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EFFECT OF DYNAMICS ON THE FORMATION OF MIXED PHASE REGIONS IN STRATIFORM CLOUDS

Alexei Korolev and Paul Field

Environment Canada, Toronto, ON, Canada
National Center for Atmospheric Research, Boulder, CO, USA

1. INTRODUCTION

The growth rate of ice particles and liquid droplets under the same environmental conditions varies because of the difference between saturation vapor pressures over ice and water. Conditions commonly exist in the atmosphere where, although condensationally unstable, an adiabatic colloidal three-phase component system consisting of water vapor, ice particles and liquid droplets may exist for a limited period of time: mixed phase. For quiescent conditions, eventually all the liquid droplets will evaporate and the system will consist of only ice particles and water vapor. The glaciation process due to ice growth by deposition at the expense of co-existing liquid droplets is known as the Wegener-Bergeron-Findeisen (hereafter WBF) mechanism (Wegener 1911; Bergeron 1935; Findeisen 1938).

Previous theoretical studies suggested that the glaciation of mixed phase clouds with ice particle concentrations N_i~10^2-$10^3 l^{-1}$ and liquid water content below $0.5 g/m^3$ should occur within 20-40 minutes. However in-situ observations have shown that mixed phase clouds are a common phenomenon. Rauber and Tokay (1991) and Pinto (1998) have described long-lived narrow layers of supercooled water overlying mixed and ice layers with cloud tops as cold as -30oC. The existence of such layers appear to conflict with the outcome expected from the WBF mechanism. Rauber and Tokay (1991), Pinto (1998) and Harrington (1999) attempted to explain the existence of such layers by an imbalance between the condensate supply rate, the bulk ice crystal mass growth rate and removal of ice freezing nuclei (IFN) by precipitating ice particles. Korolev and Isaac (2003) found that a cloud parcel undergoing vertical oscillations may be subject to an indefinitely long periodic evaporation and activation of liquid droplets in the presence of ice particles. After a certain time, the average IWC and LWC reaches a steady state. This phenomenon may explain the existence of long-lived mixed phase stratiform layers.

The purpose of this paper is to study the conditions required for activation of liquid water in ice clouds and maintaining it in mixed phase at the condensational stage.

* *Corresponding author address*: Alexei Korolev, Environment Canada 4905 Dufferin Street, Toronto, Ontario, M3H5T4 Canada, alexei.korolev@ec.gc.ca

2. NECESSARY AND SUFFICIENT CONDITIONS FOR ACTIVATION OF LIQUID IN ICE CLOUD

Korolev and Mazin (2003) found a threshold vertical velocity

$$u_z^* = \frac{b_i^* N_i \bar{r}_i}{a_0} \tag{1}$$

which is required for the activation of liquid water in ice clouds, if

$$u_z > u_z^* . \tag{2}$$

here, N_i, r_i are the concentration and average size of ice particles; a_0, b_i^* are coefficients dependent upon T (temperature) and P (pressure).

Fig. 1 shows the dependence of the threshold velocity upon ice particle integral radius $N_i\bar{r}_i$. As seen in Fig.1 for typical values of $N_i\bar{r}_i$ (gray area) the threshold velocity u_z^* changes from a few centimeters to a few meters per second. Based on in-situ observations, such vertical velocities are common in clouds and they can be generated by turbulence or regular motions (e.g. convection or gravity waves). Therefore, the diagram in Fig.1 suggests that mixed phase clouds can be frequently formed in the atmosphere due to the activation of liquid water within ice clouds.

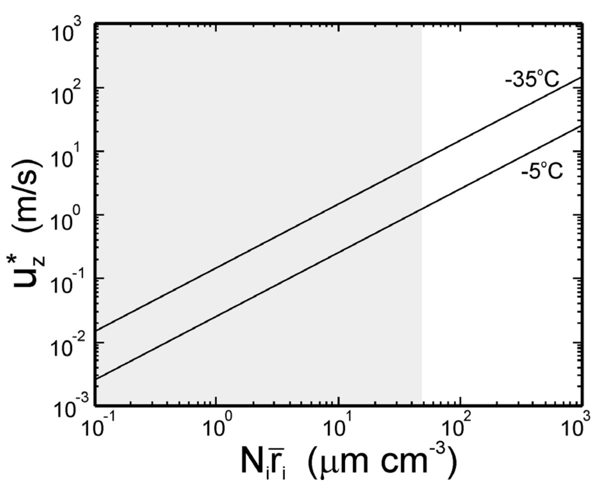

Figure 1. Threshold velocity of ascent required for activation of liquid in ice cloud versus integral radii of ice particles.

