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Generation of Available Buoyant Energy by Cloud Glaciation

By Charles F. Chappell[1]) and David R. Smith[1])

Abstract

The available buoyant energy (ABE, energy from the environment which becomes available to a parcel for buoyant accelerations) arising from glaciation is computed by integrating upward the differences in temperature between a parcel that undergoes instantaneous and isenthalpic freezing followed by an ice-saturation ascent, and one that experiences only a water-saturation ascent from the same initial cloud base conditions. This quantity is computed for three initial cloud base conditions representative of tropical, High Plains summertime, and Great Lakes wintertime cumuli. Substantial increases in parcel updraft speed are realized for all clouds if the ABE arising from glaciation is completely converted to parcel kinetic energy. Variations of the three components of parcel heating involved in the glaciation process (i.e., (1) release of latent heat of fusion from freezing of liquid water, (2) cooling or warming from sublimation or deposition as vapor pressure adjusts from water saturation to ice saturation at the post glaciation temperature, and (3) the additional warming or cooling relative to the initial water-saturation adiabat as the parcel follows an ice-saturation ascent to a specified upper reference level) are also determined as functions of glaciation temperature. It is found that sublimation substantially counteracts the parcel warming arising from the freezing of liquid water in the case of warm moist cumuli. In addition, it is found that in some instances ice-saturation ascent following glaciation can produce cooling relative to the initial departure from the water saturation adiabat. This was indicated for Great Lakes wintertime cumuli and also for warm moist cumuli with glaciations at very cold temperatures. The effect upon the buoyancy force, of the change in the mass of condensate during glaciation, is small and can usually be neglected.

1. *Introduction*

The importance of heat released by cloud seeding was first discussed by Kraus and Squires (1947). In the same year, Langmuir *et al.* (1947) pointed to the availability of another heat source during cloud seeding besides the latent heat of fusion. Langmuir said, 'when supercooled liquid water droplets are made to evaporate and condense on ice nuclei, the amount of ice formed is considerably greater than the amount of water which evaporates because the vapor pressure is lower than the water so that there is a lowering of the water vapor content in the cloud. There are thus, two sources of heat which tend to raise the temperature of the cloud, viz., the heat of fusion and the heat of sublimation of the extra amount of water which is converted from vapor to ice.' However, Macready and Skutt (1967) demonstrated that this latter effect discussed by Langmuir could act to either increase or decrease the parcel temperature,

[1]) Atmospheric Physics and Chemistry Laboratory, National Oceanic and Atmospheric Administration, Boulder, Colorado 80302, USA.

since both cooling from sublimation or heating by deposition was possible. In addition, they presented a graphical technique for estimating the increase in cloud buoyancy from seeding. FUKUTA (1972) and ORVILLE and HUBBARD (1973) also noted that cooling from sublimation may in reality counteract the warming realized from freezing liquid water and reduce cloud buoyancy.

SAUNDERS (1957) provided a basis for considering the dynamic effects of introducing ice in cloud parcels. He presented equations that describe both reversible and pseudoadiabatic expansions of air saturated with respect to plane surfaces of water and ice. He also presented an expression for evaluating the temperature change of an instantaneous and isobaric freezing process. Since this freezing technique can also be applied to freezing over a temperature interval, it may be conveniently used in cloud models.

ORVILLE and HUBBARD (1973) emphasized the importance of treating the freezing process accurately in cloud models, particularly when simulating the effects of 'dynamic' seeding (WOODLEY, 1970; SIMPSON and WIGGERT, 1971). Furthermore, they found that the imprecise manner in which the freezing process was incorporated into some cumulus models led to a bias against the natural cloud and overestimates of their seedability.

The process of glaciation produces a total parcel heating that is composed of three components (ORVILLE and HUBBARD, 1973). The first two occur at the time of freezing and are due to the release of latent heat of fusion as the supercooled liquid water freezes, accompanied by either cooling from sublimation or heating from deposition as the vapor mixing ratio adjusts to ice-saturation at the new post-freezing temperature. The third component is realized when the parcel ascends to higher levels in the atmosphere and cools at a rate specified by an ice-saturation expansion.

