Aeolian sand transport and aeolian deposits on Venus: A review

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ABSTRACT

We review the current state of knowledge about aeolian sand transport and aeolian bedforms on planet Venus. This knowledge is limited by lack of observational data. Among the four planetary bodies of the Solar System with sufficient atmospheres in contact with solid surfaces, Venus has the densest atmosphere; the conditions there are transitional between those for terrestrial subaerial and subaqueous transport. The dense atmosphere causes low saltation threshold and short characteristic saltation length, and short scale length of the incipient dunes. A few lines of evidence indicate that the typical wind speeds exceed the saltation threshold; therefore, sand transport would be pervasive, if sand capable of saltation is available. Sand production on Venus is probably much slower than on the Earth; the major terrestrial sand sinks are also absent, however, lithification of sand through sintering is expected to be effective under Venus’ conditions. Active transport is not detectable with the data available. Aeolian bedforms (transverse dunes) resolved in the currently available radar images occupy a tiny area on the planet; however, indirect observations suggest that small-scale unresolved aeolian bedforms are ubiquitous. Aeolian transport is probably limited by sand lithification causing shortage of saltation-capable material. Large impact events likely cause regional short-term spikes in aeolian transport by supplying a large amount of sand-size particles, as well as disintegration and activation of older indurated sand deposits. The data available are insufficient to understand whether the global aeolian sand transport occurs or not. More robust knowledge about aeolian transport on Venus is essential for future scientific exploration of the planet, in particular, for implementation and interpretation of geochemical studies of surface materials. High-resolution orbital radar imaging with local to regional coverage and desirable interferometric capabilities is the most effective way to obtain essential new knowledge about aeolian transport on Venus.

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1. Introduction

1.1. Aeolian processes in the Solar System and the case of Venus

In the Solar System, eight (8) planetary bodies possess significant atmospheres. Four of them do not have solid surfaces. The remaining four bodies (the Earth, Venus, Mars, and Titan, a satellite of Saturn) have observable solid surfaces in contact with significant atmospheres; on all four, aeolian transport occurs and/or occurred in the recent geological past, and a number of associated morphologies are observed at their surfaces; we recommend a nice general overview by Lorenz and Zimbelman (2014). There are systematic differences in the observed morphologies between the four bodies; these differences, however, seem minor given the huge variability of aeolian landforms on each planet (at least, on the Earth and Mars, where extensive data are available), and a factor of \(10^5\) difference in the air density (from martian mountaintops to venusian canyons). Extremely faint (two more orders of magnitude less dense than on martian mountaintops) atmosphere of Triton (a satellite of Neptune), although capable of sustaining dust plumes, is not known to create aeolian morphologies; nothing conspicuous is seen in Voyager 2 imagery at \(\sim\)1 km resolution. Recently New Horizon mission to Pluto, which possesses a comparable faint atmosphere, imaged some features resembling aeolian bedforms, however, aeolian origin of these features is not certain.

Venus gives us the case of the densest atmosphere in the Solar System. This case got an additional attention with the beginning of the Global Surveyor mission also collected radiometric data and radar altimeter data. "Resolution" is used in its original meaning, the minimal distance between two small objects that can be resolved in the image, and it is highly important to know, whether the global mixing of material occurs or not. In the absence of fluvial and glacial transport, wind is the only agent potentially capable of moving material for long distances. In this review we focus on sand transport via saltation.

For terrestrial arid land, aeolian processes move sand regionally; rivers, oceans and heavily vegetated areas prevent global sand transport. Saltation on Mars is common (e.g., Bridges et al., 2012); possibly, it was even more common in the recent geological past under different climate conditions. In places, we see clear evidence for migration of sands over hundreds of kilometers on Mars, however, no firm positive evidence for global transport has been reported so far. On Titan, the global equatorial belt of dunes (e.g., Rodriguez et al., 2014) hints at global transport by saltation. For Venus, the question about global aeolian transport of sand remains unanswered; in this review we focus on available relevant observations.

1.2. Data sources

The surface of Venus is hidden from direct view by a thick layer of clouds and possibilities to study the surface by remote sensing means are very limited. The atmosphere is transparent for electromagnetic waves with wavelengths \(\lambda\) longer than \(\sim 6\) cm. Remote sensing observations in the microwave wavelengths above this limit is the main information source about Venus geology. Active probing by radar techniques gives inherently higher resolution than passive detection of thermal radiation (radiometry). Radar images at \(\lambda = 12\) cm obtained during NASA Magellan orbital mission (1990–1994) to Venus (Saunders et al., 1992) cover the planet almost globally at \(\sim 200–300\) m resolution. (Here the term “resolution” is used in its original meaning, the minimal distance between two small objects that can be resolved in the image.) We refer these data simply as “images”. The data are available from the NASA Planetary Data System at http://pds.nasa.gov. A moderately convenient interface to global mosaics can be found at http://mapaplanet.org. In addition to these images, Magellan mission also collected radiometric data and radar altimeter data.
Another set of microwave radar data has been acquired by earth-based facilities at the same wavelength (e.g., Campbell et al., 1989; Carter et al., 2004). The obtained radar images have lower resolution and limited coverage in comparison to Magellan, but give important polarimetric information. Unfortunately, at the time of this writing these data are not available for independent analysis. Soviet orbital missions Venera-15, -16 (1983–1984) (Kotelnikov et al., 1985) covered the northern quarter of the Venus surface with radar images at λ = 8 cm. A set of microwave observations at λ = 17 cm was obtained by NASA Pioneer Venus Orbiter mission (1978–1990) (Pettengill et al., 1979). In some sense, these data are complementary to Magellan data set, however, because of their lower resolution and some other peculiarities, they are of little use for analysis of aeolian features.

The Venus’ atmosphere does not absorb electromagnetic radiation in a few narrow near-infrared “transparency windows”, in which the atmosphere is translucent, but not transparent because of intensive scattering both in the clouds and in the densest lower atmospheric layers. Thermal emission of the surface in these windows is detectable on the night side of the planet, but the resolution of obtained images is inherently low (~100 km) due to scattering in the clouds (e.g., Mueller et al., 2008).

Irrevaluable information has been obtained in-situ by Soviet landers Venera – 8–14 (1972–1982) and Vega -1, – 2 (1985), including visual panoramas of 4 landing sites and direct measurements of near-surface winds. These observations will be discussed in detail below.

1.3. Brief primer on Venus geology

A number of papers in Venus II collective monograph (Basilevsky et al., 1997; Tanaka et al., 1997; Greeley et al., 1997; Campbell et al., 1997; McKinnon et al., 1997, etc.) give a detailed review of our understanding of Venus geology and geomorphology revealed by Magellan radar observations. This subsection presents only the most basic essential facts from those publications.

Venus, a “sister planet”, has almost the same size, mean density and gravity as the Earth. Similarly to the Earth, it has a silicate crust and silicate mantle, however, there is no plate tectonics on Venus. The surface as it is seen in the radar images is dominated by vast volcanic plains, presumably, basaltic, similar to lunar maria, volcanic plains on Mars and Mercury, and possibly the large igneous provinces on the Earth. Venusian plains are disrupted by tectonic fabric, including several heavily tectonized rift zones. Variety of volcanic edifices superposed over and embayed by the regional volcanic plains span three orders of magnitude in size. Elevated topographically rough heavily tectonized radar-bright massifs of debated origin, so-called tesserae, comprise ~10% of the surface area.

About a thousand impact craters on Venus range from a few km to ~270 km in diameter; they are spread rather uniformly over the planet. The mean age of the surface inferred from these craters is on the order of 500 Ma and is rather uncertain. Unlike the Moon, Mars, and many other planetary bodies, Venus lacks geomorphological features (impact basins, heavily cratered terrains) formed early (>3500 Ma ago); in this sense the surface of Venus is young. The majority of morphological features we observe in the radar images formed tens to hundreds Ma ago (e.g., Kreslavsky et al., 2015), much earlier than all geomorphological features of comparable size that we observe on the Earth; in this sense the venusian surface is very old.

At the available image resolution, we see little morphological expression of erosion of any kind. Combined with the ages of morphological features, this indicates low erosion rates.

1.4. Review outline

We start (Section 2) with conditions at the Venus surface and review current knowledge of peculiarities of aeolian processes under such conditions. Then we consider the scarce knowledge of the major components of aeolian transport: near-surface winds (in Section 3) and availability of movable material (in Section 4). Then (Section 5) we present comprehensive overview of all traces of aeolian transport that we observe at the surface: both directly observed morphological features and indirect inference from remote sensing data. We present some attempt on synthesis (Section 6) and summarize the information pertinent to the major question about the presence or absence of global sand transport and long-range material mixing on Venus. In the end (Section 7) we briefly present our subjective view of new observations that would be critical for advance in understanding aeolian sand transport on Venus.