Eq.2 provides necessary but not sufficient conditions for the activation of liquid phase within ice cloud. For example, the condition given in Eq.2 may be satisfied when the water vapor pressure is lower than its saturation value over water. In this case, the activation of liquid water will not occur and the cloud will remain in the ice phase.

The sufficient condition for the activation of liquid water in ice cloud requires the vapor pressure of a cloud parcel to reach water saturation. In order to examine the processes of activation, condensation and evaporation of liquid droplets in presence of ice particles, we consider the changes of cloud water content in the *W-Z* (condensed water - height) coordinates (Fig.2).

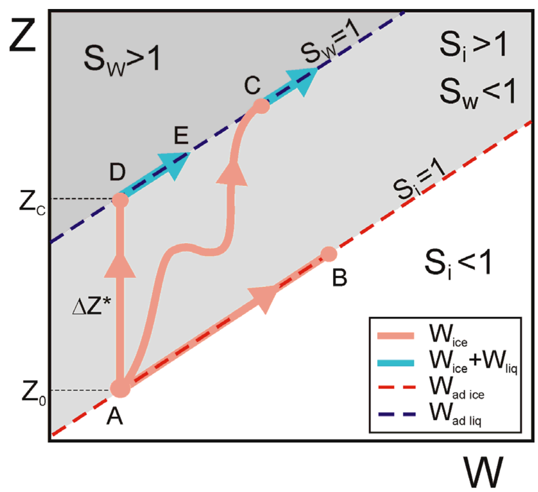

Figure 2. Conceptual diagram of the activation of liquid phase in an ascending parcel within an ice cloud. Activation of liquid phase occurs when a trajectory $W_{ice}(Z(t))$ intersects $W_{ad\ liq}(Z)$ (e.g. points C and D). The area above liquid water adiabat $W_{ad\ liq}(Z)$ is supersaturated with respect to water ($S_w>1$), the area below ice water adiabat $W_{ad\ ice}(Z)$ is subsaturated with respect to ice ($S_i<1$), the area between $W_{ad\ liq}(Z)$ and $W_{ad\ ice}(Z)$ is supersaturated with respect to ice and subsaturated with respect to water ($S_i>1$ and $S_w<1$).

We will assume that the initial relative humidity inside the cloud parcel is equal to ice saturation and that the ice particles are in equilibrium with water vapor. If the ascent velocity of the parcel is infinitesimally small ($u_z \to 0$), then the ice particles have enough time to accommodate water vapor supersaturation caused by adiabatic cooling during ascent, and the ice water content (W_{ice}) will change along an ice adiabat (line AB in Fig. 2). The relative humidity inside this parcel will always be at ice saturation, and it will never reach water saturation. If the velocity of ascent is infinitely large ($u_z \to \infty$), then the vertical gradient of condensed ice will be infinitely small i.e. $dW_{ice}/dz \to 0$ and W_{ice} will be constant. In this case, the relative humidity in the parcel will change as if it is in a cloud free volume

$$\frac{1}{S}\frac{dS}{dz} = a \qquad (3)$$

here, $S=E/E_{sw}$ is the water vapor relative humidity; E is water vapor pressure; $E_{sw}(T)$ is saturated water vapor pressure at temperature T; $a = \frac{gL_w}{c_p R_v T^2} - \frac{g}{R_a T}$; R_a and R_v are specific gas constants of the air and water vapor; L_w is the latent heat of vaporization; c_p is specific heat capacity of the air at constant pressure; g is the acceleration due to gravity.

Integrating Eq.3 yields the vertical distance required for an adiabatic parcel to ascend from the level at ice saturation up to a level where water saturation can be found (line AD in Fig.1)

$$\Delta Z_c = a^{-1} \ln\left(\frac{E_{sw}}{E_{si}}\right) \qquad (4)$$

The value of ΔZ_c gives the shortest vertical distance for a parcel within an ice cloud, initially saturated with respect to ice, to achieve saturation over water. Figure 3 shows the dependence of ΔZ_c on temperature. As seen from Fig.3 ΔZ_c monotonically decreases with an increase of temperature. For T>-20C ΔZ_c does not exceed 300m. The behavior of $\Delta Z_c(T)$ (Fig.3) suggests that the probability of formation of mixed phase clouds at warmer temperatures is higher than at low temperatures, since the occurrence of fluctuations with smaller vertical displacements, ΔZ, and velocity u_z are greater than those with larger vertical displacements.