The purpose of this paper is to explore the effects of cloud glaciation on the available buoyant energy (ABE, energy from the environment which becomes available to a parcel for buoyant accelerations). Variations of the three thermal components with glaciation temperature are also investigated for three different initial conditions. These initial conditions are chosen to represent typical cloud bases for tropical cumuli, High Plains summertime cumuli, and Great Lakes wintertime cumuli. Effects of the thermal components, which arise from glaciation and subsequent ice-saturation ascent, are integrated upward to an upper reference level to estimate their contribution to the production of ABE. Finally, the depositional growth or sublimation that accompanies the adjustment of saturation vapor pressure during glaciation is investigated, and its importance relative to thermal effects is defined.

2. *Equations and computational methods*

(*a*) *Computation of ABE*

We wish to compute the amount of ABE below some upper reference level that arises from the glaciation process, and determine quantitatively how changes in the glaciation level affect this quantity. This is accomplished by assuming that a cloud parcel follows a water-saturation (reversible adiabatic or pseudoadiabatic) ascent from an arbitrarily specified cloud base up to the reference level. Temperatures along this thermodynamic path are designated T_1. We now consider that a cloud parcel with identical cloud base characteristics experiences isobaric and instantaneous freezing at some arbitrarily specified pressure level, after which it follows an ice-saturation ascent to the reference level. Temperatures describing this second thermodynamic path are identified T_2. These two thermodynamic processes are illustrated in Fig. 1.

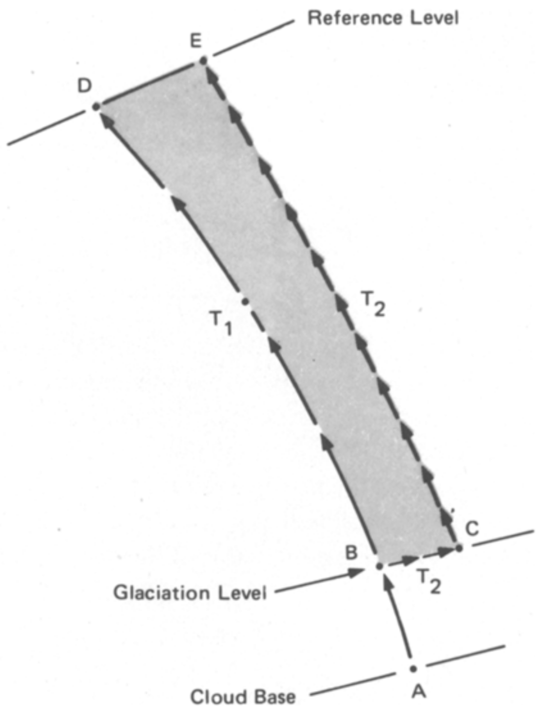

Figure 1
Diagram of relevant thermodynamic processes. ABD represents a water-saturation expansion (pseudoadiabatic or reversible). BC depicts instantaneous and isobaric freezing. CE represents an ice-saturation expansion (pseudoadiabatic or reversible).

If the temperature difference between these two thermodynamic paths is integrated from the glaciation level to the reference level, the amount of ABE below the reference level, arising from glaciation and ice-saturation ascent, is given by

$$\mathrm{ABE} = R_d \int_{p_g}^{p_r} (T_2 - T_1) d(\ln p), \tag{1}$$

where symbols are given in the appendix.

An upper reference level at 200 mb is used to integrate (1) for the tropical and High Plains summertime cumuli, while 400 mb is employed for the Great Lakes wintertime cumulus.

Solutions to (1) require that water-saturation and ice-saturation equations (reversible adiabatic and pseudoadiabatic) be computed, as well as the equation that describes instantaneous and isobaric freezing.

(b) Parcel warming during glaciation

In an isenthalpic freezing process total enthalpy of the system is conserved and the temperature T', resulting from the freezing of a supercooled cloud of droplets initially at a temperature T, is given by

$$[C_{pd} + C_{pv}q_i(T') + C_i q_{ci}](T' - T) = q_{cw}L_f + [q_w(T) - q_i(T')]L_s. \tag{2}$$

