave \(a = 1.0034\), and \(k = 0.0317\) km\(^{-1}\). Emissivity, obtained with these parameters and lapse −8.1 K km\(^{-1}\) shows linear correlation with altitude, that does not look realistic. Therefore, we adjusted the value of \(k\) to achieve absence of correlation for plains terrains (≈0.12 km\(^{-1}\)). The correlation diagram for emissivity–altitude obtained with these parameters for all orbit 470 is shown in Fig. 2. One can see that major part of points does not show correlation of emissivity with surface altitude.

The Monte-Carlo algorithm was implemented in a multi-threaded C++ code. We used scalar presentation of the scattering matrix. In the calculations the test photon packets were inserted at the bottom of the atmosphere simulating emitting surface and were collected by a “detector” at the cloud top. The modeling usually included about 16,000 photon packets in each of 1000 detector cells (4 by 4 km in size) that corresponded to the same area on the surface due to orthographic projection.

The Monte Carlo radiative transfer simulations were used to determine the atmospheric blurring function (a synthetic VMC image of a point source on the surface). The modeling gave the blurring function \(F(x, y)\) with half-width of ~50 km, which is in agreement with both apparent blurring of VMC images and Hashimoto and Imamura (2001) results (Fig. 3). The difference between our blurring functions and the one from Hashimoto and Imamura (2001) could be caused by using the different phase function of atmospheric particles and different cloud models. Atmospheric reflectance \(r\) and transmittance \(t\) for selected atmosphere model were obtained from same simulations and their values are \(r = 0.76\) and \(t = 0.21\), respectively for zero surface altitude. Also these calculations were used to check if outgoing flux on the top of the atmosphere is orthotropic, because Eq. (1) is valid only in that case.

To calculate synthetic VMC images we used the Magellan topography derived from Magellan Radar Altimeter (Ford and Pettengill, 1992). The topography data were converted into the maps of temperature and surface brightness distribution assuming thermal equilibrium with the atmosphere, constant lapse rate of −8.1 K km\(^{-1}\) (Seiff et al., 1985) and constant emissivity (exact value does not matter because of further normalization). Then synthetic VMC images were obtained by convolving the surface brightness distribution with the blurring function (Eq. (1), Fig. 3). In order to get rid of uncertainties in the VMC absolute calibration and cloud opacity, we normalized the measured images divid-

---

ing them by the brightness at a reference location individually se-
slected for each mosaic. From Eq. (1) we can derive the following
expression for the VMC normalized image $V$:

$$V(x, y) = \frac{e_0(1 - r(x, y))}{e_0(1 - e_0) + r(x, y)}$$

where $e_0$ is the assumed surface emissivity at the reference location $(x_0, y_0)$.

In addition, we considered a model case with constant surface emissivity. The expression for normalized model image is derived in a similar way:

$$M(x, y) = \frac{\int \int B[T_s(x', y')] \cdot F(x - x', y - y') dx' dy'}{\int \int B[T_s(x', y')] \cdot F(x_0 - x', y_0 - y') dx' dy'}$$

All contrasts in the model image are due to temperature differences and not emissivity variations. From Eqs. (2) and (3) we derive the following expression for the unknown surface emissivity distribution:

$$e(x, y) = \frac{R(x, y)e_0(1 - r(x, y))}{1 + r(x, y)(e_0(1 - R(x, y)) - 1)}$$

where

$$R(x, y) = \frac{V(x, y)}{M(x, y)}$$

Thus Eq. (4) allows us to derive spatial distribution of the surface emissivity from the ratio of the normalized VMC and model images and assumed emissivity $e_0$ at a reference location. We make two remarks on Eq. (4). First, it is applicable only to emissivity variations of spatial scale greater than the full width of the blurring function (∼100 km, or ∼10 VMC pixels), which holds for large-scale surface features. Second, the distance between the reference site and the place where we determine emissivity should not exceed typical scale of deep cloud inhomogeneities (∼1000 km). Both conditions are met in the areas analyzed in this paper.