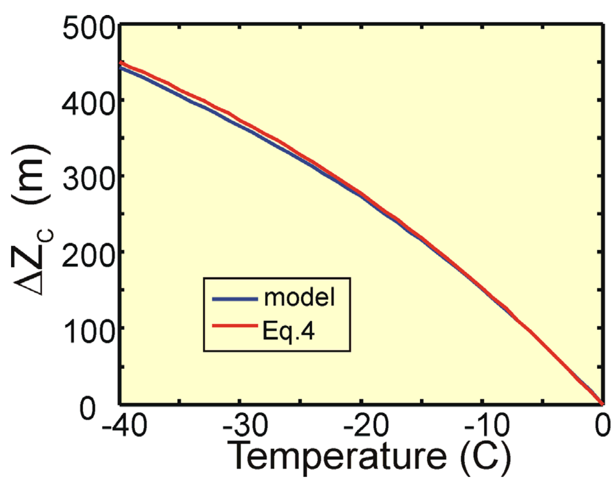

Figure 3. Distance between lifting condensation levels of ice and liquid versus temperature.

If the condition given in Eq.2 is satisfied, then the ascent of a cloud parcel beyond the point D (Fig.2) will result in the condensation of liquid water and the ice cloud will become mixed phase. Assuming that the sink of water vapor to liquid droplets is instantaneous, the total condensed cloud water content in the cloud ascending above the point D (line DE in Fig.2) with the vertical speed $u_z > u_z^*$ will be equal to the adiabatic liquid water ($W_{ad\ liq}$) passing through point D, i.e.

$$W_{ice}(Z) + W_{liq}(Z) = W_{ad\ \ liq}(Z) \tag{5}$$

The adiabatic liquid ($W_{ad\ liq}$) and ice ($W_{ad\ ice}$) water contents are defined by initial ice water content, relative humidity and temperature, and they can be found using

$$W_{ad\ \ liq}(Z) = W_{i0} + \delta W_i + \int_{Z_0+\Delta Z_c}^{Z} \beta_w(z')dz' \tag{6}$$

$$W_{ad\ \ ice}(Z) = W_{i0} + \delta W_i + \int_{Z_{0c}}^{Z} \beta_i(z')dz' \tag{7}$$

where, W_{i0} is the initial ice water content of the parcel and an offset of δW_i is included for generality to account for parcels starting off either sub- or supersaturated with respect to ice. For the example shown in Fig. 2 the initial conditions at point A are such that $\delta W_i=0$;. β_w and β_i are the adiabatic vertical gradients of liquid and ice water content, respectively.

At the beginning of this section we considered two extreme cases of the ice clouds ascending with infinitesimally small and infinitely large vertical velocities. In natural clouds such situations never occur. Due to the limited rate of condensation and evaporation of water vapor by cloud particles, the supersaturation in clouds is a function of the time history of vertical motion (Korolev and Mazin 2003), i.e.

$$S(t) = f(u_z(t)) . \tag{8}$$

It can be shown that the supersaturation and the ice water content are related

$$W_{ice}(t) = W_{i0} + \int_0^t BN_i S(t)\sqrt{r_{i0}^2 + 2A\int_0^t S_i(t')dt'}dt \tag{9}$$

here A and B are coefficients dependent upon T and P. Therefore, after substituting Eq.8 into Eq.9, the time dependent change of ice water content can be expressed as

$$W_{ice}(t) = F(u_z(t)) \tag{10}$$

i.e. $W_{ice}(t)$ can also be considered as a function of vertical velocity history. By substituting t in the left side of Eq.10 with Z found from integrating $Z = Z_0 + \int_0^t u_z(t')dt'$ we obtain

$$W_{ice}(Z(t)) = F(u_z(t)) \tag{11}$$

This formulation of $W_{ice}(t)$ enables one to draw a general conclusion about the existence of an ensemble of trajectories $u_z(t)$ that result in intersecting $W_{ice}(Z(t))$ and $W_{ad\ liq}(Z)$ (Fig.2). Such intersections represent mixed phase cloud and identify the activation of liquid water within ice.

A mathematical formulation of the above statement can be presented in the following way: the activation of liquid phase in ascending adiabatic parcels within ice clouds will occur if, and only if Eqs. 6 and 11 have a solution $W_{ice}(Z) = W_{ad\ \ liq}(Z)$

This statement gives the necessary and sufficient conditions for turning ice clouds into mixed phase. If the function $u_z(t)$ is such that Eqs. 6 and 11 are not satisfied then the cloud remains in the ice phase.

The condition for the activation of liquid water within ice clouds has a simple graphical interpretation: the activation of liquid water occurs in an ascending parcel when the curve $W_{ice}(Z(t))$ intersects $W_{ad\ liq}(Z)$ in W-Z coordinates. If the intersection of $W_{ice}(Z(t))$ and $W_{ad\ liq}(Z)$ occurs during decent, then it will correspond to the evaporation of droplets and glaciation of the parcel.

