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On Destabilization of Clouds by Radiative Cooling

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ABSTRACT—Destabilization of clouds due to long-wave radiation is computed as the difference in the radiative cooling rates across the cloud layer. The changes in destabilization and cooling rates due to changes in cloud amount, cloud altitude, cloud thickness, and temperature structure of the lower layer have been assessed separately. Changes in cloud amount cause large changes in the destabilization value and cooling rates near the top and base of the cloud. For a given cloud layer, destabilization is less in low clouds than in middle or high clouds. The presence of an inversion is equivalent to that of a cloud layer and is associated with weak destabilization.

1. INTRODUCTION

Infrared radiative heat exchange processes in an atmosphere containing clouds result in cooling at cloud top and warming at cloud base. This consequently leads to cloud destabilization. The importance of destabilization resulting from outgoing radiation from clouds was stressed earlier by Petterssen (1936). Recent studies by Danard (1969) suggest that the destabilization of clouds may make a significant contribution to the energy cycle of storms or cyclones. Petterssen and Smebye (1971), while discussing the different types of cyclogenetic mechanisms, have documented the energy structures resulting from such cyclogeneses. Their documentation was, however, restricted to kinetic energy only; that of thermal energy, including radiation, was deferred to a later study. Recently, I initiated, in association with S. Petterssen, a study of the role of radiation in the development of extratropical cyclones. Petterssen (1972) has hypothesized that a destabilization process must exist to maintain rather than to initiate the development of extratropical cyclones and to account for their remarkable longevity. As a preliminary step in our investigation, therefore, we have investigated the processes that lead to cloud destabilization and the conditions under which the destabilization is accelerated or declerated. The purpose of this paper is to present the results of this preliminary study.

2. METHOD OF APPROACH

Fluxes of long-wave radiation in the downward ($F_k^{\downarrow}$) and upward ($F_k^{\uparrow}$) directions at any reference level k, may be expressed as (Manabe and Möller 1961)

$$F_k^{\downarrow}=\pi B_c\tilde{\epsilon}_f(y_0, T_c)-\int_{B_0}^{B_c}\tilde{a}_f(y_0, T)\pi dB-\int_{B_k}^{B_0}\tilde{a}_f(y, T)\pi dB \tag{1}$$

and

$$F_k^{\uparrow}=\pi B_G-\int_{B_k}^{B_G}\tilde{a}_f(y, T)\pi dB. \tag{2}$$

The net flux, F_k^N, is given by

$$F_k^N=F_k^{\uparrow}-F_k^{\downarrow} \tag{3}$$

and the rate of heating is given by the flux divergence equation,

$$\frac{\partial T}{\partial t}=\frac{g}{c_p}\frac{\partial F_k^N}{\partial p}. \tag{4}$$

Here, B_0, B_k, and B_G are blackbody fluxes at temperatures at the top of the atmosphere, at level k, and at the earth's surface, respectively. B_c is the blackbody flux at temperature T_c where T_c=220°K (T_c is introduced for the convenience of integration),

$$y_0=|u_k-u_0|,$$
$$y=|u_k-u|,$$
and
$$u_0=0 \tag{5}$$

where u_k is the optical depth measured from the top of the atmosphere (upper boundary) to the reference level k. The terms $\tilde{\epsilon}_f(y, T)$ and $\tilde{a}_f(y, T)$ are the mean slab emissivity and the mean slab absorptivity, respectively, as defined by Yamamoto (1952). The values of $\tilde{\epsilon}_f$ and $\tilde{a}_f$ have been used as given by Manabe and Möller (1961) and Manabe and Strickler (1964).