2. Aeolian transport under venusian conditions

2.1. Conditions

The atmosphere of Venus consists mostly of carbon dioxide CO₂ (>96%) and nitrogen N₂ (3.5%); a number of other gaseous species (Ar, Ne, SO₂, H₂O, CO, H₂S, HCl, OCS, etc.) are present in small amounts (Esposito et al., 1997). The lowermost atmosphere is hot: the temperature at typical surface elevations is 720–750 K, the nominal temperature lapse rate is ~8 K km⁻¹, close to adiabatic (Seiff, 1983). The day/night and weather-related temperature variations in the near-surface atmosphere are minor (see discussion in Section 3.2) despite a very long solar day (~3.8 months).

The atmosphere is massive: the pressure at the typical surface elevations (6051–6052 km planetary radius) is 90–95 bar (9.0–9.5 MPa) (Seiff, 1983). It varies from ~120 bar in the deepest canyons to ~45 bar at Maxwell Montes summit. This pressure is higher than the critical pressure of N₂ and, except mountain tops, CO₂. Therefore, the near-surface atmosphere is actually not a gas, but a supercritical fluid.

The density of the atmosphere at the typical surface elevations is ~60–70 kg m⁻³ (and down to 33 kg m⁻³ at Maxwell Montes summit), a factor of ~50 higher than terrestrial air density and a factor of ~15 lower than water density. In a sense, Venus’ air is more similar to water than to terrestrial air. Given that venusian gravity (8.9 m s⁻²) is very similar to terrestrial, it is reasonable to expect that peculiarities of aeolian transport on Venus would be transitional between aeolian transport on the Earth and terrestrial subaqueous sediment transport.

2.2. Saltation under venusian conditions

From theoretical models of saltation physics Kok et al. (2012) found that typical saltation length under venusian condition is ~1 cm, a factor of 30 shorter than for terrestrial conditions, and typical saltation height is ~0.2 cm, a factor of 15 lower. Similar values were observed in dedicated wind tunnel experiments under venusian conditions (Greeley et al., 1984a). For aeolian saltation in terrestrial and martian conditions, a “splash” process, when landing saltating grain knocks out other grains, plays a significant role causing wind threshold hysteresis: wind speed needed to support saltation is slower than needed to initiate it. On Venus, similarly to terrestrial subaqueous saltation, the “splash” process plays no role, and no wind speed hysteresis is expected (Kok et al., 2012, and references therein).
Sand transport on Venus is sluggish in comparison to the Earth. The volumetric sand flux, according to Kok et al. (2012), can be calculated as

\[ Q_V = C \left( \frac{\rho_s}{\rho_f} \right) u_{th} (u/\tau_{sth})/g, \]

where \( \rho_s \) is the sand material density, \( \rho_f \) is the fluid (air or water) density, \( u \) is the shear velocity (also called friction speed) of the driving wind, \( u_{th} \) is the saltation threshold shear velocity, \( g \) is acceleration due to gravity, and \( C \approx 5 \) is a dimensionless empirical constant. To calculate a single flux value that would characterize sand transport on Venus we assume the wind shear velocity “slightly above” the threshold and use the following dimensional scaling:

\[ Q_V \approx 2(\rho_s/\rho_f)u_{th}^2/g, \]

which, for the venusian \( u_{th} \sim 0.02 \text{ m/s} \) (Iversen and White, 1982; Greeley and Iversen, 1985) gives \( Q_V \sim 4 \times 10^{-6} \text{ m}^2/\text{s} \) or \( \sim 1 \text{ m}^2 \) per year, a factor of ten lower than for the Earth, if calculated in the same manner.

Wind tunnel experiments (Greeley et al., 1984a) showed that the particle size with the lowest saltation threshold is \( \sim 75 \mu \text{m} \), similar to the terrestrial case. Due to much denser air the salutation threshold is much lower than on the Earth, \( \sim 0.5 \text{ m/s} \) free flow velocity (not to be confused with \( u_{th} \) for these fine sand particles, and slightly higher for coarser sand. Similarly to terrestrial subaqueous transport, larger (sub-millimeter) sand particles move by rolling rather than by saltation.

2.3. Bedforms

On the basis of a simple physical theory of saltation-driven growing instabilities of sand sheets (Kroy et al., 2002), Claudin and Andreotti (2006) proposed a simple scaling estimate of primary dune wavelength:

\[ L = \frac{53d\rho_s}{\rho_f}, \]

where \( d \) is particle size, \( \rho_s \) is again the sand material density, and \( \rho_f \) is the fluid (air or water) density. This gives 10–20 cm primary fine sand dune wavelength for Venus, compared to \( \sim 20 \text{ m} \) for the Earth. Formation of transverse dunes within10 cm wavelength under Venus’ conditions has been observed in the wind tunnel experiments by Greeley et al. (1984b). On the Earth well-developed dunes are known to grow three orders of magnitude larger than the characteristic incipient dune size, up to several kilometers. Analogous growth is reasonable to expect on Venus, winds and sand availability permitting.

Terrestrial subaerial dunes and other aeolian bedforms demonstrate a rich variety of morphologies controlled by several factors, namely, sand availability, degree of its sorting (the presence of particles of different sizes and density), wind speed, variability of wind direction, when wind speed exceeds the salutation threshold, the presence and nature of vegetation. On Venus these factors, except the latter, are also likely to affect aeolian bedform morphology and cause morphological variability.

Terrestrial subaqueous sand deposits also demonstrate variability of morphologies. Morphologies formed under stable current conditions have been reviewed by Mazumder (2003). Several specific morphologies, for example, antithunes, are forming in shallow water channels; the free surface of water plays a critical role in physics their formation; such bedforms are not expected to occur in regular subaerial environment on Venus. In deep water, turbidity currents loaded with suspended fine sediment often produce morphologies typical for shallow water due to the presence of the dividing surface between normal and denser water. Density flows (analogous to turbidity currents) on Venus and related morphologies might form in aftermath of some rare unique events like volcanic eruptions, meteoritic impacts, large landslides; pervasive turbidity flows seem very unlikely given the apparent absence of erosion.

A systematic overview of subaqueous sand bedforms formed by deep water currents has been done by Stow et al. (2009). Fig. 1 compiled from their work presents different types of bedforms as function of the particle size and flow velocity. The following morphologies have been described. Linear streaks aligned with the flow have spacing up to decimeters; crag and tail refers to the elongate mound deposited immediately downstream of an obstacle (crag) in the path of flow (tail), centimeters to decimeters long. Scour refers to elongated marks around and extending downstream from obstacles. Scour length varies from meters to
hundreds of meters and may be observed without associated obstacles. Ribbon marks are elongate moulded filaments of sand, mostly regularly spaced, with parallel to slightly sinuous planform. Most of those reported are large-scale bedforms (width 10–100 m, length 5–50 km), although smaller-scale features (width of a few meters) have been referred to as narrow ribbons, sand streamers and sand streaks. Furrows are elongate, primarily erosional features, with regular to irregular spacing and are relatively large scale features (width 5–150 m, length 1–10 km). Ripples are the smallest transverse bedforms (wavelength 0.1–0.6 m, height 0.02–0.1 m). Sand waves have a longer wavelength (5–500 m), flatter bedform (height 0.5–5 m) compared to dunes.

Extensive studies of morphologies produced by sand and wind under Venus' conditions have been carried out by Greeley et al. (1984b), Greeley and Arvidson (1990), Marshall and Greeley (1992), Greeley et al. (1997) using a wind tunnel. Their results are summarized in Fig. 2 as a diagram in the same particle size – wind velocity space as in Fig. 1. Three flow regimes were identified based on threshold wind speed for rolling/intermittent saltation, continuous saltation, and suspension. These three regimes resulted in longitudinal bedforms, transverse bedforms, and featureless beds, respectively.

The diagrams in Figs. 2 and 1 have some similarities but are not identical. The difference is not only caused by the difference between fluid density and viscosity: the wind tunnel experiments had limited run time and test section size. For example, the larger size of subaqueous transverse dunes reported by Stow et al. (2009) in comparison to the microdunes in the wind tunnel is probably caused by the fact that the latter are incipient, while the former are well developed. It is interesting that transverse dunes, typical subaerial bedforms formed under stable wind direction in the presence of abundant sand, occupy a very limited domain in the particle size – wind velocity space both in Figs. 1 and 2.