![Maps of the study area SW of Beta Regio (26°S–38.5°N, 243°–298°E). (a) Magellan SAR image; (b) Magellan topographic map (highs are bright); (c) Magellan map of microwave emissivity (brighter shades denote higher emissivity); and (d) simplified geologic map of the area: P – plains, T – tessera terrain, R – rifts, black spots – young lavas, LE – low radar emissivity deposits, Chi – Chimon-mana Tessera, Tuu – Tuulikki volcano.](image)
5. The geologic analysis of the VMC data

5.1. Study areas

The major objective that is pursued by this study is to test, if tessera terrain material is different in its chemical/mineralogical composition from the surrounding plains, which as was said above are considered to be basaltic. We approached this objective not globally but within rather small region which has the appropriate objects of the study and is well covered by the VMC images. This is the area SW of Beta Regio (Fig. 4). Here there is a relatively small but distinct massif of tessera terrain, Chimon-mana Tessera, the surface emissivity of which we try to determine and compare with that of the adjacent plains. About 1000 km to the north, among the plains, there is a relatively small volcano, Tuulikki Mons, whose morphology (gentle slopes and extended outskirts of lava flows) are indicative of basaltic composition (e.g., Head et al., 1992). Its presence in the study area is important for our analysis because its summit stands about 0.5–1 km above the plains, as the summit portion of Chimon-mana tessera does. Thus, we can eliminate the altitude effect and try to search for the effects of surface composition or texture.

For this area we considered results of VMC imaging campaigns carried out in 2007: #4 (March–April), #5 (July–August) and #6 (November) assembled into three different mosaics. Mosaics taken during the seasons 4 and 5 are significantly overlapping and rather consistent with each other and with model images. What is seen on these VMC mosaics is also consistent with what is seen on images taken by VIRTIS in the area, where the VMC and VIRTIS coverage overlap (Mueller et al., 2008). The mosaic for season 6 differs from the mosaics for seasons 4 and 5 in many details. Now we are analyzing the nature of these differences. The analysis given below is made using the two visually consistent mosaics from seasons 4 and 5. Quantitative emissivity measurements presented in Section 5.2 below are obtained using orbit-wise mosaics, built from individual images taken from the orbit 0470 (2007–08–04). From Eq. (4) and ratios of mosaic for orbit 470 to orbits 46x and 47x we estimate clouds opacity variations across studying area to be not more than 10% and most likely smaller. This gives errors in emissivity 10–20% (assuming that all other parameters in Eq. (4) are constant).

5.2. Search for compositional difference among the studied units

We compare (Fig. 5) the central part of Chimon-mana Tessera (unit 1) against surrounding regional plains (unit 2), divided into subunit 2n (northern plains) and 2s (southern plains), and relatively young Tuulikki Mons volcano (unit 3) and its summit part (unit 4) against its surrounding regional plains (unit 5). The unit altitudes are given in Table 3.

As it was said above, calculation of 1-μm emissivity e from the observed thermal emission requires two assumed model parameters: temperature lapse rate L and reference surface emissivity e0. For each pixel we calculated e for two values of the reference emissivity e0 = 0.8 and e0 = 0.58. The values L = −8.1 K km⁻¹ and e0 = 0.8 have been used in a number of previous publications (e.g., Meadows and Crisp, 1996), while e0 = 0.58 have been used by Smrekar et al. (2010).

Before discussing the calculation results, it is worthwhile to consider effects of different assumptions. The decrease of the assumed reference surface emissivity e0 from 0.8 to 0.58 should “proportionally” reduce calculated emissivity of all terrains.

The decrease of the assumed lapse rate (for example from −8 K km⁻¹ to −5 K km⁻¹) leads to a hotter model temperature of the high-standing landforms and thus to a higher 1-μm model emission, which in turn leads to lower emissivity of the high-standing landforms calculated from the comparison of the model and measured (by VMC or other instrument) data (Fig. 6a). The increase of the assumed lapse rate leads to the opposite effect, that is to a higher calculated emissivity of the high-standing landforms (Fig. 6b). This is true only if the studied landform is a kind of plateau, wide enough in comparison to the width of the blurring function.

