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VERTICAL TRANSPORT OF WATER ICE AT LOW LATITUDES IN THE MARTIAN ATMOSPHERE

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Abstract. Solar occultation measurements made from the *Phobos 2* spacecraft show that two condensation levels for H_2O are generally present at respectively ≈15 km and ≈50 km altitude at low latitudes. Above ≈15 km, a regular decrease of the water vapor mixing ratio up to ≈35 km is observed (Krasnopolsky et al, 1991) and it is interpreted as the signature of a saturated atmosphere. Around ≈50 km, thin clouds, very likely formed of water ice, are detected (Blamont et al, 1991). This means that above ≈35 km exists a region where the atmosphere is not saturated in H_2O, containing enough water vapor to give rise to the formation of clouds. We show that upwelling in the ascending branch of the Hadley cell is able to supply water ice, across the cold saturated region (≈15-35 km), to the non-saturated layer below the clouds, where ice may sublimate and condense again at cloud level (≈50 km). Different constraints may be introduced relating to the coagulation probability during the upward transport phase, the minimum amount of H_2O which must be deposited at the sublimation level and the infrared optical thickness of ice particles. It is shown that these constraints are jointly satisfied for reasonable values of parameters, indicating that a quantity (zonally averaged) of ≈0.01 pr. μm of water ice in the shape of small particles of radius 0.1-0.2 μm might be present in the middle atmosphere (15-35 km) above equatorial regions.

Introduction

Several types of information on the thermal structure of the Martian atmosphere in equatorial regions near the northern spring equinox were obtained from solar occultation measurements made on board the *Phobos 2* spacecraft. Using nine occultations of the Sun by the Martian atmosphere at 1.9 μm wavelength, an average vertical profile of the water vapor concentration between 15 km and 35 km altitude was derived from a weak water vapor band (Krasnopolsky et al, 1991). This profile shows a regular decrease of the H_2O mixing ratio above 15 km which was interpreted as the signature of a saturated atmosphere. The temperature profile was derived assuming saturation conditions. Results obtained above ≈35 km are not reliable because of instrumental problems. In five cases, a thin cloud was observed close to 50 km altitude (Chassefière et al, 1991, called CB). Because of instrumental problems, the low atmosphere was generally not observed simultaneously with a cloud except in one case. This partial overlapping of the two sets of observations, together with the apparent stability of atmospheric conditions derived from solar occultation measurements, shows that two condensation levels are generally present, at respectively 15 km and 50 km. These data are not compatible with a classical temperature profile, i.e. a nearly constant negative thermal gradient over the whole middle atmosphere. An inflexion of the profile must be present around 35 km, with a relative warming beneath, in order to link the low temperatures inferred below 35 km to the relatively high temperature required at the cloud base. An estimate of this temperature was done using a simple steady state model involving settling and eddy mixing as transport processes (CB). A most likely value of ≈170 K at 50 km was found, about 10 K above the temperature derived from the saturated water vapor profile at 35 km.

Several possible temperature profiles, consistent with the temperature inferred at cloud level and the thermal profile deduced from water vapor observations, are shown in Figure 1. There is a likely temperature inversion around 35 km with an amplitude of ≈10± 10 K. Even if no inversion exists, because of the statistical uncertainty, there is a definite inflexion of the thermal profile at 35 km, with a relative warming beneath. It may be attributed to the effect of breaking gravity waves on the general circulation (see e.g. Barnes, 1990), but this question will be dealt with in another paper. The sublimation level, i.e. the boundary between the saturated region and the non-saturated layer below the clouds, is close to 35 km altitude. The existence of this sublimation level is a direct and necessary consequence of the presence of two condensation levels. It is the central premise of this paper.

Formation of high clouds: a dynamical explanation

Possible role of Hadley circulation

In principle, the low temperature between 15 and 35 km prevents H_2O molecules from reaching higher altitudes due to the settling of ice particles. The most straightforward way to explain the permanent supply of water ice to the non-saturated layer below the clouds is upwelling. Upward motions are expected to take place in the middle atmosphere. Near equinox, models show that a global circulation of the Hadley type is established. The wind velocity may be derived from a three-dimensional model like the GCM (Pollack et al, 1990). From GCM wind and temperature plots for the late northern winter and mid northern spring (time and zonally averaged values) kindly provided by J. Barnes, we have extracted the vertical profiles of the upward velocity V_u at the mean latitude of observation, ≈10° N.L. (Figure 2). The vertical profiles are smoothed over one atmospheric scale height. The typical ascending velocity is 0.5 cm s^{-1}. The decrease of the velocity above 30 km is due to the fact that the GCM top is constrained at ≈50 km, velocities being required to be small here. If the GCM top was fixed at higher altitudes, the circulation would almost certainly extend further upwards. We will assume that V_u remains constant above the altitude where its maximum is reached (≈30 km, Figure 2).