Equation 2 can be written in terms of the total mixing ratio of water substance, q, (SAUNDERS, 1957) by noting that $q = q_i(T') + q_{ci}$ and $C_{pv} \approx C_i$. The latter relationship is exact at about -34C using the expressions for C_{pv} and C_i defined here. If we take $C_{pv} \approx C_i$ for all temperatures of interest, (2) takes the form

$$(C_{pd} + C_{pv}q)(T' - T) = q_{cw}L_f + [q_w(T) - q_i(T')]L_s. \tag{3}$$

This form of the freezing equation is amenable to physical interpretation but is not convenient computationally, since it requires an iterative process for solution. A more computationally convenient form of (3) can be formulated by assuming that $(T' - T)$ is small enough to be treated as a differential, and then equating $e_i(T')$ to a Taylor series expansion in $e_i(T)$. By employing the first two terms of this expansion, the Clausius–Clapeyron relation, and the assumption that $p - e_i(T) \approx p - e_i(T')$, the temperature change due to freezing can be expressed in terms of the temperature prior to freezing and (3) becomes

$$(T' - T) = \frac{q_{cw}L_f + \dfrac{\varepsilon L_s e_w(T)}{p - e_w(T)} - \dfrac{\varepsilon L_s e_i(T)}{p - e_i(T)}}{C_{pd} + qC_{pv} + \dfrac{\varepsilon^2 L_s^2 e_i(T)}{[p - e_i(T)]R_d T^2}}. \tag{4}$$

The procedure adopted here is to calculate T$'$ using (4), and then to insert T' into (3) to compute temperature changes associated with the terms on the right hand side of the equation. Thus, the physically meaningful thermal components, $\Delta T_f = q_{cw}L_f/(C_{pd} + qC_{pv})$ (temperature change due to freezing of liquid water) and $\Delta T_d = [q_w(T) - q_i(T')]L_s/(C_{pd} + qC_{pv})$ (temperature change due to deposition or sublimation) are evaluated directly.

The third thermal component, ΔT_l, which is realized when the parcel is lifted from the glaciation level, is computed at the upper reference level for a given set of initial conditions by $\Delta T_l = (T_2 - T_1) - (\Delta T_f + \Delta T_d)$, where $(T_2 - T_1)$ is evaluated at the upper reference level, and $(\Delta T_f + \Delta T_d)$ is the net change in parcel temperature realized at the glaciation level.

(*c*) *Condensate effects*

During the glaciation processs depositional growth or sublimation will generally accompany the adjustment of vapor pressure from saturation over a plane water surface to saturation over a plane ice surface at the new post-freezing temperature. This adjustment may produce either an increase or a decrease in the mass of condensate within the updraft parcel. The downward force corresponding to this change in the mass of condensate is then given by $-g[q_w(T) - q_i(T')]$. If this quantity is equated to the buoyancy force, $g\Delta T/T$, the change in the mass of condensate can be transformed to an equivalent temperature difference at the glaciation level, or $\Delta T = T[q_w(T) - q_i(T')]$. This equivalent temperature difference can then be divided by the parcel warming realized during glaciation $(\Delta T_f + \Delta T_d)$ to define its relative importance to buoyancy production at the glaciation level.

(*d*) *Reversible adiabatic processes*

Changes in temperature of a cloud parcel as a function of pressure during a reversible water-saturation adiabatic ascent are given by

$$\left.\frac{dT}{dp}\right)_w = \frac{\left(\dfrac{R_dT}{C_{pd}p}\right)\left(1 + \dfrac{q_w}{\varepsilon}\right)\left(1 + \dfrac{q_wL_c}{R_dT}\right)}{\left[1 + \dfrac{C_wq_{cw}}{C_{pd}} + \dfrac{C_{pv}q_w}{C_{pd}} + \dfrac{L_c^2q_w(q_w + \varepsilon)}{C_{pd}R_dT^2}\right]}. \tag{5}$$

A similar expression for the reversible ice-saturation adiabatic ascent is given by

$$\left.\frac{dT}{dP}\right)_i = \frac{\left(\dfrac{R_dT}{C_{pd}p}\right)\left(1 + \dfrac{q_i}{\varepsilon}\right)\left(1 + \dfrac{q_iL_s}{R_dT}\right)}{\left[1 + \dfrac{C_iq_{ci}}{C_{pd}} + \dfrac{C_{pv}q_i}{C_{pd}} + \dfrac{L_s^2q_i(q_i + \varepsilon)}{C_{pd}R_dT^2}\right]}. \tag{6}$$