If the studied landform is not a plateau, but a mountain with prominent summit, then at any lapse rate the summit due to blurring of the IR emission by the scattering in the clouds will not be seen as cold as it is. Because of averaging with hotter pixels surrounding the summit, the latter on the model image would appear hotter, its model emission will be higher. If the model blurring radius is smaller than that of natural smoothing due to scattering in the cloud layer, then the model emission of the summit will be lower than its measured emission and this will result in higher calculated emissivity of the mountain (Fig. 6c). If the model blurring...
radius is larger than that of natural smoothing, then the model emission of the summit will be higher than its measured emission and this will result in lower calculated emissivity of the mountain (Fig. 6d).

This effect is illustrated by calculation of model images of the surface emission done for real topography of Tuulikki Mons vicinity and area south of it. The topography of this area is characterized by presence of Tuulikki Mons (relatively large mountain with summit depression in the upper right of the image), two smaller mountains SSW of it and a crater south of Tuulikki (Fig. 6e). Fig. 6f shows the ratio of the surface emission images calculated for the blurring function diameter =50 km (numerator) and for diameter =100 km (denominator). It is seen on Fig. 6f that decrease of the blurring function radius leads to decrease of model emission (comparing to natural blurring) for the mountain tops and this should lead to their higher calculated emissivity. For the crater (antimountain) the effect is opposite.

The combined effect of higher/lower lapse rate and the degree of the model blurring may lead to significant differences in the calculated emissivity of the studied landforms.

Table 3
Surface features under study (see Fig. 5), their altitudes, area and virtual points count (see below).

<table>
<thead>
<tr>
<th>Surface type</th>
<th>Mean altitude and its std. dev., km</th>
<th>Virtual points count</th>
<th>Area, $10^3$ km$^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chimon-mana, unit 1</td>
<td>$0.6 \pm 0.4$</td>
<td>103</td>
<td>142</td>
</tr>
<tr>
<td>Plains around Chimon-mana, unit 2</td>
<td>$-0.22 \pm 0.3$</td>
<td>337</td>
<td>605</td>
</tr>
<tr>
<td>Plains to the north from Chimon-mana, unit 2n</td>
<td>$-0.4 \pm 0.3$</td>
<td>180</td>
<td>323</td>
</tr>
<tr>
<td>Plains to the very north from Chimon-mana (northern part), unit 2nn</td>
<td>$-0.6 \pm 0.2$</td>
<td>98</td>
<td>175</td>
</tr>
<tr>
<td>Plains to the south from Chimon-mana, unit 2s</td>
<td>$-0.2 \pm 0.3$</td>
<td>75</td>
<td>148</td>
</tr>
<tr>
<td>Tuulikki middle, unit 3</td>
<td>$-0.2 \pm 0.3$</td>
<td>47</td>
<td>90</td>
</tr>
<tr>
<td>Tuulikki top, unit 4</td>
<td>$0.8 \pm 0.4$</td>
<td>16</td>
<td>33</td>
</tr>
<tr>
<td>Plains around Tuulikki, unit 5</td>
<td>$-0.4 \pm 0.2$</td>
<td>144</td>
<td>260</td>
</tr>
</tbody>
</table>

Fig. 6. Dependence of measured emissivity on the atmosphere temperature lapse (a and b) and the blurring function radius (c and d). The triangles on the diagrams visualize the sign of the effect: the triangle tip up indicates the emissivity increase, the tip down, the emissivity decrease. As an example: topography map of area $1.5^5\degree$N–$13^\circ$N, $268^\circ$E–$278^\circ$E (e), and ratio of surface emission for model with 50 and 100 km blurring function width (f).
We calculated the mean $e$ and estimation of its standard deviation for each unit (Table 5). To assess significance of the observed differences in the mean $e$ we applied Welch’s test (Welch, 1947) for the unit pairs of interest. The atmosphere blurring makes our effective spatial resolution to be $\sim 100$ km, which is much larger than a formal field of view of the VMC pixel. So, one cannot consider a value of each pixel as single and independent measurement. To correct this situation the study surface was “paved” with subareas of 100 km across. The number of such “tiles” on each unit (virtual points in Table 3) was taken as the number of measurements for the test. The results of the estimates are given in Table 5.

