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Probability distribution of surface wind speed induced by convective adjustment on Venus

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ABSTRACT

The influence of convective adjustment on the spatial structure of Venusian surface wind and probability distribution of its wind speed is investigated using an idealized weather research and forecasting model. When the initially uniform wind is much weaker than the convective wind, patches of both prograde and retrograde winds with scales of a few kilometers are formed during active convective adjustment. After the active convective adjustment, because the small-scale convective cells and their related vertical momentum fluxes dissipate quickly, the large-scale (>4 km) prograde and retrograde wind patches remain on the surface and in the longitude–height cross-section. This suggests the coexistence of local prograde and retrograde flows, which may correspond to those observed by Pioneer Venus below 10 km altitude. The probability distributions of surface wind speed V during the convective adjustment have a similar form in different simulations, with a sharp peak around $\sim 0.1\ \mathrm{m\,s^{-1}}$ and a bulge developing on the flank of the probability distribution. This flank bulge is associated with the most active convection, which has a probability distribution with a peak at the wind speed 1.5-times greater than the Weibull fitting parameter c during the convective adjustment. The Weibull distribution $P(>V)$ $(=\exp[-(V/c)^k])$ with best-estimate coefficients of Lorenz (2016) is reproduced during convective adjustments induced by a potential energy of $\sim 7\times 10^7\ \mathrm{J\,m^{-2}}$, which is calculated from the difference in total potential energy between initially unstable and neutral states. The maximum vertical convective heat flux magnitude is proportional to the potential energy of the convective adjustment in the experiments with the initial unstable-layer thickness altered. The present work suggests that convective adjustment is a promising process for producing the wind structure with occasionally generating surface winds of $\sim 1\ \mathrm{m\,s^{-1}}$ and retrograde wind patches.

1. Introduction

Surface wind is an important element of surface–atmosphere interactions on terrestrial planets. However, in the case of Venus, the surface wind has not been observed over the whole planet. A small number of observations has shown near-surface flows with speeds of $\leq 1\ \mathrm{m\,s^{-1}}$ (Counselman et al., 1980; Kerzhanovich et al., 1980, 1982; Ksanfomaliti et al., 1982). Retrograde flow with respect to the surface rotation was observed in the vertical profile of the zonal mass flow below 10 km, although the wind profile was noisy (Schubert et al., 1980). If such a retrograde zonal flow is maintained at the surface, it acts as the source of the angular momentum of Venus' superrotation.

The wind streak structures on the surface also provide evidence for large-scale wind flow near the surface (Greeley et al., 1994). The presence of the large-scale surface wind structure was also supported in the global maps of near-surface winds obtained from Venus general circulation models (GCMs). Herrnstein and Dowling (2007) reported that the wind in the surface layer has a maximum speed of $\sim 27\ \mathrm{m\,s^{-1}}$ in their GCM with topography. According to Yamamoto and Takahashi (2009), in the absence of topography, simulated wind speeds are much lower than $1\ \mathrm{m\,s^{-1}}$ in the lowermost level of their GCM. In contrast, when topography is considered, near-surface winds with speeds of a few meters per second are seen in and around highlands and mountains. At the present stage, because there are large differences in the surface wind among the GCMs and in-situ observations, it might be difficult to use the GCMs for investigating the global and in-situ structures of the near-surface wind.

Recently, Lorenz (2016) estimated a probability distribution of the surface winds from in situ measurements using a Weibull probability function $P(>V)=\exp[-(V/c)^k]$ with $c=0.8$ and $k=1.9$ (where V is the surface wind speed) and suggested that the winds are strong enough to move dust and sand. The Weibull function has been widely applied to the probability distributions

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of surface wind speeds on Mars and Titan (Lorenz, 1996; Fenton and Michaels, 2010; Lorenz et al., 2012a,b). However, dynamical processes that might produce the Weibull distribution have not been fully elucidated for the Venus case. The present study investigates the impact of near-surface convection on the probability distribution.

According to Young et al. (1987) and Izakov (2002), unstable layers were observed near the Venusian surface. The presence of the unstable near-surface layers is expected to influence thermal, material, and wind distributions via convective adjustment. Here, convective adjustment means that statically unstable state is relaxed to neutral state via convective motions. Although the relaxation process of the air temperature to the adiabatic lapse rate is parameterized in GCMs, mesoscale regional models directly simulate convective motions leading to the adjustment process without the simplified convective-adjustment parameterization. Direct simulations of atmospheric and oceanic convective adjustments for Earth have been conducted in previous works (e.g., Dietich 1977; Dietich and Lin 2002). Yamamoto (2011) examined thermal and material transports due to convective adjustment near Venus' surface using an idealized weather research and forecasting (WRF) model with a grid size of 100 m. The WRF simulation resolved the convective adjustment process with scales of ~1 km and ~1 h. However, in discussing the near-surface wind structure induced by convective adjustment, the author needs to examine the sensitivity to the model domain size. Yamamoto (2011) briefly reported that the vertical eddy momentum transport is sensitive to the model domain size. Thus, we must carefully check the sensitivity of the wind structure to the model domain size in the present study.

The main goal of the present work is to explain the formation of the retrograde zonal flow in the superrotational background flow (Schubert et al., 1980) and the probability distribution of the surface wind (Lorenz, 2016). The present work focuses on the convective adjustment with scales of ~1 km and ~1 h under the Venusian condition that the background zonal wind is much weaker than the convective wind speed near the surface (i.e., extremely slow background wind). The model setting is described in Section 2. To determine the model domain size used for the simulation of the convective adjustment, the sensitivity of the convective adjustment to the model domain size is investigated in Section 3.1. After this, the structure and probability distribution of the surface wind induced by the convective adjustment are investigated in Sections 3.2–3.4, before the concluding remarks (Section 4).