(*e*) *Pseudoadiabatic processes*

Changes in temperature of a cloud parcel as a function of pressure during a pseudoadiabatic water-saturation ascent are calculated by

$$\left.\frac{dT}{dp}\right)_w = \frac{\left(\dfrac{R_dT}{C_{pd}p}\right)\left(1 + \dfrac{q_w}{\varepsilon}\right)\left(1 + \dfrac{q_wL_c}{R_dT}\right)}{\left[1 + \dfrac{q_w}{C_{pd}}\left(C_{pv} - C_w + \dfrac{L_c}{T}\right) + \dfrac{L_c^2q_w(q_w + \varepsilon)}{C_{pd}R_dT^2}\right]}. \tag{7}$$

The corresponding expression for a pseudoadiabatic ice-saturation ascent is given by

$$\left.\frac{dT}{dp}\right)_i = \frac{\left(\dfrac{R_dT}{C_{pd}p}\right)\left(1 + \dfrac{q_i}{\varepsilon}\right)\left(1 + \dfrac{q_iL_s}{R_dT}\right)}{\left[1 + \dfrac{q_iL_s}{C_{pd}T} + \dfrac{L_s^2q_i(q_i + \varepsilon)}{C_{pd}R_dT^2}\right]}. \tag{8}$$

In (7), terms that arise from including the variation of L_c with temperature have been retained while in (8) the very small variation of L_s with temperature has been neglected.

The assumption that total pressure is equal to the partial pressure of dry air is not made in (7) or (8).

(*f*) *Other computational procedures*

Values of the latent heats of condensation and fusion in cal gm^{-1} are given by

$$L_c = 597.3 - 0.566(T - 273.16)$$

and

$$L_f = 80.2 + 0.566(T - 273.16).$$

The specific heat of ice in cal gm^{-1} K^{-1} is calculated by

$$C_i = 0.503 + 0.0018(T - 273.16),$$

while values of the specific heat of water in cal gm^{-1} K^{-1} are given by

$$C_w = 1.00 \text{ for } T \geq 288.16\text{K},$$

and

$$C_w = 9.2709297 - 0.060518327T + 0.00011052521T^2 \text{ for } T < 288.16K.$$

Saturation vapor pressure (in mbs) over plane water and ice surfaces are computed with TETENS' (1930) expression or

$$\ln e_s = \ln 6.1078 + [a(T - 273.16)/(T - 273.16 + b)] \ln 10,$$

where $a = 7.5$ and $b = 237.3$ for saturation over a plane water surface, and $a = 9.5$ and $b = 265.5$ for saturation over a plane ice surface.

The amount of liquid condensate, q_{cw}, is taken to be the adiabatic water content (reversible or pseudo) which develops in the parcel as it rises from cloud base to the glaciation level following a water-saturation expansion, or

$$q_{cw} = q_w(T_{cb}) - q_w(T_1).$$

The amount of ice condensate, q_{ci}, above the glaciation level, is given by a similar expression, or

$$q_{ci} = q_w(T_{cb}) - q_i(T_2).$$

3. *Results*

The three thermal components that make up the total parcel heating are depicted in Fig. 2 as a function of glaciation temperature for initial conditions representative of a tropical cumulus cloud base. The large amount of liquid water generated under these conditions leads to large positive values of ΔT_f due to release of latent heat of fusion. However, this warming is not fully realized by the parcel during glaciation since cooling by sublimation ($\Delta T_d < 0$) substantially reduces the net warming. Note this cooling by sublimation exists at all glaciation temperatures for these initial conditions. It is interesting to note that warming by release of heat of fusion is relatively constant with glaciation temperature, reaching a slight maximum for temperatures around -20C to -25C. The slight drop off at colder glaciation temperatures is due to the decrease in the latent heat of fusion, L_f, while little additional liquid condensate becomes available to the cloud parcel at these colder temperatures.