Figure 4 shows results of numerical modeling of the activation of liquid water within an ice cloud for a parcel ascent with constant velocity u_z=0.5m/s. Since r_{im} is increasing during ascent, then u_z^* is also increasing following Eq.1. The vertical changes of u_z^* is shown in Fig.4b. In accordance with the above discussion the activation of water occurs as soon as $W_{ice}(Z)$ intersects the liquid water adiabat $W_{ad\ liq}(Z)$ at the level indicated by line Z_a (Fig 4a). Above the level Z_a total water content $W_{ice}(Z)$+$W_{liq}(Z)$ (green line) follows the liquid adiabat $W_{ad\ liq}(Z)$ (dased blue line) until it reaches the level at which the droplets have evaporated Z_e. At the level Z_e ice water content $W_{ice}(Z)$ is equal to the liquid adiabat $W_{ad\ liq}(Z)$. Liquid water content $W_{liq}(Z)$ reaches a maximum at the level Z_m where $u_z = u_z^*$ (point B in Fig.4b) and then it starts to evaporate since $u_z < u_z^*$ (Fig.4b). It should be noted that the activation of liquid water does not occur below level Z_a, even though $u_z > u_z^*$ (line OA in Fig.4b), since the water vapor pressure is below water saturation. This example illustrates the above statement that the condition given in Eq.2 is necessary but not sufficient for the activation of liquid water in ice clouds.

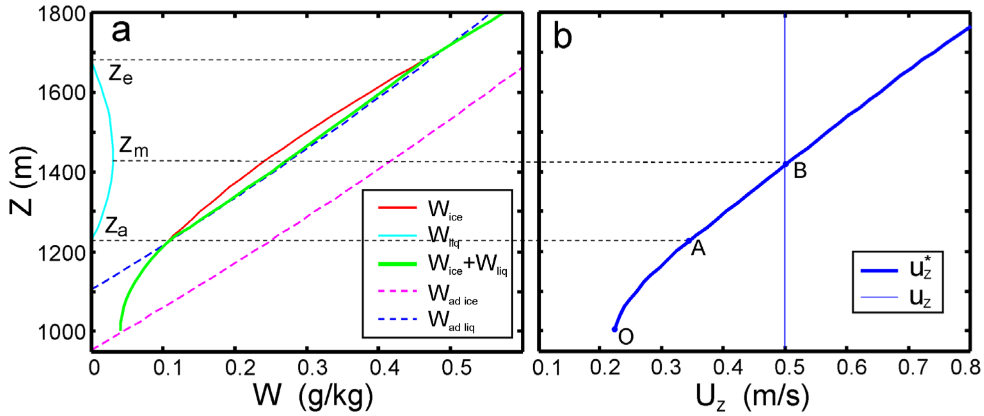

Figure 4. Numerical modeling of activation of liquid water within ice cloud in a uniformly ascending adiabatic parcel with u_z=0.5m/s, N_i=100l^{-1}, r_{i0}=50μm, S_{i0}=1; T=-10C. (a) vertical changes of ice, liquid and total water contents; (b) vertical changes of threshold velocity (Eq.1) and velocity of the parcel.

3. VERTICAL HARMONIC OSCILLATIONS

Eqs. 6 and 11 do not have an analytical solution in a general form for arbitrary $u_z(t)$. Thus, in order to predict formation of mixed phase clouds one has to solve a system of differential equations describing the cloud microphysics. However, the conditions for the activation of liquid water within ice clouds can be found for some special cases. In the following sections we discuss the formation of mixed phase clouds during harmonic oscillations of $u_z(t)$.

3.1 *Ice cloud*

Before proceeding with analysis of activation of liquid water in a vertically oscillating parcel, we will examine some characteristic features of the condensation-evaporation processes in an ice only cloud during harmonic oscillations of the vertical velocity $u_z = u_0 \sin(\omega t)$, where $\omega=2\pi/t_0$, $t_0=2\pi R/u_0$ is the period of oscillations, R is the radius of oscillations, and u_0 is the tangential velocity. No activation of liquid water is allowed during these fluctuations, and the cloud remains in a two-phase condition consisting of ice particles and water vapor.

The supersaturation with respect to ice in the parcel can be described as

$$\frac{1}{S_i}\frac{dS_i}{dt} = \left(\frac{gL_i}{c_pR_vT^2} - \frac{g}{R_aT}\right)u_z - \left(\frac{1}{q_v} + \frac{L_i^2}{c_pR_vT^2}\right)\frac{dW_{ice}}{dt} \tag{12}$$

Analysis of Eq. 10 suggests that the characteristic time of relaxation to a quasi-steady supersaturation in the ascending parcels is defined by the so-called time of phase relaxation (Mazin 1968, Korolev and Mazin 2003)

$$\tau_{ph} \approx \frac{b}{N_i\bar{r}_i} \tag{13}$$

here, b is a coefficient dependent upon P and T.