It is seen from Table 5 that, as expected, the use of surface reference emissivity $e_0 = 0.58$ has lead to the decrease of all calculated $e$ and differences between emissivities of different features. Most cases do show significant differences between selected units (Table 3); we discuss this below.

5.2.1. Plains units vs. plains units variabilities

Before we compare the 1-μm emissivities of Chimon-mana Tessera and the plains let us look what are emissivities of the plains. As it is seen from Table 5 and Fig. 5c in the study area they are variable.

The plains units around Chimon-mana (units 2n and 2s) and Tuulikki Mons (unit 5) were outlined as bands surrounding these two landforms keeping the total widths of these two areas to be about 1000 km. Then from analysis of the map of calculated emissivity (Fig. 5c) we divided unit 2n into two subunits (2nn and 2ns) which are noticeably different in their surface emissivities. The plains north from Chimon-mana Tessera (unit 2n) have significantly higher emissivity than the plains to the south (unit 2s). Statistically significant difference is observed for both the lr8-e08 and lr8-e058 models for all identified units and subunits of the Chimon-mana area Table 3.

The emissivity of plains around Tuulikki volcano (unit 5) is lower than that of units 2n and 2nn but higher than that of units 2s and 2ns. The unit 5 is 500–1000 km north of unit 2n and 1000–1500 km north of unit 2s, so potential variability in the clouds’ density at these distances makes comparisons of unit 5 with units 2n including 2nn and 2ns and 2s less reliable than for the case unit 2n vs. unit 2s, but probably acceptable. If we apply statistical estimates to these comparisons then the mentioned differences between unit 5 and units 2n and 2nn are statistically significant and between unit 5 and units 2s and 2ns are insignificant for both the lr8-e08 and lr8-e058 models. So the units 2n, 2s and 5 have different 1-μm emissivities (unit 2nn, the highest and unit 2s, the lowest) and the differences in most cases are statistically significant.

In attempt to understand the potential nature of these differences we put outlines of all the mapped units on the Magellan

---

**Table 4**

Description of used models.

<table>
<thead>
<tr>
<th>Model name</th>
<th>Model description</th>
</tr>
</thead>
<tbody>
<tr>
<td>lr8-e08</td>
<td>$T$ = 8.1, $e_0 = 0.8$</td>
</tr>
<tr>
<td>lr8-e058</td>
<td>$T$ = 8.1, $e_0 = 0.58$</td>
</tr>
</tbody>
</table>

**Table 5**

Comparison of emissivities of different surface types. Shaded lines correspond to combinations where inferred emissivity difference is formally statistically significant. See Table 4 for models and Table 3 for units description.

<table>
<thead>
<tr>
<th>Unit A/Unit B</th>
<th>$\epsilon \pm \text{std. dev}$</th>
<th>Difference at 0.05 level</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unit A</td>
<td>Unit B</td>
<td></td>
</tr>
<tr>
<td>Model lr8-e08</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1/2</td>
<td>$0.55 \pm 0.37$</td>
<td>$0.56 \pm 0.31$</td>
</tr>
<tr>
<td>1/2n</td>
<td>$0.55 \pm 0.37$</td>
<td>$0.64 \pm 0.24$</td>
</tr>
<tr>
<td>1/2s</td>
<td>$0.55 \pm 0.37$</td>
<td>$0.47 \pm 0.35$</td>
</tr>
<tr>
<td>2n/2s</td>
<td>$0.64 \pm 0.24$</td>
<td>$0.47 \pm 0.35$</td>
</tr>
<tr>
<td>1/2n</td>
<td>$0.55 \pm 0.37$</td>
<td>$0.76 \pm 0.15$</td>
</tr>
<tr>
<td>1/2ns</td>
<td>$0.55 \pm 0.37$</td>
<td>$0.50 \pm 0.27$</td>
</tr>
<tr>
<td>2n/2s</td>
<td>$0.76 \pm 0.15$</td>
<td>$0.50 \pm 0.27$</td>
</tr>
<tr>
<td>2n/2s</td>
<td>$0.64 \pm 0.24$</td>
<td>$0.76 \pm 0.15$</td>
</tr>
<tr>
<td>2n/2s</td>
<td>$0.64 \pm 0.24$</td>
<td>$0.50 \pm 0.27$</td>
</tr>
<tr>
<td>2n/2s</td>
<td>$0.63 \pm 0.07$</td>
<td>$0.55 \pm 0.04$</td>
</tr>
<tr>
<td>3/4</td>
<td>$0.63 \pm 0.07$</td>
<td>$0.53 \pm 0.45$</td>
</tr>
<tr>
<td>3/5</td>
<td>$0.55 \pm 0.04$</td>
<td>$0.53 \pm 0.45$</td>
</tr>
<tr>
<td>4/5</td>
<td>$0.64 \pm 0.24$</td>
<td>$0.53 \pm 0.45$</td>
</tr>
<tr>
<td>2n/5</td>
<td>$0.47 \pm 0.35$</td>
<td>$0.53 \pm 0.45$</td>
</tr>
<tr>
<td>2nn/5</td>
<td>$0.76 \pm 0.15$</td>
<td>$0.53 \pm 0.45$</td>
</tr>
<tr>
<td>2nn/5</td>
<td>$0.50 \pm 0.27$</td>
<td>$0.53 \pm 0.45$</td>
</tr>
</tbody>
</table>