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Theoretical and observational constraints

Our aim is to show that enough H_2O can be transported from the low condensation level (z_b≈15 km) up to the sublimation level (z_t≈35 km) in realistic dynamical conditions and to estimate the quantity of water ice involved in this process. The model assumes that vapor and ice are conserved separately, i.e. particles are always in a saturated atmosphere. Thermal coagulation, settling, and advection processes are allowed. GCM results are used to constrain atmospheric velocities. The model works on zonally averaged quantities. A steady state is assumed globally, and justified, in both vertical and meridional directions at low latitudes and in the altitude range 15-50 km. Explicitly free parameters are the particle radius, the number density, and the altitude of the sublimation level. Observed constraints are ice opacity at the cloud level and atmospheric opacity profiles. Consistent solutions for the free parameters are obtained.

Three basic constraints on the radius and the number density of particles may be introduced:

1- The time for particles to move up from z_b to z_t is smaller than their mean coagulation time. The radius of particles, whose size distribution is supposed monodisperse, is considered to a first approximation as constant over the relevant altitude range. In the case of a broad size distribution at $z \approx z_b$, the radius of particles reaching z_t cannot exceed a critical value r_c defined by:

$$\int_{z_b}^{z_t} n(z)\, V_T(z)\, 4\pi r_c^2 \frac{dz}{V_u(z)-V_s(z)} = 1 \qquad (1)$$

where V_T is the average relative thermal velocity of particles of radius r_c and n their number density ($V_T^2 = 16kT\,(\pi m_i)^{-1}$; m_i: mass of particles; Fuchs, 1964). The product $n\, V_T\, 4\pi r_c^2$ is the number of particles colliding with one particle per unit of time, the differential term $dz/(V_u-V_s)$ being the time increment dt. The sticking efficiency is assumed to be 1. If not, r_c is larger and the constraint somewhat alleviated. This equation means that the thermal coagulation probability for particles of radius r_c moving up from z_b to z_t at the velocity (V_u-V_s) is 1. In presence of a large number of particles of different sizes at $z \approx z_b$, the combined effects of coagulation and settling prevents some of them from climbing. Because the size distribution function decreases for decreasing radii, the mass of ascending particles may be supposed to be concentrated around the critical value r_c defined by Eq. 1 and the truncated size distribution to be nearly monodisperse. On the other hand, if only a small number of particles are available at $z \approx z_b$, there is no regulation of particle size and number by coagulation and their average radius is smaller than r_c. Eq. 1 provides an upper limit on the particle radius.

2- The flux $\Phi = n(z_t)\,(V_u(z_t) - V_s(z_t))$ of water ice particles at the top must be large enough to supply the required amount of H_2O to high clouds. The column volume W_i of water ice in clouds (observed five times), averaged over the 38 available occultations, is 10^{-4} mm^3 cm^{-2}, or ≈ 10^{-3} pr. μm, the thickest cloud containing ≈10^{-2} pr. μm. The time τ during which an amount of H_2O equal to the average one present in the clouds is brought from below may be defined in the following way:

$$\Phi \frac{4}{3}\pi r_c^3\, \tau = W_i. \qquad (2)$$

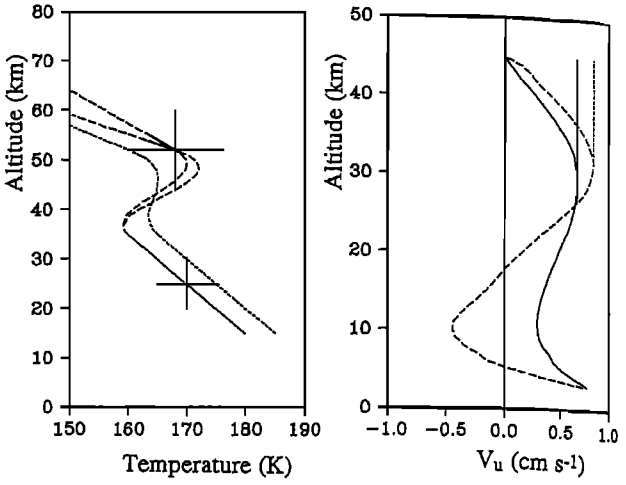


Fig. 1 (left). Three thermal profiles consistent with water vapor and cloud observations (at resp. ≈15-35 km and ≈50 km). The long-dashed curves are obtained from an interpolation by a cubic polynom below the cloud and a linear extrapolation above the cloud for the two extreme values of the thermal gradient (-1.5, -2.5 K km^{-1}; see CB) at the most likely cloud altitude and temperature. If the temperature below ≈35 km is underestimated by 5 K, the temperature inversion is less marked (short-dashed curve).

Fig. 2 (right). Vertical profiles of the upward velocity for late northern winter (dashed line) and mid northern spring (solid line). Thin vertical lines above ≈30 km altitude are profiles retained for the present study. These curves are derived from GCM plots provided by J. Barnes.

Typically, clouds are removed from equatorial regions by the meridional circulation in a time τ_m of the order of L/V_m where L is the distance from the equator to mid-latitudes (≈2500 km) and V_m the meridional wind velocity, in the range 1-4 m s^{-1} following GCM calculations. It results that τ_m ≈ 7-25 days. Moreover, the time τ_u for a small particle to move up from z_b to z_t is of the order of $\Delta z/V_u$ where $\Delta z = z_t - z_b$ (≈20 km) and V_u ≈ 0.4 cm s^{-1} (Figure 2), i.e. τ_u ≈ 50 days. The zonal wind may reach 20 m s^{-1} at 30-40 km altitude, implying that the time τ_z for an air parcel to go round the planet at the cloud altitude is about 10 days, i.e. $\tau_z \approx \tau_m$. Although no rigorous steady state is expected ($\tau_u \approx \tau_m$), these times give an order of magnitude of the characteristic time τ which must be introduced in Eq. 2. The fact that $\tau_z \approx \tau_m$ shows that it is justified to average the water ice content of clouds as is done when establishing Eq. 2. Indeed, water ice particles which are supplied to the region below the clouds originate at all longitudes due to the vertical gradient of the zonal velocity, increasing from ≈10 m s^{-1} near 15 km altitude to ≈20 m s^{-1} at higher altitudes. The value of τ which must be used in Eq. 2 is therefore in the range from 2 weeks ($\approx\tau_m$) to 2 months ($\approx\tau_u$).

3- The tangential optical depth of ice particles around z_m≈25 km altitude at 1.9 μm wavelength must be very small, lower than ≈0.01 (CB). Denoting by σ the extinction cross-section of ice particles, which varies as $\approx r^6$ for small particles, this condition may be written:

$$N(z_m)\, \sigma(r_c) < 0.01 \qquad (3)$$

where $N(z_m)$ is the column number density of ice particles along a tangent line of sight at z_m=25 km.

Vertical distribution of water ice particles

In the steady state, the gas flux ρ **V**, where ρ is the volumic mass of the gas and **V** its velocity vector, obeys the law of conservation of mass:

$$\text{div } \rho \mathbf{V} = 0 \quad (4)$$

where div denotes the divergence operator. The same equation may be written for ice particles. Since there is no coagulation due to constraint (1), the motion of particles, which may be considered as inert in the absence of sublimation and/or condensation, is driven by the fluid atmosphere. The conservation equation for particles may be written:

$$\text{div } \rho_i (\mathbf{V}+\mathbf{V_s}) = 0 \quad (5)$$

where ρ_i is their volumic mass and $\mathbf{V_s}$ their settling velocity. $\mathbf{V}+\mathbf{V_s}$ is the velocity vector, denoted $\mathbf{V_i}$, of these particles. Denoting by m their mixing ratio ρ_i/ρ, Eq. 5 may be written:

$$\text{div } [m (\rho\mathbf{V}+ \rho\mathbf{V_s})] = 0 \quad (6)$$

Since the settling velocity of particles is inversely proportional to the volumic mass of the gas, the product of these two quantities is constant with altitude and $\text{div}(\rho\mathbf{V_s}) = 0$. Using this equality and Eq. 4, Eq. 6 yields:

$$\mathbf{V_i} \cdot \mathbf{grad}\ m = 0 \quad (7)$$

where **grad** denotes the gradient operator. Eq. 7 shows that the gradient of the particle mixing ratio is perpendicular to the particle velocity vector, i.e. m is constant along any stream line of the particle flow. On a global scale, the reasonable assumption that the horizontal distribution of particles is uniform around 15 km altitude, at the foot of the stream lines, can be made. It follows that the mixing ratio of particles of radius r is constant in the region located below the critical altitude such that settling counteracts upwelling (V_s=V_u). It is therefore justified to assume, as an exploratory hypothesis, that the mixing ratio of particles is constant with altitude.

Results

Case without eddy mixing

The constraints derived from Eq. 1, 2 and 3 are shown in Figure 3 using mid northern spring conditions, near the conditions of observation, which are more favourable than those of the late northern winter because there is no reversal of the vertical velocity at low altitudes (Figure 2). n_b is the number density of particles at the altitude z_b: n_b=$n(z_b)$. The two curves corresponding to the non-coagulation and H_2O supply conditions are limited on the ordinate axis by a horizontal asymptotic straight line corresponding to the radius such that $V_s(z_t)$=$V_u(z_t)$. The three conditions cannot be reconciled adopting the GCM vertical profile of V_u in mid northern spring conditions and use a characteristic time τ of 2 weeks. By multiplying by 1.5 the GCM profile in such a way as to obtain about the same maximum as in late northern winter

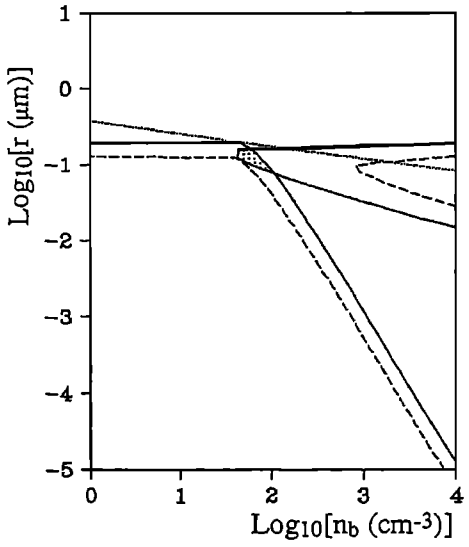


Fig. 3. The values of the radius r and the number density n_b=$n(z_b)$ which fulfil the three constraints are indicated by a grey area, limited by the two curves (solid lines) corresponding to the constraints of non-coagulation (Eq. 1) on the left and of a sufficient supply of H_2O to the clouds (Eq. 2) on the right for τ=2 months and the vertical velocity derived from the GCM in mid northern spring conditions multiplied by 1.5. The oblique dotted line corresponds to the constraint on the tangential optical thickness at 25 km altitude (Eq. 3). Dashed curves are obtained using the GCM velocity in mid northern spring conditions and a characteristic time τ of 2 weeks. There is no solution in this case.

($\approx$0.85 cm s^{-1}) and increasing the time τ up to 2 months, it is possible to reconcile the first two conditions. The third condition on the tangential optical depth at 25 km altitude is also satisfied. Results corresponding to different values of the input parameters, including the sublimation altitude, are presented in Table 1. The most likely number density of particles at z=z_b is $\approx$50-100 cm^{-3}. Their radius is around 0.1-0.2 μm, smaller than 0.4 μm in all cases. For a sublimation altitude of 40 km, the transport of ice particles is inhibited by settling and a large value of the upward velocity is required (3 V_u(GCM) $\approx$ 2 cm s^{-1} around 30 km). The column volume of these particles is $\approx$5 10^{-3} pr. μm, i.e. five times larger than the zonally averaged amount contained in high clouds.