With time, dunes can grow and coalesce to form a hierarchy of progressively larger dunes. On the basis of theoretical considerations and terrestrial observations, Andreotti et al. (2009) argued that this growth is limited by the dune wavelength approximately equal to the thickness of the convective boundary layer in the atmosphere. Lorenz et al. (2010) showed that spacing of dunes on Titan is likely equal to the boundary layer thickness. On Venus, the dynamics of the lowermost atmosphere is poorly understood. If the atmosphere were not convecting at all, the diurnal temperature variations at the equator would reach ~7 K (Gierasch et al., 1997). On the other hand, Magellan microwave radiometry experiment (Pettengill et al., 1992) did not detect diurnal variations above its 2–5 K detection threshold. This suggests that the convective boundary layer does exist on Venus. The atmosphere above 12 km altitude is reliably known to have a high static stability, which suggests that the boundary layer is thinner than 12 km. The only one sufficiently accurate pressure and temperature profile through the lowermost 12 km of the atmosphere has been recorded during Vega 2 lander descent. The static stability profile derived from these data (e.g., Zasova et al., 2006) suggests ~4 km thick convective boundary layer, however, unexpectedly high negative static stability at 3–4 km make these data suspicious. Andreotti et al. (2009) and Lorenz et al. (2010) used the following estimate for the boundary layer thickness:

\[ \lambda = \Delta T / \Gamma \]

where \( \Delta T \) is the seasonal temperature amplitude, and \( \Gamma \) is the potential temperature gradient. Extrapolation of this scaling relationship from quickly spinning Earth and Titan to slowly spinning Venus obviously requires replacement of seasonal temperature amplitude with diurnal. A “standard” value of \( \Gamma \) for the lowermost atmosphere is ~0.5 K/km (Seiff et al., 1985), although this value is a guess and is not based on actual measurements; assumption of \( \Delta T \approx 2 \, K \) at the detection threshold gives \( \lambda \approx 4 \, km \), consistent with the Vega 2 profile. It looks like a few km is a likely guess for the boundary layer thickness and therefore for spacing of the largest possible aeolian bedforms on Venus.

3. Wind regime at the venusian surface

3.1. Direct wind measurements at the surface

Remarkable direct measurements of wind speed on Venus have been carried out on Venera-9 lander with specially designed anemometers (Avduevskii et al., 1976, 1977) (Fig. 3). The anemometers themselves were calibrated before flight in a small pressurized wind tunnel under venusian air density; the density
effect on calibration turned out to be small. Two anemometers were mounted on the lander at ~1 m above the surface at 135° angle from each other with respect of vertical axis of the lander. Performance of the anemometers mounted on a spare copy of the lander was assessed under terrestrial conditions in the field. It was found that the measured wind speed is very close to the true wind speed at the same height above the ground, unless the anemometer is in the lee of the upper part of the lander; in the latter case the other anemometer is always outside the lee and gives correct measurements.

Measurements on Venus were taken at ~31°N latitude on October 22, 1975, about 1 pm local solar time. Eight (8) measurements were taken by one anemometer and 9 by the other. The first measurement was taken ~6 min and the last is ~49 min after the landing. Each measurement was an average of 6 readings taken within 16-s long intervals. The measurements are scattered around ~0.5 m s⁻¹ within 0.3–0.7 m s⁻¹ range; there were no systematic differences between two anemometers and no systematic trend within 43 min of measurements. The latter indicates that the measured wind is a regular wind at the landing site in early afternoon and not the result of atmospheric disturbance caused by the lander descent. The anemometers were not designed to measure wind direction. It seems that none of 17 measurements was taken in the lee of the upper part of the lander (otherwise, we would expect much lower speed in some measurements). If wind direction changed completely randomly, the probability that at least one measurement is taken in the lee is ~80%; thus, there is some indication for the stability of wind direction.

The same anemometers were mounted on Venera-10 lander, which landed at ~16°N on October 25, 1975 about 2 pm local solar time. Total 34 readings were obtained from one of the anemometers during first 1.5 min after landing (Avduevskii et al., 1976, 1977). The readings were scattered around ~1.0 m s⁻¹, no systematic trend within 1.5 min was observed.

Ksanfomaliti et al. (1983) reported estimations of wind speed at Venera-13 and -14 landing sites from the level of acoustic noise.
registered by microphones located outside the landers at ~0.3 m above the ground. The microphone on Venera-14 certainly worked well, because during initial 3 min after landing it registered high-level sounds produced by lander operations at known moments. During the first 4 min between noisy operations and then within 10 short measurement windows until 54 min after landing the microphone registered rather stable slightly varying signal interpreted as noise produced by stable wind in the microphone and its supporting frame. Ksanfomaliti et al. (1983) tried to calibrate the noise level in terms of wind speed by measuring wind-induced acoustic noise in the same microphone and frame in wind tunnel in laboratory environment, rescaling the results to venusian air density, and involving some additional considerations. They reported ~0.3 m s\(^{-1}\) wind speed and are cautious about exact absolute value. We are even more skeptical about their absolute calibration, however, the detection of stable wind without systematic trend during 54 min of observations looks reliable. This observation was made at ~13°S latitude on March 5, 1982 about 10 am local solar time.

The signal from the same microphone on Venera-13 was saturated because of unlucky amplifier settings different from Venera-14. Ksanfomaliti et al. (1983) applied their calibration and reported ~0.5 m s\(^{-1}\) lower limit for wind speed.

Venera-13, however, gave another spectacular evidence for winds at the surface. It landed at ~7°S on March 1, 1982 about 10 am local time. At the touchdown some local loose material was emplaced on the base ring of the lander (Fig. 4). During ~85 min after landing, a scanning camera took several repeating panoramas of the landing site, including the ring, which was imaged 5 times. Some loose material was progressively removed from the ring, apparently, by wind (Selivanov et al., 1982). Individual particles are not resolved on the panoramas, and particle size of the loose material is unknown. However, this observation clearly indicates the stable (not immediately related to the lander descent) wind capable of moving loose material.

Principally different direct measurements of winds are obtained from tracking of probes and landers during their descent through the atmosphere. For the lowermost, near-surface altitudes such measurements are not very informative: the measured velocities are within measurement precision (1–3 m s\(^{-1}\)) from zero; they are summarized, e.g., by Kerzhanovich and Marov (1983), and Schubert et al. (1980). All mentioned observations were recently reanalyzed and statistically modeled by Lorenz (2016). He concluded that winds capable of moving sand and dust are rather common.

3.2. Dynamics of the lower atmosphere

Motion of venusian atmosphere at the level of the clouds (45–65 km above the surface) is dominated by quick super-rotation: the atmosphere rotates from east to west much quicker than the planet. This motion is well studied with numerous remote sensing observations. Observational data about the lowermost parts of the venusian atmosphere are insufficient for detailed understanding of their dynamics. Some understanding can be gained from first principles.

The major part of solar irradiation is absorbed in the clouds, however, a small but non-negligible portion penetrates down to the surface and absorbed there. This cause formation of diurnal convective boundary layer, however, winds induced by this convection are difficult to estimate (see Gierasch et al., 1997 for detailed discussion). Dobrovolskis (1993) showed that irradiation of inclined surfaces should cause diurnal cycle of upslope and downslope winds with poorly constrained speed on the order of a few m s\(^{-1}\).

Due to very low obliquity of spin axis with respect to the orbital plane, there are no seasons on Venus. The surface irradiation is greater at the equator than at high latitudes, while the surface temperature (lapse compensated) is uniform within a few K precision achieved by microwave radiometric measurements. The trend in irradiation and uniformity of temperature means the existence of advective heat transport. The first principles predict (e.g., Gierasch et al., 1970; Golitsyn, 1970) a Hadley-type circulation pattern of the lower atmosphere with equatorial upwelling, polar downwelling, equatorward near-surface flow slightly deflected to the east by weak Coriolis force. (The Coriolis force is weak because Venus rotates slowly; it deflects the equatorward flow to the east because, unlike the Earth, Venus spins from east to west.) Quantitative analysis of such kind of circulations by Stone (1975) predicted very small latitudinal temperature contrasts (~0.1 K), meridional flow velocities on the order of ~2 m s\(^{-1}\), and zonal flow velocities on the order of ~1 m s\(^{-1}\). As we mentioned above, such velocities are consistent with constraints obtained by descending probes and landers. Flow velocities give reasonable proxy (but not exact values) for expected wind speed.

Analysis of temperature measurements in the lowermost layers of the atmosphere made by descending probes and landers (e.g., Schubert et al., 1980) indicates that the near-surface temperature field is not as uniform as predicted from the first principles: there are temperature variations on the order of ~5 K, lapse rate compensated. This indicates the presence of some air motion in addition to the stable Hadley cells described above.