If the characteristic time of vertical motion

$$\tau_z >> \tau_{ph}\,, \tag{14}$$

then ice particles have enough time to accommodate (release) water vapor during ascents (descents), and bring the two-phase “ice-vapor” system into quasi-equilibrium. It has been shown that the supersaturation $S_i(t)$ in a cloud parcel approaches a quasi-steady value $S_{qsi}(t)$, with time (Korolev and Mazin 2003). For example, typically over $3\tau_{ph}$ the difference between $S_i(t)$ and $S_{qsi}(t)$ becomes smaller than 10%. Since, $\lim_{t\to\infty} S_{qsi} = 0$ (Sedunov, 1974), then $\lim_{t\to\infty} S_i = 0$ as well. Therefore, $\lim_{t\to\infty} W_{ice} = W_{ad\ ice}$, since $S_i \propto (W_{ice} - W_{ad\ ice})$. Thus, for characteristic vertical fluctuations with a time scale $\tau_z >> \tau_{ph}$, (Eq.14) ice water content W_{ice} approaches its adiabatic value.

If Eq.14 is not valid then, W_{ice} may be significantly different from $W_{ad\ ice}$., depending on the ratio of τ_z and τ_{ph}.

For fluctuations with a vertical scale ΔZ and the characteristic velocity $\bar{u}_z$ Eq.14 can be rewritten as

$$\Delta Z >> \Delta Z_{ph} \tag{15}$$

where $\Delta Z=\bar{u}_z\,\tau_z$, and $\Delta Z_{ph} = \bar{u}_z\tau_{ph}$. In the following discussion ΔZ_{ph} will be referred to as the “vertical phase scale” and has a simple physical interpretation: ΔZ_{ph} is a vertical scale such that at $\Delta Z >> \Delta Z_{ph}$, regardless of the trajectory $u_z(t)$, cloud water content will approach its adiabatic value. For turbulent fluctuations an expression similar to Eq.15 was found in general form by Kabanov and Mazin (1970).

Figure 5 illustrates the different behaviors of W_{ice} in four parcels having the same amplitude of vertical oscillations ΔZ_{ph} Z=200m, but different velocities u_0=0.02m/s, 0.2m/s, 1m/s and 5m/s. The initial conditions for all of the parcels were the same: N_{ice}=1000l^{-1}; r_i=20μm; T=-15C, S_{i0}=1.01.