The fluxes are computed at reference levels that are spaced at intervals of 50 mb from 100 mb (upper boundary of the atmosphere) downward to the ground surface. The effects due to water vapor alone have been considered; those due to carbon dioxide have been found to be small by an order of magnitude, especially in the troposphere (Godbole and Kelkar 1971, Manabe and Strickler 1964). The presence of clouds has been taken into account—for a cloud cover of less than 100 percent a prorated amount of flux is computed from the flux for overcast and for cloudless sky conditions (Godbole 1963). For convenience of computation, the heights of base and top of the cloud are approximated to the nearest reference levels. The radiosonde data for 0000 GMT on Dec. 18, 1967, used in this

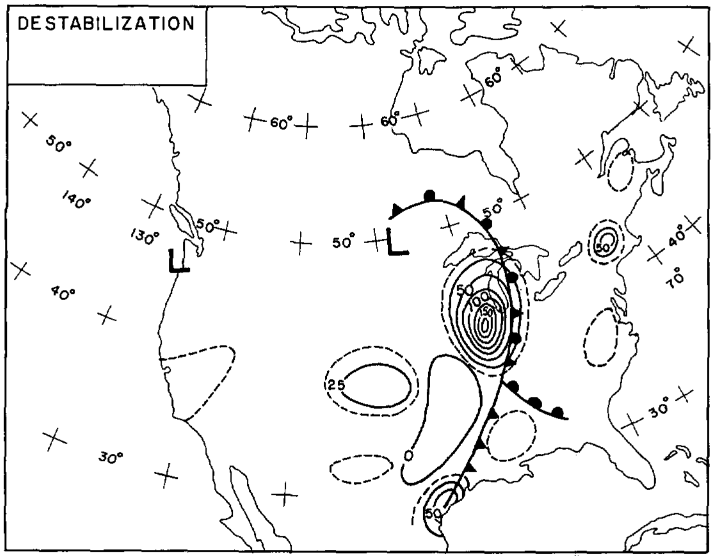

FIGURE 1.—Distribution of cloud destabilization (10^{-2}°C·day^{-1}·mb^{-1}) due to long-wave radiation at 0000 GMT, Dec. 18, 1967.

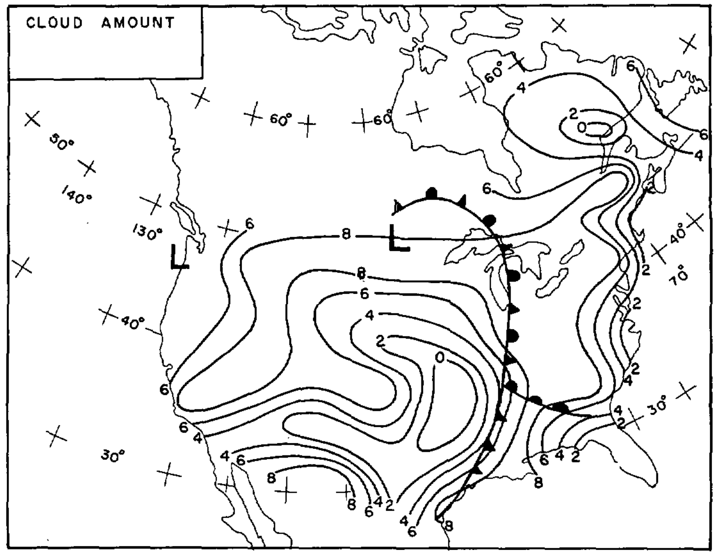

FIGURE 2.—Distribution of cloud cover (okta) at 0000 GMT, Dec. 18, 1967.

study were kindly provided by the National Environmental Satellite Service, NOAA. The above data form a part of the data used in a more broad-based, on-going investigation of the effects of long-wave radiation on a cyclone located west of the Rocky Mountains during the period from 0000 GMT on Dec. 16, 1967, to 1200 GMT on Dec. 18, 1967 (Petterssen and Smebye 1971).

A cloud is considered to act as a blackbody with respect to long-wave radiation. The net flux at a cloud top is the difference between the upward blackbody flux from the cloud top and the downward flux from the atmosphere above. Since the downward radiation originates at lower temperatures and equals only a fraction of the blackbody radiation corresponding to these temperatures, it will be exceeded by the upward flux. In addition, the net flux in the interior of a cloud must be zero; therefore, the top of a cloud loses heat by radiation. At the base of a cloud, the downward flux is exceeded by the flux from below; the base of a cloud is, therefore, heated. The combination of radiational cooling of the top and heating of the base causes the lapse rate in a cloud to become steeper, resulting in destabilization of the cloud layer.