Model lr8-e058

<table>
<thead>
<tr>
<th>Unit A/Unit B</th>
<th>$\epsilon \pm \text{std. dev}$</th>
<th>Difference at 0.05 level</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unit A</td>
<td>Unit B</td>
<td></td>
</tr>
<tr>
<td>Model lr8-e058</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1/2</td>
<td>$0.43 \pm 0.03$</td>
<td>$0.42 \pm 0.27$</td>
</tr>
<tr>
<td>1/2n</td>
<td>$0.43 \pm 0.03$</td>
<td>$0.48 \pm 0.19$</td>
</tr>
<tr>
<td>1/2s</td>
<td>$0.43 \pm 0.03$</td>
<td>$0.35 \pm 0.32$</td>
</tr>
<tr>
<td>2n/2s</td>
<td>$0.48 \pm 0.19$</td>
<td>$0.35 \pm 0.32$</td>
</tr>
<tr>
<td>1/2n</td>
<td>$0.43 \pm 0.03$</td>
<td>$0.55 \pm 0.11$</td>
</tr>
<tr>
<td>1/2ns</td>
<td>$0.43 \pm 0.03$</td>
<td>$0.38 \pm 0.24$</td>
</tr>
<tr>
<td>2n/2s</td>
<td>$0.55 \pm 0.11$</td>
<td>$0.38 \pm 0.24$</td>
</tr>
<tr>
<td>2n/2s</td>
<td>$0.48 \pm 0.19$</td>
<td>$0.55 \pm 0.11$</td>
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<tr>
<td>2n/2s</td>
<td>$0.48 \pm 0.19$</td>
<td>$0.38 \pm 0.24$</td>
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<tr>
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<td>$0.43 \pm 0.02$</td>
</tr>
<tr>
<td>3/5</td>
<td>$0.48 \pm 0.04$</td>
<td>$0.38 \pm 0.04$</td>
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<tr>
<td>4/5</td>
<td>$0.43 \pm 0.02$</td>
<td>$0.38 \pm 0.04$</td>
</tr>
<tr>
<td>2n/5</td>
<td>$0.48 \pm 0.19$</td>
<td>$0.38 \pm 0.04$</td>
</tr>
<tr>
<td>2n/5</td>
<td>$0.35 \pm 0.32$</td>
<td>$0.38 \pm 0.04$</td>
</tr>
<tr>
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</tr>
<tr>
<td>2nn/5</td>
<td>$0.38 \pm 0.24$</td>
<td>$0.38 \pm 0.04$</td>
</tr>
</tbody>
</table>
images (Fig. 7). It is seen in this figure that plains of the study area are geologically variegated. On the background of radar-dark regional plains are patches of radar-bright ones and the younger volcanic centers with radially spreading lobate flows, also radar-bright. One of such volcanic centers is Tuulikki Mons volcano, which will be discussed below. Unit 2n is the most abundant in these radar-bright spots (the most abundant is subunit 2nn), unit 2s is the least abundant and unit 5 is in this respect intermediate. Fig. 7b shows that unit 2n is on average at the lowest altitudes while unit 2s, on the highest and unit 5, at the intermediate, but close to unit 2n. Fig. 7c and d show that unit 2n has on average the higher microwave emissivity and the lower Fresnel reflectivity comparing to units 2s and 5. We interpret the observed microwave emissivities and Fresnel reflectivities of these units (Fig. 7c and d) as indication that the surface material within unit 2n and especially subunit 2nn is on average more consolidated comparing to units 2s and 5. We conclude that the surface material of the unit 2nn is less weathered and this is why it has the higher 1 μm emissivity. As it was shown in Section 2.4 the weathered basalts typically have the lower 1 μm emissivity. This is probably due to the presence of highly reflective anhydrite, which according to thermodynamic modeling (e.g., Zolotov, 2007) is a typical product of weathering of basalts on Venus. The lowest 1-μm emissivity typical for unit 2s may also be because of smaller surface grain size due to presence of eolian dust. The relatively high altitude typical for unit 2s is probably not the reason for its lowest 1-μm emissivity because the altimetrically higher main body (unit 3) and summit (unit 4) of Tuulikki Mons (see below) show an increase rather than a decrease of emissivity.