3.3. Climate change

The present-day eccentricity of the Venus’ orbit (0.007) is anomalously low. It varies chaotically with a characteristic time scale of ~0.05 Ma. The median eccentricity over the last 10 Ma is 0.032, and the peak values exceed 0.07; the last such eccentricity peak occurred ~1.4 Ma ago. Eccentric orbit leads to an uneven distribution of incoming solar energy flux through the year. At the highest eccentricity, the peak solar radiation flux is 15% higher than now, and 32% higher than in the minimum, while the increase of the total annual insolation is minor (0.2%). As far as we know, nobody has assessed the effect of Venus eccentricity on climate. In principle, since the radiative time scale of Venus’ atmosphere is much longer than the orbital period, we would not anticipate any change in the wind regime at a different eccentricity. However, intricate internal feedbacks in the climate system make accurate prediction about winds difficult. The latter is also true for climate change forced by variations of minor abundances of greenhouse gases (Bullock and Grinspoon, 1996, 2001). Such variations could be caused by pulses of volcanism and/or secular change in volcanic gas supply in response to secular change of geodynamic style.

3.4. Summary

Our knowledge of near-surface winds and their variability on Venus is very poor. However, both direct observations and first-principle inferences point to the presence of winds with typical speed about of 0.5–1 m s\(^{-1}\) at ~1 m above the surface. This wind speed is exceeding or similar to the saltation threshold for sand particles (Section 2.2). Thus, if sand capable of saltation is available, we would expect saltation to occur.

4. Availability of sand

4.1. Sources of sand

On the Earth, the most powerful processes that disintegrate volcanic rocks to produce sand-size particles are chemical weathering and mechanical erosion by flowing water, mechanical erosion by
glacial ice, and the freeze–thaw cycle. The very hot atmosphere and the lack of liquid water on Venus make these weathering and erosion processes impossible. Thus we would expect production of particulate material on Venus to be much slower in long-term average in comparison to the Earth. Nevertheless, there are some other processes that can produce particulate material; we consider them below.

4.1.1. Chemical weathering

Venus atmosphere has orders of magnitude greater content of highly corrosive and chemically reactive gases such as HF, HCl, SO₂; the main atmospheric component, supercritical CO₂, is also reactive (Fegley et al., 1997). This points to the possibility of direct chemical interaction of the atmosphere with the surface. Several lines of evidence indicate that such chemical weathering of the primary volcanic rocks in the absence of liquid water indeed occurs. According to thermodynamic calculations, concentration of chemically active atmospheric gases is sufficient to support chemical alteration processes of primary minerals of mafic and ultramafic igneous rocks such as carbonization, oxidation, formation of sulfates, sulfides, and F- and Cl-bearing minerals (Urey, 1952; Mueller and Kridelbaugh, 1973; Zolotov and Volkov, 1992; Kargel et al., 1994, etc.). Moreover, the observed abundance of HCl and HF is consistent with buffering by reactions with surface rocks (Fegley and Treiman, 1992; Treiman and Bullock, 2012).

Several observations on the basis of the surface remote sensing data can be explained by chemical weathering and thus serve as evidence for it. (1) Surface material at high elevations have very unusual electromagnetic properties, in particular, anomalously low microwave emissivity and high microwave radar cross-section (image brightness) (e.g., Pettengill et al., 1988, 1992; Klose et al., 1992). Several hypotheses explaining this phenomenon involve the presence of some specific products of chemical weathering of igneous rocks (e.g., Wood, 1997; Treiman et al., 2016). Such explanations, however, are not unique: there is another class of hypotheses based on condensation of some atmospheric species (e.g., Fegley et al., 1997). (2) Smrekar et al. (2010) reported association of stratigraphically young volcanic lava flows with areas of low near-IR emissivity. If those observations are technically correct, they can be naturally explained by near-IR emissivity increase with time due to chemical weathering. (3) Bondarenko et al. (2003) showed that stratigraphically older volcanic units have higher microwave emissivity, which again can be explained by progressive accumulation of high-emissivity material with time. (4) Similarly, Campbell et al. (1992) noted that the youngest impact craters usually have floors of anomalously low microwave emissivity, which again is consistent with emissivity increase with time. The latter two observations, however, can alternatively be explained by accumulation of wind-transported material. The microwave signature (3,4 above) indicate at least a few cm thick presumably altered layer, while the near-IR signature (2 above) can be produced by a several-microns-thick veneer.

Fig. 5. Magellan radar image of a diffuse radar-dark parabola associated with crater Stuart (30.8°S, 20.2°E, D = 67 km). On this and all other radar images brighter shades correspond to higher radar cross-section values, north is at the top.

Fig. 6. Magellan radar image of a diffuse radar-dark halo associated with crater Lind (50.2°N, 355.0°E, D = 25).

Fig. 7. Magellan radar image of Al-Uzza Undae. Faint NE-SW-trending lineaments are interpreted as transverse dunes; bright SE-NW-trending lineaments are wind streaks.
The surface of Venus seen by the panoramic cameras of Venera 10, 13, and 14 consists of slabs of material a few centimeters thick (Florensky et al., 1977, 1983; Basilevsky et al., 1985) (Fig. 4). The finely layered slab material is mechanically soft, as deduced from spacecraft landing dynamics (Basilevsky et al., 1985) and penetrometer data (Kemurdzhian et al., 1983). This soft material has been interpreted as chemically altered lavas (Garvin et al., 1984), although other interpretation (Florensky et al., 1983) involves lithified aeolian deposits.

Chemical reactions of possible mineral assemblages and atmospheric gases discussed above go with changing of material density. The weathering of primary basalts to the stable secondary mineral assemblages on Venus would be accompanied by volume increase (density decrease) (Klose et al., 1992). This would lead to disintegration of solid rocks into particulate materials. Although dominant particle sizes are difficult to predict, formation of sand-size particles is likely. Even if chemical weathering goes without a density or volume change, re-crystallization of new mineral assemblages can still cause disintegration.

The thickness of weathered layer depends on chemical reaction at the gas-mineral interface, and resurfacing at different scales (Zolotov and Volkov, 1992; Fegley et al., 1997) which would expose unaltered materials. Purely diffusion-limited weathering is extremely slow: accumulation of a thin veneer of altered material would quickly reduce the weathering rate.

4.1.2. Impacts

The dense atmosphere shields the surface from impacts of small meteoroids; impact craters smaller than ~1 km do not form, and the rate of formation of 1–25 km craters is reduced (e.g., McKinnon et al., 1997). Impacts by large meteoroids, however, do occur on Venus. Large impacts are known to produce abundant dis-
integrated material, although detailed quantitative knowledge of the yield is not accurate. The size-frequency distribution of large craters is such that a few larger craters contribute more to the production of fines than a larger number of smaller craters. Garvin (1990) estimated that the total ejected volume of venusian craters is equivalent to several meters thick equivalent global layer, however, a dominant part of this material form proximal ejecta well seen in the images. He estimated the total amount of fines capable of suspension (< /C24 30 l particles) of 1–2 mm global-equivalent layer. Amount of sand-size material is significantly smaller than the total ejecta volume, but probably larger than the amount of fines.

Radar data give reliable observational evidence for deposits of particulate material resulted from large impacts. About 67% of all craters with a diameter >5 km are associated with so-called dark diffuse features (DDFs) (e.g., Basilevsky et al., 2004). They are seen in Magellan radar images (Figs. 5 and 6) as large (hundreds to over a thousand of kilometers) radar-dark areas surrounding craters. Their boundaries are diffuse; numerous small features (tectonic lineaments, lava flow boundaries) can be traced from outside into the dark areas except their darkest innermost parts. Their global distribution is seen in the global radar image mosaic in Fig. 10b.

Some DDFs (~10% of all craters) have a very specific planform of a parabola surrounding its crater with its apex pointing to the east (Fig. 5), so that the parabola is open to the west. General mechanism of parabola formation is well understood (Vervack and Melosh, 1992; Campbell et al., 1992). The impact ejects particulate material into space; the particles first travel above the atmosphere along ballistic trajectories, then re-enter the atmosphere and descend with Stokes vertical velocities. While passing through the middle atmosphere (~30–70 km above the surface) the particles are transported westward by fast super-rotation atmospheric flow. Then they reach the surface and produce a deposit. Kinematical modeling (Vervack and Melosh, 1992; Schaller and Melosh, 1998) does reproduce the parabolic planform.

Schaller and Melosh (1998) used geometry of the parabolas themselves to independently estimate the size distribution of particulate material ejected by large impacts. They used known dependence of the descent velocity as a function of particle size, assumed the present-day superrotation wind speed, and used assumptions about ejecta velocity distribution based on observations on the Moon and numerical modeling. Their inference led to an estimate of ~8 cm equivalent global layer of particulate material created by all observed craters. Unfortunately, these calculations are not perfectly consistent: the inferred size distribution leads to parabolas dominated by particles a few cm in size, which contradicts to the observed radar signature of the parabolas; dark semitransparent appearance of the deposit in radar images indicates that the particles are smaller than a centimeter. To settle and not being suspended the particles should be larger than tens of microns. Thus, the parabolas probably are made of fine pebble-size particles with possible sand-size particle component.