Cloud destabilization, D_s, may be defined as

$$D_s=\frac{\delta}{\delta p}\left(\frac{\partial T}{\partial t}\right) \tag{6}$$

where ∂p denotes the thickness of cloud in millibars. Destabilization results if $D_s>0$. If $D_s<0$, there is no destabilization. Equation (6) may be written as

$$D_s=\frac{\delta}{\delta t}\left(\frac{\partial p}{\partial T}\right). \tag{7}$$

Therefore, since $\partial T/\partial p$ represents the stability parameter, destabilization may also be defined as the rate of change of stability with time. If a cloud is introduced by some means, into an initially stable atmosphere, instability is created across the cloud depth; according to eq (7), this process will lead to destabilization of the cloud mass. We show in the next section that, once a cloud layer is introduced and destabilization is initiated, the destabilization value decreases with time.

Values of destabilization have been computed at individual stations taking into account the amount of cloud and the height of cloud base and top as observed. The computations were then repeated by introducing given variations, such as changing the amount of cloud, shifting the cloud layer as a whole upward or downward, increasing the thickness of the cloud layer, and introducing an inversion in the lower layers. The variations are introduced arbitrarily in only one parameter at a time, keeping the other parameters unchanged.

3. RESULTS AND DISCUSSION

Figure 1 shows the distribution of destabilization. The pattern is not very impressive because it tends to show singularities at scattered locations. Maximum destabilization (1.84°C·day^{-1}·mb^{-1}) occurs at Peoria, Ill. Large values (>0.5°C·day^{-1}·mb^{-1}) also occur at Victoria, Tex., Green Bay, Wis., and Albany, N.Y. Minimum (zero) destabilization is found over parts of Missouri, Arkansas, Oklahoma, and Kansas.

The distribution as shown in figure 1 does not suggest any clear relationship with the prevailing weather, such as the position of fronts and the areas of low or high pressure. However, since destabilization is essentially caused by the presence of clouds, one should expect its distribution pattern to be directly related to that of some cloud parameter such as cloud amount, thickness, or mean altitude. The distribution of total cloud amount is, therefore, shown in figure 2 for comparison. There is little correspondence between the two patterns except for the common location of zero destabilization and clear skies. One is led to believe, therefore, that thickness and mean altitude of the clouds must also play a role in determining the distribution pattern of destabilization.

Estimates of cloud thickness and cloud altitude for all the reporting stations are available. It is easy to see that, for the same thickness of cloud, destabilization is different for different cloud heights. Also, for the same cloud height, destabilization is different for different cloud thickness. This relationship exists because the blackbody emission from the level of cloud base or cloud top is a function of the temperature at that level.

To understand the influence of cloud thickness and cloud altitude on destabilization, one should probably examine the results of a few stations rather than the distribution patterns of these parameters. For this purpose, the following five stations have been selected.

1. Columbia, Mo. — CBI–445
2. Topeka, Kans. — TOP–456
3. Peoria, Ill. — PIA–532
4. Omaha, Nebr. — OMA–553
5. Green Bay, Wis. — GRB–645

These stations were chosen because they provide a good contrast in the destabilization field shown in figure 1.

Table 1 shows the instantaneous cooling rate (°C/day) at all the reference levels from the surface to 100 mb and the associated destabilization (10^{-2} °C·day^{-1}·mb^{-1}) for the selected stations. Cloud data and surface pressures used in the computations of cooling rate are given in table 2. Note that no strong cooling or warming occurs at any level for stations CBI and TOP, which reported clear skies. The cooling rate for these stations is simply a function of temperature and gradient of water vapor in the vertical (Godbole 1963). For PIA and GRB, which have overcast skies, the cooling rate at the top of respective cloud layers is large compared to that at OMA, which has only a fractional cloud cover (3 oktas). Note also that the height of the cloud top is important in determining the magnitude of cooling. For example, maximum cooling of 51.5°C/day occurred at PIA where cloud top was lowest (950 mb). Cooling rates of 26.4°C/day and 2.7°C/day were computed at GRB and OMA where cloud tops were 900 mb and 350 mb, respectively.