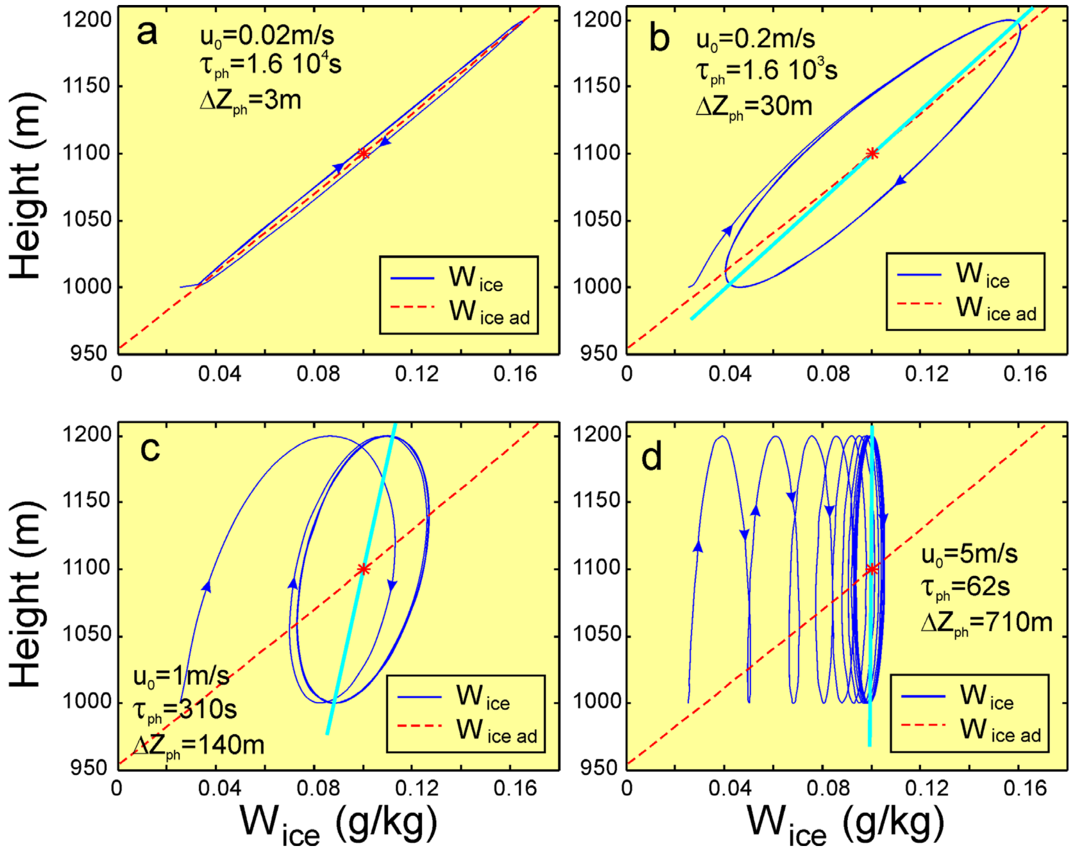

Figure. 5. Changes of ice water content in an adiabatic ice cloud when the vertical velocity varies as $u_z = u_0 \sin(\omega t)$. For all four cases N_i=1000l^{-1}, r_{i0}=20μm, ΔZ_{ph} Z=200m, τ_{ph}≈220s, T_0=-10C, S_{i0}=1.01. (a) ΔZ>>ΔZ_{ph}, τ_z≈1.6 10^4s; (b) ΔZ>ΔZ_{ph}, τ_z≈1.6 10^3s; (c) ΔZ~ΔZ_{ph}, τ_z≈310s; (d) ΔZ<ΔZ_{ph}, τ_z≈62s,

As seen in Fig.5 W_{ice} approaches a limit-cycle having a quasi-elliptical shape. The minor axis of the limit-cycle ellipse increases and then decreases with the increase of u_0, whereas the major axis monotonically decreases with the increase of u_0. The slope of the major axis of the limit-cycle $\chi = \left(dW_{ice}/dZ\right)_e$ changes from zero to β_i as u_0 increases from 0 to ∞.

The center of the limit-cycle is located near the intersection of the $W_{ad\ ice}(Z)$ and the axis of rotation $Z=Z_0+\Delta Z/2$. For fluctuations with ΔZ >500-700m the center of limit-cycle deviates to the left side from $W_{ad\ ice}(Z)$ due to its non-linear dependence $W_{ad\ ice}$ versus Z. Since ΔZ and initial conditions are the same, the locations of the centers of limit-cycles for all four cases shown in Fig.5 are also the same. Therefore, the average size of ice particles for the limit-cycle can be estimated as

$$\bar{r}_{im} = \left(\frac{3W_{im}}{4\pi\rho_i N_i}\right)^{1/3} \tag{16}$$

where W_{im} is the adiabatic ice water content in the center of limit-cycle. For ΔZ<500-700m it can be assumed that β_i is a constant, and, therefore, W_{im} can be estimated from Eq.5 as

$$W_m = W_{i0} + \delta W - \frac{1}{2}\beta_i \Delta Z \tag{17}$$

Since the centers of the limit-cycles are the same, the average sizes of the ice particles are also the same for all four cases. Substituting N_{i0}, r_{i0}, S_{i0} into Eqs. 16, 17 and 12 yields $\bar{r}_{im}$ =32μm and τ_{ph}≈220s.

The characteristic vertical velocity along ascending or descending branches can be estimated as $\bar{u}_z = \frac{2u_0}{\pi}$. Hence, the characteristic time ($\tau_z = \Delta Z/\bar{u}_z$) and vertical phase scales ($\Delta Z_{ph} = \bar{u}_z \tau_{ph}$) of the cases shown in Fig, 5a, b, c, and d will be, respectively: τ_z≈1.6 10^4s, 1.6 10^3s, 310s, 62s and ΔZ_{ph} ≈3m, 30m, 140, 710m.

As seen in Fig.5, the behavior of $W_{ice}(Z)$ during harmonic fluctuations is in agreement with the above discussion. If $\Delta Z >> \Delta Z_{ph}$ (or $\tau_z >> \tau_{ph}$) then $W_{ice}(Z)$ is close to the adiabatic ice water content $W_{ad\ ice}(Z)$. This case corresponds to that shown in Fig. 5a. If $\Delta Z << \Delta Z_{ph}$ (or $\tau_z << \tau_{ph}$), then the fluctuations take place as if in cloud free air. W_{ice} and r_i during such fluctuations remains approximately constant (Fig.5d). The cases shown in Fig.5 b and c are intermediate cases between $\Delta Z >> \Delta Z_{ph}$ and $\Delta Z << \Delta Z_{ph}$.