5.2.2. Chimon-mana Tessera (unit 1) vs. adjacent plains (unit 2)

Here we compare the 1-μm emissivity of Chimon-mana Tessera with the adjacent plains to the north (unit 2n) and to the south (unit 2s). It is seen in Table 5 that the calculated emissivity of this tessera is lower than that of the unit 2n and especially lower comparing to unit 2s and higher than that of the unit 2s. In these
cases the difference is statistically significant for both models lr8-e08 and lr8-058. As it follows from the consideration given in Section 5.2.1, the northern plains (especially the unit 2nn) seem to be more pristine and less weathered than the southern plains. So we may conclude from Table 5 that the surface material of Chimon-mana Tessera has the lower (by 15–35%) 1-μm emissivity than the basaltic material. This agrees with the results published by Herbert et al. (2008), Mueller et al. (2008), Hashimoto et al. (2008), and Gilmore et al. (2011a), where lower (comparing to supposedly basaltic plains) emissivity for other tessera massifs have been reported.

If the lower (comparing to tessera) emissivity of the southern plains would be due to weathering of their material, one could expect that tessera surface material, which was exposed to the atmosphere for the longer (comparing to the plains) time should also be weathered. However, thermodynamic calculations done by Barsu-
kov et al. (1980, 1982) show that felsic materials should be stable in Venus surface environment and the weathering-involved changes of their mineralogy and thus emissivity are not expected.

The tessera surface is on average higher by ≈0.6 km than the surface of adjacent plains (see Table 3). On Venus at higher altitudes winds should be stronger than at the lower ones (Kerzhanovich et al., 1983) and this may control the surface grains size: the higher the surface, the stronger the wind, and probably the coarser the surface material grain size. This change, however, should favor the higher (comparing to the plains) emissivity of the tessera surface material and the fact that when comparing tessera with the northern plains we do not see the increase, but see a decrease, is an indication that the altitude effect even if it exists is weaker than the effect of mineralogic composition.

5.2.3. Tuulikki Mons volcano main body (unit 3) vs. surrounding plains (unit 5)

It is seen in Table 5 that calculated emissivity of the material of the main body of Tuulikki Mons volcano (unit 3) is higher than that of surrounding plains (unit 5) and this difference is statistically significant for both models lr8-e08 and lr8-e058. As it was mentioned above, the Tuulikki morphology, the radial assemblage of rather long lobate lava flows on very gentle slopes, suggests a basaltic composition. The most part of the volcano is only slightly higher than the adjacent plains, but lobate flows, composing it, are geologically younger than the surrounding plains. So it is natural to expect that the Tuulikki material is less weathered than that of the surrounding plains and this probably explains its higher 1-μm emissivity. The unit 3 emissivity is virtually the same as that of unit 2n and this supports our suggestion that the units 2n material is not significantly weathered.