To illustrate the effects on radiative cooling and destablization of varying the parameters one at a time, we have selected a single station (TOP) for study. The results of calculations using TOP data are shown in table 3. For convenience of comparison, the values of cooling rate and associated destabilization from table 1 are reproduced in column 2 of table 3.

Effect of Cloud Cover

Cloud amount is arbitrarily increased from zero (observed) to 8 oktas (i.e., overcast sky). For our first variation, we assumed the cloud base to be at 400 mb and the top at 300 mb (col. 3 of table 3). Cooling of 22.5°C/day occurs at cloud top and warming of 18.6°C/day occurs at cloud base. The effect of the cloud on cooling rate seems to penetrate through all the layers from cloud base down to the ground surface; above cloud top, only two layers are affected. The associated destabilization is 0.41°C·day^{-1}·mb^{-1}.

TABLE 1.—*Instantaneous rate of temperature change (°C/day) and the associated destabilization (10^{-2}°C·day^{-1}·mb^{-1}) caused by long-wave radiation*

Pressure (mb)	CBI-445	TOP-456	PIA-532	OMA-553	GRB-645
100					
	−0.46	−0.48	−0.46	−0.46	−0.48
150					
	−0.18	−0.19	−0.18	−0.22	−0.19
200					
	−1.03	−1.79	−1.15	−0.17	−1.28
250					
	−1.22	−1.20	−1.39	−1.93	−1.38
300					
	−1.11	−0.86	−1.71	−1.51	−1.23
350					
	−1.13	−1.02	−1.30	−2.68	−1.15
400					
	−1.18	−1.22	−0.96	−0.61	−1.20
450					
	−1.37	−1.34	−0.95	−0.50	−1.30
500					
	−1.36	−1.25	−1.08	−0.65	−1.28
550					
	−1.32	−1.11	−1.37	−0.84	−1.28
600					
	−1.27	−1.06	−1.45	−0.89	−1.31
650					
	−1.20	−0.80	−1.18	−1.00	−1.20
700					
	−0.99	−0.90	−0.97	−1.03	−1.12
750					
	−1.08	−1.04	−1.21	−1.19	−1.35
800					
	−1.17	−0.96	−1.25	−1.29	−1.50
850					
	−0.98	−0.94	−1.32	−0.95	−1.43
900					
	−1.28	−1.36	−1.67	−1.15	−26.40
950					
	−1.54	−1.52	−51.50	-------	0.00
Surface					
Destabilization	0	0	184	41	53

TABLE 2.—*Observed surface pressure and cloud amount* (N) *with heights of base and top of the cloud*

Station	Surface pressure (mb)	N (okta)	Cloud data base (mb)	Top (mb)
CBI–445	970	0	---	---
TOP–456	967	0	---	---
PIA–532	978	8	978	950
OMA–553	947	3	400	350
GRB–645	973	8	950	900

TABLE 3.—*Instantaneous rate of temperature change (°C/day) and the associated destabilization* (10^{-2}°*C·day*$^{-1}$·*mb*$^{-1}$) *dueto changes in cloud amount* (N), *cloud altitude* (H), *cloud thickness* (ΔH), *and temperature lapse rate.* N *is in oktas and* H *is in mb.*