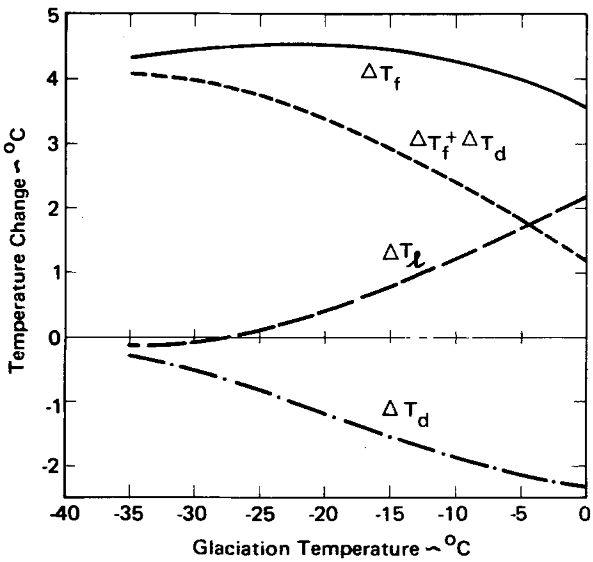

Figure 2
Individual temperature changes associated with cloud glaciation as a function of glaciation temperature for a tropical-cumulus cloud base (p = 900 mb and T = 22C). ΔT_f is the temperature change due to freezing of liquid water. ΔT_d is the temperature change due to sublimation or deposition as vapor pressure adjusts from saturation over a plane water surface to saturation over a plane ice surface at the new post-glaciation temperature. ΔT_l is an additional departure of temperature from the initial water-saturation adiabat that is realized as the cloud parcel ascends from the glaciation level following an ice-saturation adiabat to an upper reference level of 200 mb.

Note also in Fig. 2 that lifting from the glaciation level along a reversible ice-saturation adiabat results in a parcel temperature at 200 mb that may be warmer or cooler relative to its initial departure from the reversible water-saturation adiabat. At warmer glaciation temperatures, however, additional relative warming is realized through lifting.

Variations of the three thermal components with glaciation temperature for initial conditions representative of a High Plains cumulus cloud base are shown in Fig. 3. These curves are similar to those for the tropical cumulus case except that the magnitudes of the temperature changes are generally reduced for all thermal components.

In the case of initial conditions representative of Great Lakes wintertime cumulus clouds, the variations of the three thermal components with glaciation temperature show some important differences. Note in Fig. 4 that warming due to deposition now exists for all glaciation temperatures, and is comparable to the warming associated with the release of heat of fusion, especially at warmer glaciation temperatures. Lifting of the cloud parcel along a reversible ice saturation adiabat now results in a relative cooling (again relative to the initial reversible water-saturation adiabat) at 400 mb for most glaciation temperatures.

It is apparent in Figs. 2–4 that the thermal components associated with the sublimation and lifting processes are nearly equal, but opposite in sense, over a large range

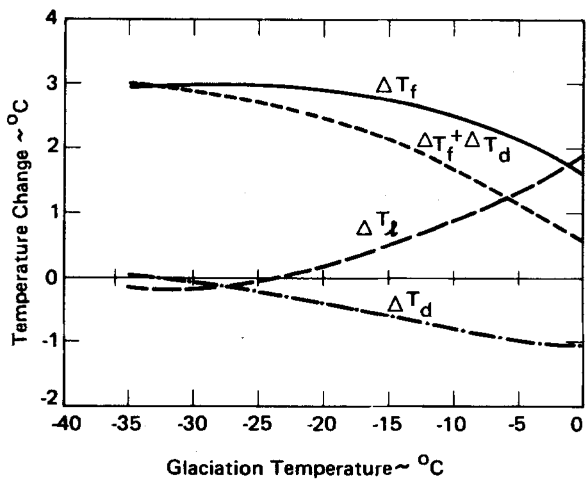

Figure 3
Same as Fig. 2 except for a high plains cumulus cloud base (p = 650 and T = 11C).

of glaciation temperatures for all clouds. Moreover, if the ΔT_l curve is added to the $\Delta T_f + \Delta T_d$ curve in Figs. 2–4, the parcel warming realized at the upper reference level, $\Delta T_f + \Delta T_d + \Delta T_l$, is nearly constant with glaciation temperature. Thus, as pointed out by SAUNDERS (1957), and discussed further by ORVILLE and HUBBARD (1973), the final parcel temperature obtained is largely independent of the temperature at which glaciation occurs.