5.2.4. Tuulikki Mons summit (unit 4) vs. the main body of volcano (unit 3)

It is seen in Table 5 that calculated emissivity of the material of the Tuulikki Mons summit (unit 4) is lower than that of the volcano main body (unit 3) and this difference is statistically significant for both the model lr8-e058 and the model lr8-e08. This suggests that the difference is probably real. The lower emissivity of the summit material can be explained neither by the differences in the degree of weathering (on the volcano summit and slopes it should be approximately the same) nor by the coarser grain size of the summit surface material due to its higher altitude/higher wind velocities (it should work in the opposite direction). The reason may be different (more felsic) composition of the summit part of the volcano. This suggestion is supported by the presence of a steep-sided dome on the volcano top (Fig. 8). As it was mentioned above in Section 2.1, steep-sided domes were considered to be formed by eruptions of lavas geochemically more evolved comparing to basalts (Pavri et al., 1992) although other suggestions on their compositions have been also published (Fink and Griffiths, 1998; Bridges, 1995; Pavri et al., 1992).

5.2.5. The 1-μm emissivity vs. altitude correlation diagrams

As additional evidence of natural clustering of the identified units in the altitude – 1-μm emissivity space we have made for them correlation diagrams of these characteristics (Figs. 9 and 10). The figures show that in the values of 1-μm emissivity and...
altitude, plains show a significant scatter that obviously reflects their geologic variabilities as it was mentioned in Section 5.2.1. However, despite the mentioned variabilities, the considered units, are rather well clustered suggesting their geologic individualities.

5.2.6. Summary of the analysis of the 1-µm VMC data

Summarizing the results of our analysis of the 1-µm channel VMC images for the area SW of Beta Regio we can say that the plains here are rather variegated in their 1-µm emissivities. This seems to be mostly due to the degree of the weathering of their surface materials, that, in turn, probably depends on the geologic age with the younger materials being less weathered. The calculated emissivity of the surface material of Chimon-mana Tessera is about 15–35% lower than that of the less weathered plains. So the tessera material here has lower emissivity than the material of supposedly basaltic plains. The lower 1-µm emissivity of tessera material may be indicative of its non-basaltic, probably felsic composition as it was suggested, first by Nikolaeva et al. (1992) and then by Helbert et al. (2008), Mueller et al. (2008), Hashimoto et al. (2008), and Gilmore et al. (2011a). The calculated emissivity of the main body of Tuulikki volcano is very close to that of the plains which are considered to be the least weathered and higher than that of the surrounding plains of supposedly intermediate degree of weathering. The emissivity of the Tuulikki summit is somewhat lower than that of the volcano main body that can be due to different (more felsic?) composition of the surface material of the summit.

6. Discussion

During the 1990 Galileo Venus flyby, the Near Infrared Mapping Spectrometer (NIMS) investigated the night-side atmosphere of Venus in the spectral range 0.7–5.2 µm. The acquired data were analyzed by Hashimoto et al. (2008) to study Venus surface emissivity at 1.18 µm wavelength in the part of Venus disk from 20°W to 90°E. To reduce the random noise the data were averaged within a circle with radius 250 km. The temperature lapse used in the analysis is given by the Venus International Reference Atmosphere (Seiff et al., 1985), that is close to –8 K km⁻¹.

Hashimoto et al. (2008) mostly do not discuss a regional difference, but analyze surface emissivity as a function of surface altitude. They found that the majority of observed lowlands (<0 km altitude) has higher emissivity compared to the majority of highlands (>2 km altitude). Their interpretation is that the highland materials are generally composed of felsic rocks (granites?), while the lowlands are basaltic. Some regional differences are nevertheless mentioned: Ishtar Terra, Eistla Regio, and Alpha Regio have relatively low emissivity, while Bell Regio and a band from Tahmina Planitia and Fonoucha Planitia have relatively high emissivity values.

Most highlands on Venus are tesserae (Ivanov and Head, 1996; Tanaka et al., 1997; Ivanov, 2008) so the lower emissivity of highlands found by Hashimoto et al. (2008) seems to be attributed to tessera terrain. However if we consider the mentioned above three particular regions of relatively low emissivity the association of low emissivity with tessera is not so straightforward: Two of these regions, Ishtar Terra and Alpha Regio, are dominated with tessera, but Eistla Regio is the area of extensive, essentially young volcanism with morphologies suggesting basalts.