Pressure (mb)	$N=0$ $H=0$ $\Delta H=0$ (existing)	$N=8$ $H=350$ $\Delta H=100$	$N=8$ $H=650$ $\Delta H=100$	$N=8$ $H=900$ $\Delta H=100$	$N=8$ $H=625$ $\Delta H=650$	$N=0$ Inversion
100						
	−0.48	−0.48	−0.48	−0.48	−0.48	−0.48
150						
	−0.19	−0.19	−0.19	−0.19	−0.19	−0.19
200						
	−1.79	−1.85	−1.80	−1.79	−1.35	−1.79
250						
	−1.20	−1.38	−1.20	−1.20	−1.38	−1.20
300						
	−0.86	−22.51	−0.87	−0.86	−22.51	−0.86
350						
	−1.02	18.51	−1.04	−1.02	0	−1.02
400						
	−1.22	0.25	−1.25	−1.22	0	−1.22
450						
	−1.34	−0.03	−1.41	−1.36	0	−1.34
500						
	−1.25	−0.13	−1.38	−1.27	0	−1.25
550						
	−1.11	−0.23	−1.38	−1.14	0	−1.11
600						
	−1.06	−0.25	−28.69	−1.10	0	−1.05
650						
	−0.80	−0.17	7.64	−0.85	0	−0.78
700						
	−0.90	−0.24	0.30	−0.98	0	−0.85
750						
	−1.04	−0.45	−0.10	−1.14	0	−0.76
800						
	−0.95	−0.53	−0.12	−1.26	0	−1.58
850						
	−0.94	−0.45	−0.05	−30.12	0	−2.16
900						
	−1.36	−0.93	−0.63	−0.83	−0.83	−0.94
950						
	−1.52	−1.08	−0.81	0	0	−0.61
Surface						
Destabilization	0	41	36	29	3.3	(1.3)

Next, the cloud layer as a whole is shifted downward keeping the cloud amount and thickness constant. In the first instance, the cloud base is put at 700 mb and the top at 600 mb. In the second instance, the base is lowered farther to 950 mb with the top at 850 mb. Columns 4 and 5 of table 3, which show the corresponding results, suggest that the destabilization is less when the cloud layer is at lower altitude (Möller 1951). Columns 3, 4, and 5 also show that cooling is greater when the cloud top is lower, whereas warming is greater when the cloud base is higher (Haltiner and Martin 1957). The rate of warming at the base is less than the rate of cooling at the top.

Column 6 shows the effect of stretching a cloud layer in which the base is at 950 mb and the top is at 300 mb.

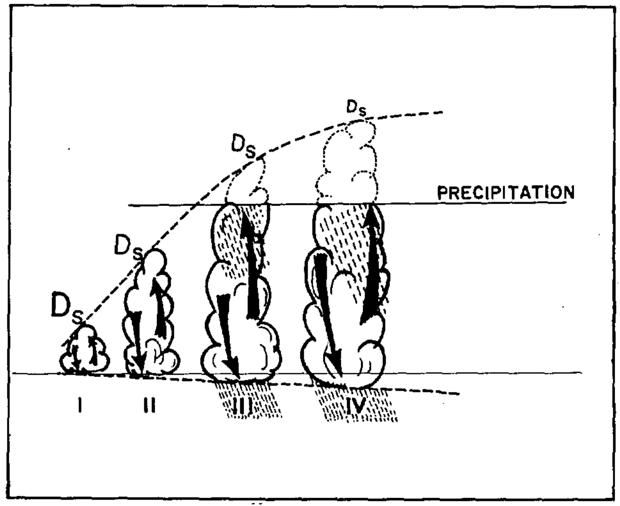

Figure 3.—Schematic representation of cloud growth with associated magnitude of destabilization, D_s.

Destabilization (0.033°C·day^{-1}·mb^{-1}) decreases as cloud thickness increases, obviously because the process of destabilization operates over a larger vertical extent. Cooling at cloud top or warming at cloud base, or both, would decrease depending upon how the cloud layer is stretched in the vertical. In a typical towering cumulus development where the condensation level remains more or less constant (cols. 5, 6, table 3), the cooling from the cloud top decreases progressively with development accompanied by a gradual reduction in the amount of destabilization. In other words, on the basis of radiational processes alone, the cloud development appears to be a self-sustaining process in which an initial value of large destabilization causes convective overturning that further leads to cloud development at the expense of destabilization. A diagram showing successive cloud development accompanied by smaller and smaller destabilization values is presented in figure 3. In the presence of long-wave radiation processes alone, a cloud will continue to develop vertically until destabilization becomes zero (limiting value). However, because of dynamical and microphysical processes, the cloud will tend to precipitate and dissipate before this limiting value is reached. Early morning showers are known to be the result of cloud destabilization.