Both estimates based on the cratering mechanics (Vervack and Melosh, 1992) and analysis of microwave probing results (Bondarenko and Head, 2004) consistently yield thickness on the order of meters in the thickest parts near the largest craters.
Crater-related deposits, including parabolas are usually well seen in microwave emissivity maps obtained by passive radiometry onboard Magellan. In the emissivity maps, such deposits are usu-
ally larger than DDFs in the radar images; this indicates the pres-
ence of thin (down to decimeter) deposits transparent for radar in
the peripheral parts of the DDFs.

Non-parabolic DDFs are usually interpreted to form by progres-
sive degradation of parabolas (Izenberg et al., 1994; Basilevsky and
Head, 2002, 2006) at tens of Ma time scale. Morphological signa-
ture of such degradation is controversial (Bondarenko and Head,
2009), however independent morphological observations confirm
that the craters with parabolas are among the youngest on the
planet. It is not excluded that some or all non-parabolic DDFs formed
during some epoch(s), when there was no atmospheric superrot-
tation. In summary, details of evolution of DDFs are uncertain, but
they are certainly meters-thick deposits of particulate material
with probable significant proportion of sand-size particles.

Ivanov et al. (1992) argued that the impacts themselves cause the
erosion due to accompanying atmospheric shock waves in the
dense atmosphere. If the surface is made of friable lightly indu-
rated soil, such material can be mobilized and eroded away up to
200 km from the crater center. This effect may reactivate old indu-
rated sand deposits.

4.1.3. Volcanism

On the Earth volcanism is a source of the fragmented material
that covers underlying surface in two ways: as pyroclastic flows
and ash fallout deposits. The mix of fragmented material, tephra,
composed by ash particles, their aggregates and rocks, with vol-
canic gases can form pyroclastic density flow originated in the vent
and extended downslope. We do not include such flows them-
selves into the aeolian process inventory. Pyroclastic flows can
form during explosive volcanic eruptions. Based on possible Venus’
conditions and reasonable magma properties, Head and Wilson
(1986) predicted relatively small amount of pyroclastic material on
Venus. On the other hand, recent study of conditions required
to establish and sustain buoyancy column during eruptions by
Glaze et al. (2011) expected that pyroclastic flows would be rela-
tively more prevalent on Venus in comparison to the Earth.

On the Earth, and perhaps on Venus either, the deposits left by such
flows are solidified and form rocks, often soft and friable; they
are source of sand-size particles produced by subsequent erosion,
but not particulate materials themselves.

Explosive volcanic eruptions generate large amounts (>50% of
total erupted mass) of fine ash particles which are dispersed into
the atmosphere by buoyant plumes above the volcanic vents and
pyroclastic density currents (e.g., Carey and Sigurdsson, 1982;
Hildreth and Drake, 1992; Durant and Rose, 2009). Grain size spec-
trum of the ash is wide and includes significant amount of fine
sand size grains. Settling ash particles often form aggregates. In
summary, volcanism does produce fine sand particles capable of
aeolian transport, but we do not know, whether this source is sig-
nificant or negligible in comparison to impacts.

4.1.4. Other sources

Lineaments of tectonic origin are abundant in radar images; they
are surface expressions of tectonic faults of different spatial
scales. Tectonic deformation acts as a cause for the crust ruptures
and consequent fragmentation of bedrock into debris. On the Earth,
the upper crust in tectonically active regions is fragmented into
blocks down to the scale of boulders and smaller (e.g., Molnar
et al., 2007). Tectonic faults themselves could be the source of fine
material that is formed under shear stresses. A small proportion of
sand-sizes particles formed in faults could be exposed on the sur-
face and involved into aeolian transport. On the Earth this source of
sand is very minor in comparison to the effect of flowing water; on
Venus its contribution might be more significant.

Disintegration of rocks forming sand-size particles also occurs
due to collision processes: grain-to-grain collisions lead to grain
fracturing by rapid stress changes associated with sudden unload-
ing. The collision processes include rock falls, large-scale land-
slides/rock avalanches, as well as wind abrasion.

On the Earth, earthquakes often initiate a lot of rock falls and
rock avalanches. They can be a significant source of sand-size
grains locally. For example, Pollet and Schneider (2004) showed
that a rock avalanche produced up to ~10–15% in volume of fine
particles (< 1 cm, mean size is ~30 m km) with a significant pro-
portion of sand. On Venus, the net contribution of rock falls and
slides/rock avalanches into sand production can be greater than
on the Earth due to the absence of the most powerful sand sources.

Wind abrasion is the physical process of erosion by the impact
of wind-blown saltating particles onto a bedrock surface. In areas
covered by either consolidated sedimentary or crystalline rocks
wind abrasion is dominated factor in the wind erosion for remov-
ing of material from bedrock surfaces (Laity and Bridges, 2009;
Laity, 2011). Depending on the bedrock type, abrasion can produce
a significant amount of sand-size particles. On the Earth, abrasion
rates of consolidate rocks range from 0.004 to 0.4 mm/yr
(Rohrmann et al., 2013). On Mars, this process might be more
effective due to a higher salination velocity (Laity and Bridges,
2009), while on Venus with its lower saltation velocity, abrasion
is expected to be less effective. Nevertheless, the Venus Wind Tun-
nel experiments (Marshall et al. 1991) showed that rock abrasion
does occur under Venus condition; therefore, it can be a source of
sand material. The abrasion rate on Venus is not known: it depends
on availability of saltating particles and physical properties of the
surface material.

4.2. Sinks of sand

The absence of the main terrestrial source of particulate mate-
rial, erosion by water, was used by Weitz et al. (1994), Greeley
et al. (1997) to explain the apparent scarcity of aeolian deposits.
This logic, however, is imperfect, because the absence of water
means also the absence of main terrestrial sink of sand capable
of saltation: transport by flowing water and sedimentation in
lakes, seas and oceans. Therefore, particulate material, being pro-
duced at lower rates, may lasts longer.

4.2.1. Lithification

One of the possible final fates for the loose particulate material
on Venus is its lithification, consolidation of loose sand into rock.
On the Earth, lithification through chemical cementation usually
involves precipitation from aqueous solutions, which is impossible
on Venus. Burke et al. (1994) have suggested that formation of cal-
cite (CaCO3) and/or anhydrite (CaSO4) by interaction of Ca-bearing
silicates with atmospheric CO2 and/or SO can produce chemical
cementation on Venus in the absence of aqueous solutions. Their
laboratory demonstration of the process, however, involved
artificial CaO as a source material instead of silicates (Burke
et al., 1994).

Sintering also called “cold welding” is a physical process leading
to lithification, which is much more effective on Venus than on the
Earth. The physics of this process involves plastic deformation (due
to dislocation migration) and viscous flow (due to diffusion) at par-
ticle contacts in response to molecular (capillary, surface tension)
forces. Theoretical consideration of such processes has been pub-
lished by Starukhina (2008); her results for the case of the lunar
regolith showed good correspondence to its known properties.
In particular, the capillary forces are responsible for small amount
of small (micron size) particles in natural powders. On Venus,
due to a higher temperature and therefore, higher diffusion coefficients, the solid-state flow at grain contacts is much more effective. Assuming temperature of 450 °C and corresponding diffusion coefficient of $\sim 10^{-7} \text{ cm}^2/\text{s}$, we estimated time for diffusion smoothing of particles' shapes according to Starukhina's (2000) consideration. Surface irregularities having sizes of 1 μm can be smoothed in ~2 days, 10 μm in ~50 years, and 100 μm in ~5 $\times$ 10$^5$ years. These estimates show that on Venus sintering can be responsible for geologically quick lithification of loose sand. The stickiness of sand-size particles observed under Venusian conditions by Marshall et al. (1991) in their wind tunnel experiments was explained by "cold welding", the same mechanism based on plastic flow at grain contacts as discussed above.

4.2.2. Disintegration

Sand-size particles capable of saltation are disintegrated through mutual collisions. On the Earth, quartz sand is abundant due to its tolerance to collisions. Sand particles of basaltic composition are much more fragile. However, on Mars there are abundant aeolian bedforms of generally basaltic composition; there is observational evidence of sand migration over hundreds of kilometers on Mars, however, it looks like the transport is not global and the majority of basaltic sand deposits have local sources (Tirsch et al., 2011). On Venus, the saltation velocities are significantly slower than on the Earth and Mars (Section 2.2 above); therefore, the rate of disintegrations is expected to be significantly lower, and we expect that basaltic sand-size particles are potentially capable of migrating over regional distances.