VIRTIS is an IR mapping spectrometer on-board Venus Express which acquired a large amount of data including the 1-µm night images of Venus surface (Drossart et al., 2007). These data cover mostly the southern hemisphere of the planet and thus are complementary to those acquired by VMC. So it is very appropriate to compare our results with theirs. Also their processing and analysis are rather similar to ours and the comparisons are appropriate from the methodological point of view too.

The issue of emissivity differences of various landforms and terrains is described in Helbert et al. (2008), Mueller et al. (2008), and Gilmore et al. (2011a). Their major finding is that the studied by them several massifs of tessera terrain show 1-µm emissivity lower than that of the surrounding supposedly basaltic plains. This implies that tessera material may be felsic although other options are also considered: different (from the plains) weathering regime and different surface grain size. Gilmore et al. (2011a) suggest one more option: tessera material could be non-igneous, which would affect the emissivity through the difference in composition and/or in grain size. Some, but not all, volcanic edifices, according to these works, show emissivity higher than that of surrounding plains. In Lada Terra, high emissivities were measured for young volcanic flows extending from the rim of Boala Corona, nested inside Quetzalpetlatl crater. Mueller et al. (2008) explains their emissivity increase by possible ultramafic composition of the lavas.

Our results lead to generally the same conclusions: our calculated 1-µm emissivity of tessera surface material is lower than that of relatively fresh lavas of plains and volcanic edifices. This suggests that the tessera material is probably not basaltic and may be felsic.

We have found that the surface materials of plains are very variegated in their 1-µm emissivity that probably reflects variability of their local geologic histories, mostly the degree of chemical weathering with less weathered materials showing the higher emissivities.

We have also found a possible decrease of the calculated emissivity at the top of Tuulikki Mons volcano which may be due to different (more felsic?) composition of volcanic products on the volcano summit comparing to its slopes. This suggestion seems to be supported by the observation that at the volcano summit there is a steep-sided dome.

We did not find any indication of the increase of surface material emissivity at higher altitudes which could result from the expectedly higher wind velocities (and thus the coarser grain size of the surface materials) on the higher altitudes. This suggests that within the considered altitude range, which is only 1–1.5 km, this effect, if exists, is not noticeable.

7. Conclusions

(1) The night-time VMC images provide reliable information on spatial variations of the NIR thermal emission of the Venus surface, which potentially may be interpreted in terms of geological characteristics of the studied area, including possible compositional differences between the geologic units.

(2) Our calculations for the area SW of Beta Regio showed that 1-µm emissivity of tessera surface material is lower than that of relatively fresh supposedly basaltic lavas of plains and volcanic edifices. This is consistent with the hypothesis that the tessera material is not basaltic and may be felsic. These results are in agreement with the results of Helbert et al. (2008), Mueller et al. (2008), Hashimoto et al. (2008), and Gilmore et al. (2011a) and with early suggestions of Nikolaeva et al. (1992). If the felsic nature of venusian tesserae is confirmed in further studies, this may have important implications for geochemical environments in early history of Venus, indirectly supporting a hypothesis of water-rich early Venus (e.g. Kasting et al., 1984; Kasting, 1988; Grinspoon and Bullock, 2003).
(3) We have found that the surface materials of plains in the study area are very variegated in their 1-μm emissivity, which probably reflects variability of their local geologic histories, mostly the degree of chemical weathering with less weathered materials showing higher emissivities. Future studies in the areas of geologically more homogeneous plains would be helpful in proving this suggestion.

(4) We have also found a possible decrease of the calculated emissivity at the top of Tuulikki Mons volcano which, if real, may be due to different (more felsic?) composition of volcanic products on the volcano summit comparing to its slopes. This suggestion seems to be supported by the observation that at the volcano summit there is a steep-sided dome. More evolved lavas in the latest stages of evolution of basaltic magma chambers are rather typical for magmatism of Earth (e.g. McBurney, 2006).

In our future studies we plan to involve new data acquired in the continuing Venus Express mission and to apply techniques of quantitative analysis to the existing and future data.

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