Possible role of eddy mixing

Random motions due to eddy diffusion are not expected to modify the general results obtained from the simplified treatment applied above. Moreover, eddy mixing alone, without upwelling, cannot explain the supply of H_2O to high clouds. The mixing ratio of particles of radius r in a static atmosphere dominated by eddy mixing decreases with an equivalent scale height $H_i \approx H\ V_e\ V_s^{-1}$ where V_e is the eddy diffusion velocity (V_e = K/H; K: eddy diffusion coefficient), and H the atmospheric scale height (see CB for more details). The mixing ratio m of particles at the sublimation level is:

$$m(z_t) \approx m(z_b) \exp\left(-\frac{V_s(z_t)-V_s(z_b)}{V_e}\right) \quad (8)$$

Table 1. Ranges of possible values of the radius r and the number density n_b at 15 km altitude of ice particles as a function of the sublimation altitude z_t, the time τ for the water present in clouds to be brought up, and the ratio V_u/V_u(GCM) between the upward velocity and the GCM upward velocity (mid northern spring). Underlined bold types indicate the most likely solution. Bold types correspond to the extreme case of a strong upward convection together with a long characteristic time and a low sublimation level. Empty spaces correspond to the cases where there is no solution.

z_t (km)	τ (days)	V_u/V_u(GCM)	r (μm)	n_b (cm^{-3})
30	60	3	**0.03-0.4**	**0.5-500**
		1.5	0.05-0.3	4-150
		1	0.1-0.2	20-60
	15	3	0.05-0.4	1-400
		1.5	0.1-0.3	20-100
		1	-	-
35	60	3	0.04-0.3	3-400
		1.5	**0.1-0.2**	**50-100**
		1	-	-
	15	3	0.06-0.3	10-300
		1.5	-	-
		1	-	-
40	60	3	0.05-0.2	20-300
		1.5	-	-
		1	-	-
	15	3	0.1-0.2	100-200
		1.5	-	-
		1	-	-

The upward eddy flux at this level may be approximated by:

$$\Phi \approx n_b \exp(-\frac{\Delta z}{H}) \exp(-\frac{V_s(z_t)-V_s(z_b)}{V_e}) V_e \qquad (9)$$

and the mean residence time τ_r of particles between z_b and z_t is obtained by ratioing the total content of H_2O, $\approx n_b H$, to this flux:

$$\tau_r \approx \tau_e \exp(\frac{\Delta z}{H}) \exp(\frac{V_s(z_t)-V_s(z_b)}{V_e}) \qquad (10)$$

where $\tau_e \approx (H^2/K)$ is the eddy mixing time, ≈20 days assuming $K \approx 5\ 10^5\ cm^2\ s^{-1}$ (CB; Kahn, 1991). The uncertainty on this value is about a factor of 3 on either side. The first exponential term of Eq. 10 is ≈10. The second one is ≈10 within a factor of 10. It follows that the most likely value of τ_r is ≈6 yr. Even using our highest admissible value of K of $\approx 10^6\ cm^2\ s^{-1}$, τ_r (≈300 days) is still much larger than it is in the case of upwelling ($\tau_u \approx$20-50 days). Such a long time favours coagulation and eddy mixing is therefore not a good candidate.

On the other hand, eddy mixing could play a determining role in the release of water ice particles from the ice-rich convective flow to the non-saturated region below the clouds. For dynamical reasons which will be developed in another paper, the inflexion of the thermal profile around 35 km altitude could correspond to the top of the Hadley cell. The eddy diffusion velocity V_e is of the same order as the upward velocity V_u in the last scale height below z_t, suggesting that eddy mixing may remove ice particles from the flow around 30-40 km to inject them in the region beneath where they sublimate, maintaining a high negative gradient of the water ice mixing ratio at the boundary.

Conclusion

On a global scale, the major recycling mechanism for H_2O entering the formation of high clouds observed at low latitudes by solar occultation is upwelling rather than eddy mixing. The same mechanism may act to bring up the smallest dust particles (r_d<0.1 μm), which will serve as condensation nuclei, from the lower atmosphere where they are present in large amounts. A zonally averaged quantity of water ice of nearly 10^{-2} pr. μm, of the same order as the content of the thickest cloud, could be present in the atmosphere in the shape of small solid ice particles of radius ≈0.1-0.2 μm. The tangential optical thickness of these particles is smaller than 0.01 in the near infrared. On a local scale, clouds are sustained by eddy mixing rather than upwelling whose role is to bring water ice from the low condensation level (≈15 km) up to the non-saturated region below high clouds (≈35-50 km). Due to the likely horizontal variability of thermal structure and water vapor concentration in the non-saturated layer, the cloud coverage is not uniform, clouds not being present where temperature is higher and/or water vapor less abundant.

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