Stratiform clouds frequently develop into cumuliform-type clouds through the destabilization process; the resulting cellular motions lead to clear or thin cloud areas in the downdraft regions (Haltiner and Martin 1957). Another example, which is due in part to this destabilizing effect, is the nocturnal development of thunderstorms from clouds that could not develop beyond the swelling cumulus stage during the day (Hess 1959). Observations generally indicate that, on such occasions, cloud development is rapid in the initial stages and then slows down. These observations indirectly support the one order of magnitude difference between destabilization values for a shallow cloud (table 3, col. 5) and a thick cloud (col. 6). Determination of whether or not the destabilization actually decreases exponentially as depicted in figure 3 is deferred until a later report.

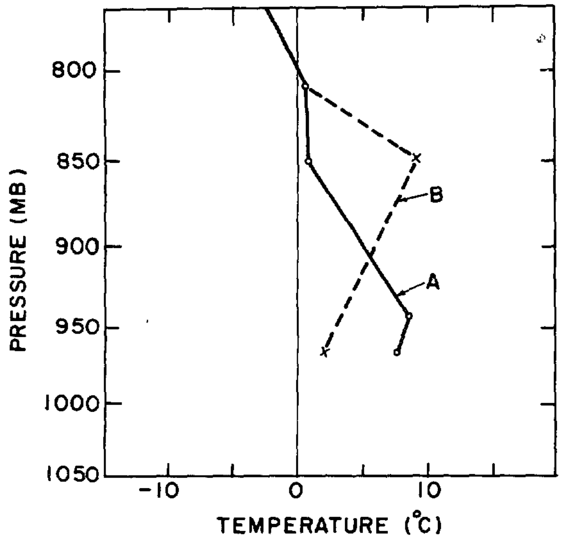

Figure 4.—Vertical distribution of temperature (°C) for (A) observed case and (B) inversion case.

Effect of Inversion

Figure 4 shows how an inversion is introduced arbitrarily in the lower layer. The top of the inversion layer is at 850 mb and the base is at the ground surface. Cooling rate values obtained with this inversion are shown in column 7 of table 3. Maximum cooling (2.2° C/day) occurs at the top of the inversion layer, suggesting that the presence of an inversion layer is equivalent to that of a cloud layer covering a fraction of a celestial dome (col. 5, table 1) or a cloud having partial blackbody characteristics. The amount of destabilization across the inversion layer is $0.013°C \cdot day^{-1} \cdot mb^{-1}$, which is small compared to that across the actual cloud layer.

An inversion just below the cloud layer should offset the warming at the cloud base. This effect is illustrated in columns 5 and 6 of table 3, where an inversion between the surface and 950 mb (fig. 4) gives rise to cooling instead of warming at the cloud base.

4. CONCLUSIONS

The foregoing results may be summarized as follows:

1. Destabilization is sensitive to changes in cloud amount.
2. Destabilization increases as the height of the cloud layer increases and decreases as the thickness of the cloud layer increases.
3. Destabilization contributes to the cloud development at its own expense.
4. An inversion layer is equivalent to a cloud layer from the standpoint of radiative exchange processes and is associated with weak destabilization.

The above results bring out the importance of having an accurate knowledge of both the total cloud amount and cloud type (high, medium, or low). In addition, it is essential to know accurately the heights of cloud bases and cloud tops. The present investigation suffers in this respect. Instead of considering low, medium, and high cloud layers separately, we considered only a single cloud layer, using the base of lowest cloud and the top of the highest cloud as the respective base and top of the layer. In addition, we approximated the heights of the cloud bases and tops to the nearest reference levels, which are spaced at 50-mb intervals. To that extent, we have introduced uncertainties into the computations.

It is obvious that the effects of cloud destabilization are pronounced during the night when heating by solar radiation is absent. It would be interesting, therefore, to examine the relationship between cloud destabilization and nighttime precipitation. The convective instability created by destabilization could be parameterized to obtain estimates of precipitation, which could be compared with observations. Absolute precipitation amounts alone would probably not be meaningful, but a comparison of the variation in destabilization and accompanying precipitation with the variation in observed precipitation could prove enlightening. A study along this line is in progress and will be the subject of a future report.

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