2. Model settings

The WRF community model (Skamarock and Klemp, 2008; Skamarock et al., 2008) has been applied to planetary atmospheres as Planet WRF (Lee et al., 2006; Richardson et al., 2007; Newman et al., 2011) and used for idealized experiments in planetary fluid dynamics (Yamamoto, 2011, 2014; Dias Pinto and Mitchell, 2014). The present study simulates eddies with scales of ~1 km under Venus' near-surface conditions using an ideal LES (Large Eddy Simulation) case of the Advanced Research WRF (ARW) model (version 3.1), which is a fully compressible and nonhydrostatic model that uses the third-order Runge–Kutta scheme for time integration and a staggered Arakawa C-grid. Based on the Venus international reference atmosphere (Seiff et al., 1985), the physical constants and parameters in the model are changed to the Venusian ones. The surface pressure (P_S) is 92,000 hPa, the acceleration of gravity (g) is 8.87 m s^{-2}, the gas constant (R) is 191.4 J kg^{-1} K^{-1}, and the specific heat at constant pressure (c_p) is 1181.0 J kg^{-1} K^{-1}. The Coriolis parameter is set at zero because of the slow planetary rotation. A Deardorff TKE scheme is used to compute the subgrid-scale eddy viscosity and eddy diffusivity, and a Monin–Obukhov scheme (MM5 similarity) is chosen as the surface-layer option (Skamarock et al., 2008). To determine the appropriate domain size, the sensitivity experiments to the model domain size (Exps. SS1, S1, C1, L1) are conducted in the present work. The model domain sizes and resolutions are listed in Table 1. A double periodic boundary condition is used along x (along a line of latitude) and y (along a line of longitude). The surface upward flux of a passive tracer at the bottom boundary is defined as zero. The frictional velocity is computed by the surface-layer scheme.

In the Venusian radiative transfer models (Matsuda and Matsuno, 1978; Takagi et al., 2010; Lee and Richardson, 2011), the convective adjustment is used to prevent from exceeding adiabatic lapse rate. If the convective adjustment is not considered, the unstable state is easily formed in the radiative equilibrium in the Venus lower atmosphere (Matsuda and Matsuno, 1978; Takagi et al., 2010). In fact, the observations suggest the presence of the unstable layer (Young et al., 1987; Izakov, 2002) leading to the convective adjustment. The unstable layer might be formed by the surface heat flux and radiative transfer during a relatively long time over the global area. However, because it is difficult to simulate these long-term processes in the present model with a grid spacing of 100 m and a time-step size of 0.25 s, the present work investigates convective adjustment during a few hours after forming an unstable layer.

The radiative and diurnal processes are excluded because the simulation time is much shorter than these characteristic scales (10,000 Earth days for infrared radiation and 117 Earth days for a Venusian solar day). The heat budget at the surface depends on the surface net radiative fluxes, geothermal heating, and wind speed. The local and transient imbalance of the surface heat budget is considered as the surface heat flux magnitude Q. In the experiment with surface upward heat flux of 76.5 W m^{-2} (0.001 K m s^{-1}) of Yamamoto (2011), active convection with wind magnitudes of ~1 m s^{-1} does not occur. The surface heat flux is very small compared with the eddy heat flux in the convective adjustment experiments (0.1–1 K m s^{-1}). Under the condition of extremely slow background flow ($V \sim 0.1$ m s^{-1}), surface sensible flux is also very small. For the air-surface temperature difference of ~1 K and the bulk coefficient of 0.002 (Rossow 1983), $Q \sim 0.0002$ K m s^{-1}. This means that the surface net radiative flux (e.g., Revercomb et al., 1985) and sensible heat flux do not directly influence the convective adjustment for a few hours. Thus, for simplicity, the surface upward heat flux is set to zero in the experiments of convective adjustment and the surface temperature is assumed as the initial surface air temperature.

The parameter Γ_θ, an initial 'lapse' rate of potential temperature θ, is defined as $-d\theta/dz$ in an initially unstable layer below 2 km, which is capped by a stable layer. This definition means that positive lapse rate ($\Gamma_\theta > 0$) is unstable (i.e., the static stability $d\theta/dz$ is negative). The parameter range of Γ_θ is based on Yamamoto (2011). Although negative stability with a magnitude of ~1 K km^{-1} has been observed near the surface (Young et al., 1987; Izakov, 2002), there are insufficient observations to give typical values for Γ_θ. To investigate the sensitivity of the convective adjustment to the initial lapse rate, a wide range of Γ_θ (0.5–4 K km^{-1}) is assumed here. In addition, to kick off the convection, a random perturbation of ±0.05 K is imposed on the initial temperature field at the lowest four grid levels.

The initial profiles of the potential temperature are shown in Fig. 1 and the model conditions are listed in Tables 1 and 2. The convective adjustment near Venus' surface is simulated in sensitivity experiments with Γ_θ altered (Exps. C0.5 to C4 in Table 1, Fig. 1a). To evaluate the vertical mixing of minor constituents near the surface, the initial step function of a passive tracer is assumed (Fig. 1b). The area-mean vertical distribution and eddy vertical flux magnitude are used to compare the sensitivity experiments in the present work and Yamamoto (2011).