4.3. Evidence from landers

Panoramas taken by Venera-9, -10, -13 and -14 landers (Fig. 4) provide direct view of Venus surface. Loose material ("soil") is abundant in Venera-9 and -13 landing sites. Individual particles are not resolved in the panoramas, and soil particle size is essentially unknown: the particles are granule-size (a few mm) or finer. The presence of sand-size particles is not directly observable, however, the presence of such fraction in this soil does not contradict observations. The clumps of soil removed by wind from Venera-13 lander (Selivanov et al., 1982) (see Section 3.1) are likely to be made of sand-size particles. The soil has been thought to form by degradation of local rocks (e.g., Florensky et al., 1977; Garvin et al., 1984; Basilevsky et al., 1985) or airfall deposit of distal impact ejecta (Basilevsky et al., 2004).

Layered rocks or platy slabs are present at all four landing sites (Florensky et al., 1977, 1983; Basilevsky et al., 1985). Mechanical properties of surface material at Venera 13, -14 landing sites derived from lander touchdown dynamics (Basilevsky et al., 1985) and penetrometer data (Remurzhchan et al., 2017) indicate that these rocks are porous and mechanically weak. They have been interpreted as heavily altered lavas (Garvin et al., 1984) or indurated/lithified sediment (e.g., Florensky et al., 1977; Basilevsky et al., 1985). Under the latter interpretation, indurated/lithified aeolian bedforms are not excluded, although no unambiguous diagnostic interbedding is observed.

4.4. Summary

Production of sand-size material on Venus is slow, however, the total production through the observed geological history exceeds a centimeter of equivalent global layer and possibly much more. Sand beds are geologically quickly lithified by sintering, however, they might be disintegrated back into sand by aeolian abrasion and impact-caused atmospheric shocks.

5. Observational evidence of aeolian transport and aeolian deposits

5.1. Active wind transport

Progressive removal of soil particles from parts of Venera-13 lander (Selivanov et al., 1982) (see Section 3.1) is the only observational evidence of active wind transport on Venus. The moved material, however, is soil particles initially mobilized by landing, therefore it is not an observation of transport in natural environment.

About a half of Venus surface was imaged with Magellan radar more than once, typically with ~8 or 16 months period between images. Comparison of the images has not revealed any surface changes of either origin. This is not surprising: the resolution of the radar images was ~200 m, and the observational geometry for successive images was different, which means that the smallest changes that can be reliably identified are of ~1 km spatial scale. The rates of morphological surface change in the regions of the most active aeolian transport on the Earth and Mars (Bridges et al., 2012) are on the order of several meters per year, well below the Magellan detection threshold. Similarly, no difference between Magellan images and Venera-15, -16 radar images taken ~8 years earlier can be reliably attributed to surface changes, because the comparison is hindered by the difference in imaging geometry.

The thin saltation layer (Section 2.2) is not observable with remote sensing techniques through the thick Venus atmosphere even in principle. Dust lifting might accompany saltation. Panoramas from Venera landers and optical measurements during lander descent indicate that the lower atmosphere is typically not dusty. Some temporal variability of night-side near-infrared emission from the surface detected with Venus Express instruments (e.g., Mueller et al., 2008; Basilevsky et al., 2012) might be caused by dust storms, however, its conventional explanation is variability in reflection from the dense inhomogeneous lower cloud deck.

In summary, no evidence of natural active sand transport is observed on Venus; however, such evidence is not expected given limitations of the data available.

5.2. Direct imaging of aeolian bedforms

Surface panoramas taken by Venera-9, -10, -13, -14 (see Fig. 4, Section 4.3) do not show the presence of obvious aeolian features in the close vicinity. The clumpy appearance of soil in Venera-9 and -13 landing sites is not consistent with terrestrial examples of well-sorted active sand.

Fine layering of rocks (Section 4.3) in the landing sites might be consistent with indurated or lithified paleo-aeolian bedforms. The diagnostic cross-bedding is in the broken layered rocks is not observed, however, if present, it would be difficult to identify in available images.

Two fields of transverse dunes have been identified in Magellan radar images, namely, Al-Uzza Undae (68°N, 90°E), previously referred as Fortuna-Meshkenet dune field, and Menat Undae (25°S, 339°E), previously known as Aglaonis dune field.

On the radar images, Al-Uzza Undae (Fig. 7) appear as a set of sub-parallel lineaments 0.5–10 km long with ~0.5 km spacing (Weitz, 1994) over an area of ~17,000 km$^2$. Their identification as a field of transverse dunes is solely based on the spatial pattern of these lineaments. No diagnostic morphological details of the dunes are resolved. Our analysis showed that radar signatures are inconsistent with the presence of slip faces of significant extent; estimations by Lorenz (2015) yielded low dune aspect ratios that also suggest the absence of steep slopes. Thus, it seems probable that this dune field is currently not active. About 40
radar-bright linear wind streaks superposed over the dune filed are approximately orthogonal to the lineaments, which is consistent with their interpretation as transverse dunes. The orientation of the wind streaks indicates south-easterly winds.

Al-Uzza Undae are located in a wide depression between Ishtar Terra and Mheshkenet Tessera within a peripheral part of a dark diffuse parabola (see Section 4.1.2) associated with crater Jadwiga (68.4\(^\circ\)N, 91.0\(^\circ\)E; \(D \approx 13\) km). It is not excluded that the parabola material is the source of saltating sand for this dune field, however, a characteristic mean sand thickness of tens of meters inferred from Lorenz (2015) results seems too thick for the radar-transparent peripheral part of a parabola associated with a small crater. Alternatively, deposition of granule-size material of Jadwiga parabola might quench pre-existing dunes. In this case the sand source can be a parabola associated with a much larger crater La Fayette (70.25\(^\circ\)N, 107.5\(^\circ\)E, \(D \approx 39\) km); the highly tectonized nearby tesserae could be considered as a possible source for the material forming these dunes.

From rough estimates of the dune height (Lorenz, 2015) and the total dune area, the total sand volume in Al-Uzza Undae is a few hundreds of cubic kilometers, or on the order of 0.1 \(\mu\)m equivalent global layer. This volume is negligibly small in comparison to the whole impact-generated sand volume on Venus, however, it exceeds the whole volume of the La Fayette parabola-forming material, tens of cubic kilometers, as estimated by Lorenz (2000) on the basis of the models by Schaller and Melosh (1998). This may indicate a significant role of non-impact sand sources.

Spacing of Al-Uzza dunes is narrower than the probable thickness of the convective boundary layer discussed in Section 2.3, which might indicate that the dunes have not evolved to the maximal possible size. On the other hand, it is probable that at this high latitude the boundary layer is thicker than in Vega-2 data at the equator (because the diurnal temperature amplitude is lower), and the rather uniform dune spacing here reflects the local boundary layer thickness analogously to the largest sand seas on the Earth, Mars and Titan.

Menat Undae cover a much smaller area, \(\sim 1300\) km\(^2\), and consist of much smaller, barely resolved transverse dunes (Fig. 8). The dunes exhibit speckle-like appearance in the images consistent with west-facing slip faces. Individual dunes have length of \(\sim 200–300\) m or longer and are narrow (possibly, below \(\sim 150\) m resolution). Abundant wind streaks in the area indicate easterly winds consistent with west-facing slip faces. Menat Undae are located within a parabola and an extensive set of diffuse radar features and wind streaks associated with a large crater Carson (24.2\(^\circ\)S, 344.1\(^\circ\)E, \(D \approx 39\) km). The dune field is also located close to three older large craters Aglaonis (26.4\(^\circ\)S, 339.9\(^\circ\)E, \(D \approx 64\) km), Danilova (26.4\(^\circ\)S 337.2\(^\circ\)E, \(D \approx 49\) km) and Saskia (28.6\(^\circ\)S, 337.1\(^\circ\)E, \(D \approx 37\) km), therefore the place is favorable for a large amount of impact-generated sand.

The total area covered by resolvable aeolian bedforms on Venus is \(\sim 0.004\%\) of the total surface area, vanishingly small in comparison to Mars (\(\sim 0.7\%\)), the Earth land surface (\(\sim 1.5\%\)), and Titan (\(\sim 17\%\)). These numbers are the result of our very crude estimates of the area of dune fields that would be identifiable on these planetary bodies at Magellan image resolution.