The effects of glaciation temperature upon the ABE and the resultant theoretical updraft speed at the upper reference level are depicted in Fig. 5. Data are displayed for both pseudoadiabatic and reversible adiabatic processes, and for initial conditions representative of tropical, High Plains, and Great Lakes wintertime cumulus cloud bases. Glaciation produces greater amounts of ABE and larger updraft speeds in the tropical cumuli. For these cloud conditions glaciation at -5C or warmer produces theoretical increases in updraft speed of over 40 mps for both pseudoadiabatic and

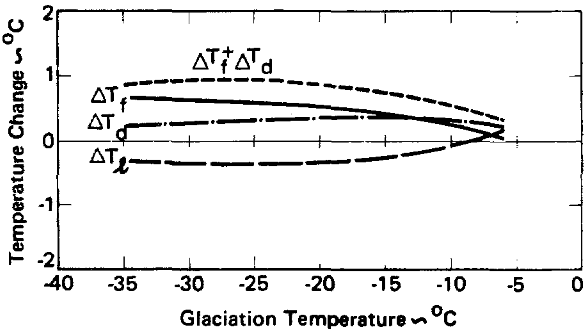

Figure 4
Same as Fig. 2 except for a Great Lakes wintertime cumulus cloud base (p = 900 and T = −5C) and an upper reference level of 400 mb.

reversible adiabatic processes. These values decrease to around 20 mps for glaciation at −35C. The ABE and updraft speed at 200 mb generated in the High Plains cumulus is less than for the tropical cumulus for all glaciation temperatures. Values of updraft speed generated at 200 mb vary from about 35 mps for glaciation at −5C to near 18 mps for glaciation at −35C.

In the case of the Great Lakes wintertime cumulus, the ABE and the resultant theoretical updraft speeds generated at 400 mb are considerably smaller, and exhibit only a slight increase with glaciation temperature. This reflects the small amount of liquid condensate available in these relatively cold clouds, and also the fact that the integration is terminated at 400 mb.

Note in Fig. 5 that values of ABE and theoretical updraft speeds at the upper reference level are greater when computed using the pseudoadiabatic equation. This reflects the greater warming of the cloud parcel by latent heat release, since none of the latent heat is expended to warm condensate in the pseudoadiabatic system.

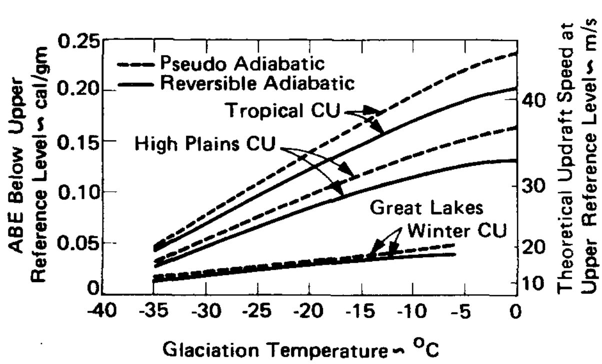

Figure 5
ABE and theoretical updraft speeds realized at the upper reference level due to glaciation as a function of glaciation temperature.

The effect of deposition or sublimation, when saturation vapor pressure adjusts during the glaciation process, has not been included in the results presented so far. However, the change in the mass of condensate arising from this adjustment will have some effect on cloud buoyancy, and therefore on the generation of updraft speed. The relative importance of this effect at the glaciation level is depicted in Fig. 6. Note that the effect on the buoyancy force at the glaciation level, of changing the mass of condensate by deposition or sublimation during glaciation, amounts to less than 5 percent of the buoyancy force produced by parcel warming for all glaciation temperatures

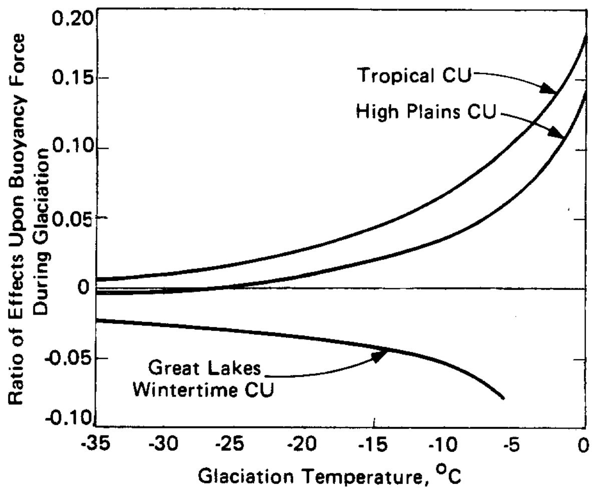

Figure 6

Ratio of the effects upon buoyancy force of condensate change and parcel warming realized during glaciation. The change in condensate during glaciation, due to deposition or sublimation as the vapor pressure adjusts from water saturation to ice saturation at the new glaciation temperature, is converted to an equivalent temperature difference, or $\Delta T = T[q_w(T) - q_i(T')]$. This quantity is then divided by the parcel warming realized during glaciation or $(\Delta T_f + \Delta T_d)$.

colder than −14C. This effect reaches 10 percent to 18 percent at glaciation temperatures warmer than −6C for tropical cumuli. Note that the effect on the buoyancy force is mainly positive (i.e. sublimation reduces the mass of ice condensate) for tropical and High Plains cumuli, but is negative for Great Lakes wintertime cumuli.