3.2 *Activation of liquid phase during vertical harmonic oscillations*

Figure 6 shows a conceptual diagram of the activation of liquid water in ice clouds for two limit-cycles of $W_{ice}(Z)$ formed during harmonic oscillations of u_z. The limit-cycle (1) has the center of rotation at point O_1 and the amplitude of the vertical oscillations is $\Delta Z = 2\Delta Z_c$. When the parcel reaches point A_1 the vapor pressure reaches water saturation. However, the activation of liquid water will not occur since at point A_1 the vertical component of the velocity $u_z = u_0 \sin(\pi) = 0$. The limit-cycle (2) rotates around point O_2, and its vertical amplitude is $\Delta Z > 2\Delta Z_c$. The activation of liquid water occurs at point A_2 and the layer Δh=A_2B_2 is mixed phase (Fig.6). As seen from Fig.6 the depth of the mixed phase layer is

$$\Delta h = \frac{\Delta Z}{2} - \Delta Z_c \quad (18)$$

For the harmonic oscillations the vertical velocity at point A_2 can be calculated as

$$u_z = u_0\sqrt{1 - 4\frac{\Delta h^2}{\Delta Z^2}} \quad (19)$$

Based on the above consideration it can be concluded that for activation of liquid water in ice cloud during harmonic fluctuations it is necessary and sufficient to satisfy two following conditions

$$u_0 > U_z^* \quad (20)$$

$$\Delta Z > \Delta Z^* \quad (21)$$

here

$$U_z^* = u_z^*\left(1 - 4\frac{\Delta h^2}{\Delta Z^2}\right)^{-1/2} = \frac{\Delta Z u_z^*}{2\sqrt{(\Delta Z - \Delta Z_c)\Delta Z_c}} \quad (22)$$

$$u_z^* = \frac{b_i(E_{sw} - E_{si})}{a_i E_{si}} N_i \bar{r}_{im} \quad (23)$$

here a_i and b_i are coefficients dependent on T and P.

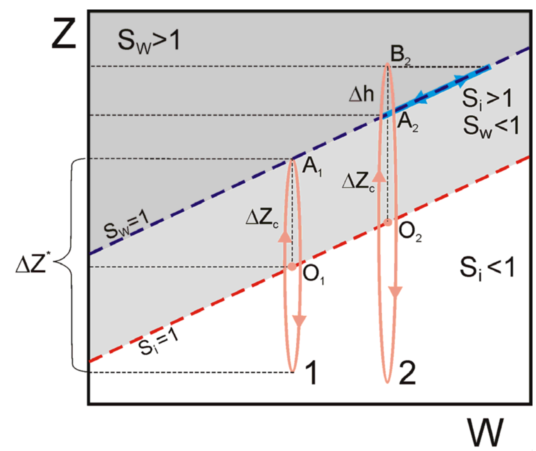

Figure 6. Conceptual diagram of the activation of liquid water during vertical fluctuations

Figure 7 shows some results of numerical modeling of the activation of liquid water during harmonic oscillations. The purpose of the cases shown in Fig.7 is to demonstrate the necessary and sufficient conditions for activation of liquid water within ice clouds. Figure 7a shows the case when the velocity condition is satisfied, but the vertical displacement is insufficient: $u_0 > U_z^*$, and $\Delta Z < \Delta Z_c$. For this case the activation of liquid water is not possible for any u_z. For the case shown in Fig.7b $\Delta Z > 2\Delta Z_c$, with $u_0 < U_z^*$, there is again no activation of liquid water, because this time the vertical velocity condition is not satisfied. For the case shown in Fig.7c where $u_0 > U_z^*$ and $\Delta Z_c < \Delta Z < 2\Delta Z_c$, activation of water occurs only during first few cycles, but as $W_{ice}(Z)$ approaches the limit-cycle $\Delta Z < 2\Delta Z_c$, and the activation of water is discontinued. For the case in Fig.7d where both conditions $u_0 > U_z^*$ and $\Delta Z > \Delta Z^*$ are satisfied, the activation and evaporation of liquid water occurs indefinitely.