5.3. *Indirect remote sensing evidence: backscattering anisotropy from radar images*

Indirect evidence for the presence of asymmetric aeolian bedforms comes from observations of radar backscattering anisotropy. A wide zone in the southern midlatitudes of Venus was imaged by Magellan twice at similar radar looking angles but from opposite directions: from the west during the first cycle of Magellan mission and from the east during the second cycle. Areas with a significant differences between east-looking and west-looking images have been found in the vicinity of three craters, namely, Stowe (43.2\(^\circ\)S, 233.0\(^\circ\)E, 75.3 km), Eudocia (59.1\(^\circ\)S, 201.9\(^\circ\)E, 25.9 km) and Guan Daosheng (61.1\(^\circ\)S, 181.8\(^\circ\)E, 43.0 km) (Weitz et al., 1994). The differences in radar cross-sections reached 7–9 dB. Such a difference was interpreted to occur due to presence of microdunes, which are unresolved in radar images. As an example, crater Eudocia with surrounding surface is shown in Fig. 9. While some diffuse boundaries between radar-bright and radar-dark terrains are identical in these images, east-looking image (Fig. 9, left) has areas with enhanced brightness in comparison to west-looking one (Fig. 9, right).

Weitz et al. (1994) invoked Bragg scattering to explain how the microdune fields with spacing comparable to the wavelength produce strong radar echo in certain directions. Bragg scattering requires the coherence length of a spatial structure to exceed or be comparable to the resolution element. For 10–20 cm microdune spacing (the smallest expected microdunes on Venus, Section 2.3) and \(\sim 100\) m resolution of the Magellan imaging radar, the required coherence length is \(\sim 500 \times \) spacing of the spatial pattern. Natural geomorphological patterns never have such high coherency. The statistical measures of regularity of some terrestrial and martian dune patterns reported by Bishop (2007, 2010) correspond to the coherence lengths of \(\sim 2\times\) spacing and shorter; our measurements of extremely regular transverse dunes in Olympia Undae, Mars yielded a coherence length of \(\sim 4–5\times\) spacing. Thus, Bragg scattering cannot account for the observed phenomenon.

A better explanation for the enhancing of radar echo strength by microdune fields could be that the microdunes cover a significant proportion of the surface and their steeper slopes are as steep as the angle of repose (\(\sim 30\%\)), which suggests fields of active transverse dunes with west-facing slip faces.

Weak backscattering anisotropy of 0.5–2.0 dB in the east-west direction is widely observed over plain areas as reported by Kreslavsky and Vdovichenko (1999). The weak anisotropy has been interpreted as a result of the presence of unresolved asymmetric aeolian bedforms (Kreslavsky and Vdovichenko, 1999).

5.4. *Indirect remote sensing evidence: backscattering anisotropy from the Doppler Centroid*

Another indication of backscattering anisotropy came from Magellan radar altimeter experiment. The Doppler shift of the radar echo allows separation of contributions from parts of the radar footprint located ahead and behind the sub-spacecraft point along the orbit, approximately in the north–south direction. For isotropic backscattering, the echo is symmetric with respect to the sub-spacecraft point, which corresponds to zero Doppler centroid shift of the echo, \(f_D = 0\). Skewed echo means non-zero \(f_D\), which thus can be considered as an indirect measure of backscattering anisotropy in the south-north direction (Tyler et al., 1992). Unlike backscattering anisotropy in radar images, which is most sensitive to steep (25–45\(^\circ\)) facets, the Doppler centroid is mostly sensitive to directional prevalence of slightly tilted (2–5\(^\circ\)) meter-size and larger facets. A global map of Doppler centroid is shown in Fig. 10 along with the global radar image mosaic.

For regions with significant regional slopes, the Doppler centroid reflects the north–south component of those slopes. Tyler et al. (1992) noted that in the plains, in the absence of regional slopes, the Doppler centroid measurements are not randomly scat-
tered, but form large consistent regions of dominated positive or negative values and is related to asymmetry of the small-scale surface topography. Bondarenko et al. (2006) showed that boundaries of those regions often coincide with geologic boundaries and/or boundaries of surficial deposits. They argued that the asymmetric topography responsible for the observed anisotropy is made by meter-scale and/or larger aeolian bedforms. In particular, Bondarenko et al. (2006) showed that the radar-dark diffuse features including crater-related parabolas and haloes usually exhibit isotropic scattering with $f_D \approx 0$. The peripheral parts of crater-related deposits that do not have pronounced dark signature in the radar images often have significant N-S slope asymmetry (Kreslavsky and Bondarenko, 2015).

There is a global hemispherical trend of the N-S scattering anisotropy clearly seen in the Doppler centroid map (Fig. 10a): in the northern hemisphere larger areas typically have $f_D > 0$, while on the southern hemisphere the sign is typically the opposite, although numerous exceptions of this trend exist. The observed hemispheric signature corresponds to a greater area of gentle pole-facing slopes; in other words, the asymmetric small-scale topography has relatively steeper equator-facing slopes. If these small-scale features are small transverse dunes or barchans with gentle stoss sides and steep lee sides, the observed hemispheric signature corresponds to equatorward global wind pattern consistent with Hadley cell circulation (see Section 3.2 above) and coinciding with the hemispheric signature of wind streak orientation (see Section 5.5 below). Consistent patches of significant backscattering anisotropy of the opposite sense might be related to local deflection of prevailing strongest winds from the global Hadley circulation pattern caused by local to regional topography. Alternatively, they might represent fields of lee dunes having gently-sloping lee sides.

5.5. Wind streaks

Wind streaks on Venus are well seen in the radar images and are abundant in some regions. They have been systematically studied, classified, and catalogued (about 6000 individual features) by Greeley et al. (1995, 1997). Wind streaks exhibit a variety of shapes and appearances; they may be radar-bright with respect to their surroundings, radar-dark, and mixed. Their spatial dimensions range from the resolution limit to hundreds of kilometers. A few examples of wind streaks are in Figs. 7, 8, 9, 11c.

On the Earth and Mars wind streaks seen in visual orbital images are often very thin dust veneers and are formed by dust rather than sand transport. Such features would not be seen in radar images. In the most favorable arrangements, the deposits need to be at least a centimeter thick on average to be distinguishable in Magellan radar images. It is more probable that they are related to decimeters to meters of surficial material. Such thicknesses are more likely to be related to sand rather than dust transport.

Greeley et al. (1995, 1997) have interpreted the radar-bright streaks as areas where the wind has removed mantling sediments, exposing rougher substrates. Radar-dark streaks were interpreted to be rather smooth deposits of aeolian sediments (sand sheets). The Earth and Mars give many examples of features that are seen as wind streaks at Magellan-like resolution and actually are barchan chains, bunches of seif dunes, gaps in dune fields (e.g., Fig. 11), etc. We believe that at least some wind streaks on Venus may be fields of small aeolian bedforms or gaps in such fields. Variations of weak backscattering anisotropy found for some wind streaks by Kreslavsky and Vdovichenko (1999) are consistent with such nature of wind streaks.

Greeley et al. (1995, 1997) have interpreted one set of narrow sharp high-contrast streaks as yardangs. Magellan image resolution is insufficient to establish such interpretation reliably: for example, unresolved barchan chains are equally consistent with the observed radar signature. Nevertheless, some wind streaks certainly can be produced by aeolian erosion.

Spatial distribution of wind streaks is highly inhomogeneous. The majority of streaks were found in latitudinal bands of 17°S to 36°S and 5°N to 53°N (Greeley et al., 1992, 1995). Wind streaks are mostly found in relatively smooth plains, possibly, because it is difficult to distinguish them in complicated patterns of dissected terrains. Their occurrence seems independent on elevation. A high concentration of wind streaks is observed in a large region surrounding crater Mead (12.5°N 57°E, $D = 270$ km), the largest impact crater on Venus, which is also probably relatively young. This concentration can be caused by the presence of abundant sand-size particles created or re-activated by this unique impact.

Wind direction distribution inferred from the wind streaks is strongly anisotropic. Greeley et al. (1995) distinguished a subset of wind streaks that are spatially associated with craters and crater-related diffuse features (an example is shown in Fig. 8). They have westward inferred wind orientation. Such streaks have been interpreted as results of impact events and westward transport due to the atmospheric superrotation. It is possible that the material transport related to formation of such streaks occurs high in the atmosphere rather then by saltation; therefore, such wind streaks are formed by air fall deposits rather than by saltation.

The rest of the wind streaks shows prominent hemispheric trend of the inferred wind direction: winds are directed toward the equator in both hemispheres. This direction is consistent with surface winds related to a Hadley circulation in the lower atmo-
A small set of short streaks are related to relatively steep slopes, and the inferred wind direction is downslope. Such features may be wind streaks produced by catabatic winds, however, they also may be aprons of mass-wasted material (Greeley et al., 1995).

In summary, wind streaks in radar images indicate that a significant amount of material was transported by winds. Some of them may be actually fields of small-scale aeolian bedforms or gaps in such fields.