4. *Summary and conclusions*

The variations of the three thermal components with glaciation temperature are interesting. While the parcel warming due to release of fusion heat is always positive, the decrease in the latent heat of fusion with temperature more than compensates for smaller increases in liquid water to produce a maximum warming for glaciation temperatures between −20C to −30C for warm moist cumuli. The adjustment of vapor pressure, from saturation over a plane water surface to saturation over a plane ice surface at the new post-glaciation temperature, may produce warming or cooling of the parcel. This effect generally acts to increase the buoyancy of wintertime cumulus clouds and to decrease the buoyancy of warm moist cumuli. Lifting following glaciation does not always result in further relative warming of the cloud parcel.

Relative warming is mainly realized with warm moist cumuli and with glaciations that occur at relatively warm temperatures. The total parcel warming realized from glaciation and subsequent ice-saturation ascent, compared with a water-saturation ascent, amounts to 3C to 4C at 200 mb for warm moist cumuli.

As one might anticipate, in spite of eccentric variations of two thermal components, total parcel heating integrated with height yields an increase in ABE and theoretical updraft speeds at the reference level that increase rather smoothly with glaciation temperature. Also, the warmer the temperature at which glaciation occurs, the greater the ABE and theoretical updraft speed attained at the reference level for all cloud systems. The change in ABE with glaciation temperature decreases above −10C, for warm moist cumuli, so that small differences in results are indicated between glaciations at 0C and −5C. This flattening effect is most pronounced for the reversible adiabatic system.

This study emphasizes the important role that glaciation plays in the dynamics of cumulus clouds. In addition, it demonstrates that substantial amounts of energy are available to be tapped by controlling glaciation temperature through cloud seeding. While we have shown theoretical increases in updraft speed arising from glaciation processes, in reality much of this ABE is likely to be dissipated in entrainment and turbulent processes. Actual updraft speeds attained by cloud parcels are therefore likely to be considerably less than these theoretical values.

Appendix
List of Symbols

C_i	specific heat of ice
C_{pd}	specific heat at constant pressure for dry air [= 0.240 cal gm^{-1} K^{-1}]
C_{pv}	specific heat at constant pressure for water vapor [=0.441 cal gm^{-1} K^{-1}]
C_w	specific heat of water
e_i	saturation vapor pressure over a plane ice surface
e_w	saturation vapor pressure over a plane water surface
g	acceleration due to gravity
L_c	latent heat of condensation
L_f	latent heat of fusion
L_s	latent heat of sublimation [= 677.5 cal gm^{-1}]
p	atmospheric pressure
p_g	pressure at the glaciation level
p_r	pressure at the upper reference level

R_d	specific gas constant for dry air [$= 0.06852$ cal gm^{-1} K^{-1}]
q	mixing ratio of total water substance
q_i	saturation vapor mixing ratio over a plane ice surface
q_w	saturation vapor mixing ratio over a plane water surface
q_{ci}	mixing ratio of ice condensate
q_{cw}	mixing ratio of liquid water condensate
T	temperature
T'	temperature of a cloud parcel after isobaric freezing
T_1	locus of cloud parcel temperatures described by a water saturation ascent from cloud base to an upper reference level
T_2	locus of cloud parcel temperatures described by a water-saturation ascent from cloud base to an arbitrarily specified glaciation level, followed by isobaric freezing and subsequent ice-saturation ascent to an upper reference level
ΔT_d	temperature change during glaciation due to sublimation or deposition as vapor pressure adjusts from saturation over a plane water surface to saturation over a plane ice surface at the new post-glaciation temperature
ΔT_f	temperature change during glaciation due to freezing of liquid water
ΔT_i	additional departure of temperature from the initial water-saturation adiabat that is realized as the cloud parcel ascends after glaciation following an ice-saturation adiabat
ε	the ratio of the molecular weights of water vapor to dry air

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