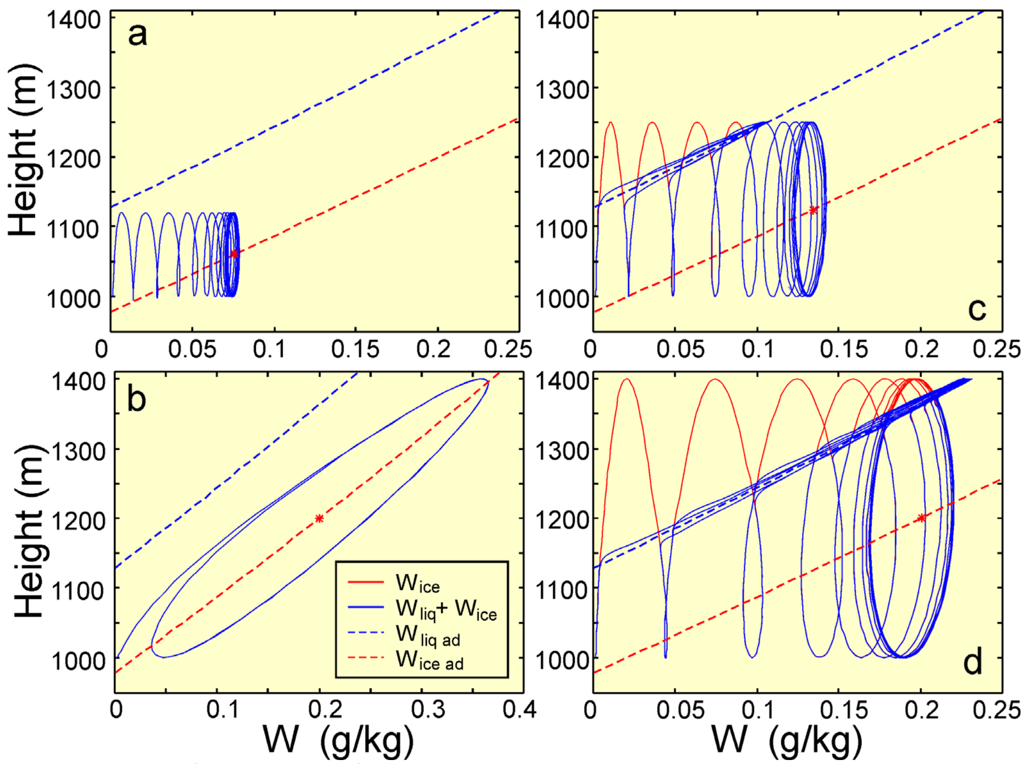

Figure 7 Numerical modeling of the activation of liquid water in ice cloud during vertical harmonic oscillations. For all four cases N_i=50l^{-1}, r_{i0}=20μm, S_{i0}=1.01, T_0=-10C, ΔZ=153m. (a) ΔZ=125m, u_0=0.5m/s, u_z^*=0.08m/s; (b) ΔZ=250m, u_0=1m/s, u_z^*=0.10m/s; (c) ΔZ=400m, u_0=0.05m/s, u_z^*=0.12m/s; (d) ΔZ=400m, u_0=1m/s; u_z^*=0.12m/s

Figure 8 shows a comparison of the threshold velocity calculated from Eqs.22,23 with that obtained from numerical simulations of the activation of liquid water within ice clouds. The modeling was conducted for different amplitudes of fluctuations 340m<ΔZ<800m and velocities (0.05m/s<u_0<6m/s). The concentration of ice particles in different runs varied from 50l^{-1} to 5000l^{-1}. As seen from Fig.8 all points are aligned, but deviate from the 1:1 line by approximately 20%. The diagrams in Fig.8 support the theoretical framework developed in the present study.

4. CONCLUSIONS

The following conclusions have been obtained:

(a) During vertical harmonic oscillations of an adiabatic two-phase, ice-vapor, colloidal system, the microphysical parameters such as particle size, condensed water, supersaturation, temperature, etc. approached a limit-cycle.

(b) Necessary and sufficient conditions for activation of liquid water in ice phase cloud during vertical fluctuations have been found. These conditions can be used for the parameterization and forecast of mixed phase clouds in numerical models.

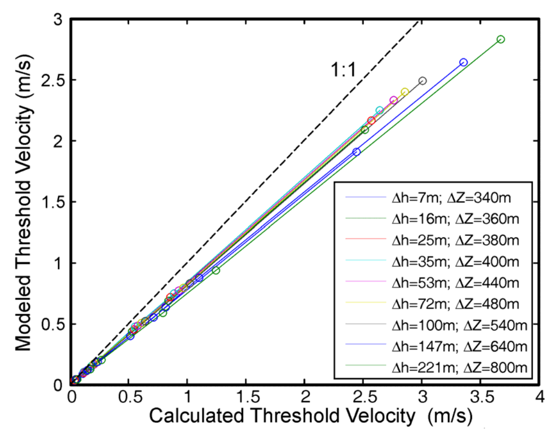

Figure 8. Comparison of the theoretical threshold velocity U_z^* (Eqs. 22, 23) and deduced from numerical modeling of activation of liquid water in liquid clouds. T_0=-10C, 340m<ΔZ<800m; 0.05m/s<u_0<6m/s; 50l^{-1} <N_i< 5000l^{-1}.

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