5.6. Possible crater infill

Large impact craters on Venus could be roughly divided into two groups: with radar-bright and radar-dark floors (Fig. 12; examples of bright-floor craters are also in Figs. 5 and 9). Craters with radar-dark parabolas, presumably, the youngest ones, usually have bright floors. Craters with dark floors usually do not have associated diffuse dark features, therefore, they are presumably relatively old. On the Moon, fresh large craters also have radar-bright floors.

Herrick and Rumpf (2011) have found that dark-floor craters are systematically shallower than the bright-floor craters, although the variability of the depth within both groups is high, and the systematic difference is comparable to the formal depth measurement accuracy. Herrick and Rumpf (2011) have explained this trend to be due to a modest volcanic infill of the dark-floor craters, which simultaneously made craters shallower and dark (similar to regional plains) in the radar images. An alternative explanation would be gradual infill of craters with aeolian sediment transported by either saltation or suspension and trapped in the steep-walled depressions of the craters. The existence of the whole range of crater floor radar cross-section from typical bright to radar-dark (Herrick and Rumpf, 2011) is more consistent with the latter mechanism. Accumulation of hundreds of meters of sand in the craters, if it indeed occurs, would indicate the presence of pervasive aeolian transport on Venus.

5.7. Summary

While resolved dunes occupy only a tiny area, the unresolved aeolian bedforms are ubiquitous and cover a significant part of plains on Venus, as shown by the observed anisotropy of backscattering and abundance of wind streaks in radar images. The suspected infill of large craters with aeolian material, if it indeed occurs, suggests pervasive global aeolian transport, while spatial association of wind streaks with the largest impact craters suggests the opposite.

6. Discussion and attempts of synthesis

6.1. Lack of dunes on Venus

The paper by Weitz et al. (1994) summarizing Magellan mission findings on aeolian bedforms on Venus was titled “Dunes and microdunes on Venus: Why were so few found in the Magellan data?” Schultz (1992) has suggested “lack of winds” explanation; he argued that all terrain-shaping work is done only by shock waves and very strong transient winds in the aftermath of large impact events; this explains the observed spatial association of wind-related features with large impact craters. However, as we saw in Section 3, regular winds on Venus are likely to be strong enough to cause salination of loose sand, and the “lack of winds” explanation does not look viable.

An alternative explanation is “lack of sand”; it has been preferred by Greeley et al. (1997). Spatial association of wind streaks with craters suggests that the large impacts are the main source of sand, and estimates show that the amount of sand produced by impacts is insufficient to cover the whole planet with well-developed large dunes. This explanation works only if we invoke geologically quick lithification of sand and “quenching” wind streaks; otherwise, the spatial association with parent impacts would be dispelled at a time scales much less than 1 Ma, while the youngest large craters are several Ma old.

The ubiquity of weak backscattering anisotropy interpreted as ubiquity of meter-scale aeolian bedforms (Sections 5.3.5.4) suggests a complimentary explanation: aeolian deposits are abundant; however, they are not immediately apparent in the Magellan images. Both the wind tunnel experiments and analogy with subaqueous sand transport indicate that formation of the classic transverse dunes with well-expressed slip faces occur within a limited domain in grain size/wind velocity space (Figs. 1 and 2); therefore, formation of such dunes may be a rare case. This would explain why strongly anisotropic backscattering making the microdunes apparent is observed only in three locations.

Due to the high atmospheric density, the incipient dunes on Venus are small, ~10 cm (Section 2.3); they need to evolve through a long growth sequence before they reach hundreds of meters and
become resolvable in Magellan images. In a sense, this growth sequence is much longer than on the other planetary bodies, where the incipient dunes are larger. It is possible that the dunes on Venus just do not have time to evolve to observable sizes. It is also possible that lack of sand and/or sand bed lithification halt dune growth before they reach their maximum possible spacing of a few kilometers (Section 2.3).

If dunes get larger, the time interval between saltation episodes for each particle increases, which in turn increases the chances for lithification through diffusive sintering. It is possible that there is a threshold size, above which a dune is lithified and become inactive, while active saltation still can occur in thin layers. Therefore, it is possible that there is pervasive ongoing aeolian transport in thin layers of active sand despite the fact that larger features are "quenched" by lithification.

In summary, our preferred answer to the "why so few" question would be that the bedforms are actually abundant and ubiquitous, however, they are small and obscured. It is also not known, whether they are active or not.

6.2. Global sand transport?

Both first-principle considerations of lower atmosphere dynamics (a single Hadley cell circulation, Section 3.2) and observational evidence (in-situ measurements, Section 3.1; consistency of the Doppler centroid signatures, Section 5.4, and wind streaks, Section 5.5, with the single Hadley cell wind pattern) suggest persisting stable near-surface winds on Venus. In this case, the time required for global sand transport is scaled as \( R/h Q \), where \( R \) is the planetary radius, \( h \) is the thickness of sand to be transported, and \( Q \) is the characteristic volumetric sand flux estimated in Section 2.2. For \( h \sim 0.1 \) m, a thickness adequate for incipient dunes on Venus (Section 2.3), this scaling estimate is 1 Ma. This time scale is significantly shorter than the mean surface age (100 s Ma), indicating that the sluggishness of sand transport on Venus does not prevent global mixing of the surface layer. On the other hand, this time scale is sufficiently long to think that sand lithification may stop transport early.

Indirect observational evidence about global sand transport is contradictory. The spatial association of the observed dunes, microdunes, wind streaks with large impact craters suggests that the sand does not travel for long distances, while ubiquity of anisotropic small-scale topography suggests the opposite. As discussed above, it is possible that thick sand deposits forming observable dunes and wind streaks are lithified and quickly become inactive, while active saltation occurs in thin (\( \sim 0.1 \) m) layers providing transport at longer distances, possibly but not necessarily globally. Long-distance aeolian transport of sands would also naturally explain some other observations: generally lower radar contrasts for stratigraphically older terrains (Arvidson et al., 1992), apparent decrease of dielectric permittivity of volcanic surfaces (Bondarenko et al., 2003) and impact crater floors (Campbell et al., 1992) with age.

If global sand transport indeed occurs, the persisting near-surface wind pattern would cause net migration of sand from moderately high latitudes toward equator. The observed absence of wind streaks at high latitudes seems consistent with this prediction. However, the Magellan radar images at high latitudes are obtained as systematically lower incidence angles and have a higher noise level; both these factors are not favorable for wind streak identification; therefore, the apparent absence of wind streaks at high latitudes may be an observational bias.

In summary, at the current state of knowledge, neither the presence of global sand transport, nor its absence can be rejected. New observations are needed to resolve this important question.

7. Conclusion

We reviewed the current state of knowledge about aeolian sand transport and aeolian bedforms on Venus. Near-surface winds routinely exceed the saltation threshold; therefore, sand transport would be pervasive, if sand capable of saltation is available. Sand production on Venus is slow on average; sand is produced by meteoritic impacts, and also possibly by volcanism and slow chemical weathering of the surface. Aeolian transport of sand is probably limited by lithification. Dunes detectable in currently available radar images occupy a tiny area, however, unresolved aeolian bedforms probably are ubiquitous. It is not clear, whether winds move sand material over long distances causing regional or global contamination of the surface layer, or sand transport is local and limited in time to some aftermath of large impacts.

Understanding of the spatial and temporal scale of sand transport is the key question in understanding of aeolian processes on Venus. This question is critical for future scientific exploration of the planet, in particular, for implementation and interpretation of geochemical studies of surface materials.

Obviously, comprehensive and detailed understanding of the surface-atmosphere interaction on Venus, including aeolian transport, requires a wide complex of observations from focused in-situ geological studies at carefully selected sites at the surface to global long-lasting meteorological network to global uniform high-resolution imaging. However, due to thick atmosphere and harsh conditions at the surface, exploration of Venus is difficult and expensive, and there is no hope for complex extensive exploration program in the foreseeable future. A focused in-situ geological study at a carefully selected landing site is hardly possible, because currently available data are insufficient for selecting such a site. Therefore, the first breakthrough in understanding aeolian transport requires remote sensing studies. Remote optical imaging of the surface in the narrow near-infrared spectral windows, where the atmosphere is not absorbing, is only possible from low-flying balloons, a few km above the surface, due to intensive scattering by the dense atmosphere (Moroz, 2002). This significantly limits coverage and requires long-term operations under very high temperatures, which makes such investigations difficult to implement. Microwave radar imaging from the orbit is the most useful and affordable tool for studies of aeolian processes and bedforms (Kreslavsky et al., 2014). A significant increase in resolution in comparison to Magellan is essential for a breakthrough in aeolian bedform studies. Interferometric radar capabilities would be very beneficial, because they allow detection of surface changes with time below the spatial resolution limit and therefore facilitate detection of active bedforms. We hope that such break-through observations will be taken during lifetime of the present-day generation of researchers.

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References

