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Research Project MIPR ES-7-967
Project No. 6698
Task No. 669802
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Scientific Report No. 2
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NAVAL POSTGRADUATE SCHOOL
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Abstract

Calculations with a synoptic case study show that long-wave radiative cooling tends to reduce the available potential energy, especially in the upper troposphere. Synoptic-scale precipitation amounts resulting from destabilization of clouds by long-wave cooling are computed. These range up to 1.4 mm in 12 hours. This destabilizing effect may be important in explaining the nocturnal maximum of precipitation over the sea. It may also contribute significantly to cyclone development.

1. Introduction

This paper is a continuation of Scientific Report No. 1 published earlier under this Research Project (Danard 1968).

Calculations of Katayama (1967) show that long-wave radiation, absorption of solar radiation and release of latent heat are about equally important in the annual mean generation of available potential energy over the northern hemisphere. Section 2 is concerned with the first of these processes. Cloud-top cooling reduces temperature contrasts if the cloud mass is in the warm air.

The existence of a nocturnal maximum in frequency of precipitation over the sea has been demonstrated by Kraus (1963). He studied the records of weather ships in the eastern Pacific and North Atlantic Oceans. With 8-10 years of data for July, he found that precipitation was reported in 19 per cent of all nocturnal synoptic reports compared to 12 per cent of the daytime reports.

Kraus proposed that during the day, absorption of short-wave radiation reduces the production of liquid water in clouds, thereby lowering the probability of precipitation. At night, this effect is absent.

The nocturnal maximum in precipitation occurrence over the central United States in summer was studied by Bleeker and Andre (1951). Their explanation was that low-level differential cooling at night due to variations in topography would result in upward motion between the

Appalachians and the Rockies.

In Section 3 it is suggested that an important factor is the destabilizing influence of long-wave cooling at night from cloud-top. The amount of precipitation associated with convective overturning is computed. The destabilization may also augment the large-scale vertical velocity. This would further increase the precipitation as well as accelerate cyclone development.

2. Effects of long-wave radiation on the available potential energy.

Consider the thermodynamic energy equation

$$\frac{\partial \alpha}{\partial t} = - \vec{V} \cdot \nabla \alpha + \sigma \omega + \frac{RH}{C_p p} \tag{1}$$

where $\sigma = -\alpha \partial \ln \theta / \partial p$, $\omega = dp/dt$, and H is the rate of input of heat per unit mass. In (1), let $\alpha = \overline{\alpha} + \alpha'$, $\sigma = \overline{\sigma} + \sigma'$, etc., where the bar denotes the average over an isobaric surface of area S and the prime the departure therefrom. Eq. (1) may then be transformed to (Danard 1966 or Wiin-Nielsen 1964)

$$\frac{\partial A}{\partial t} = - \frac{1}{2g} \int_{p_t}^{p_b} \int_S \vec{V} \cdot \nabla \alpha'^2 \, dS \, dp + \frac{1}{g} \int_{p_t}^{p_b} \int_S \alpha' \omega' dS \, dp + \frac{1}{g} \int_{p_t}^{p_b} \int_S \frac{\alpha' \sigma' \omega'}{\overline{\sigma}} dS \, dp + \frac{R}{C_p g} \int_{p_t}^{p_b} \int_S \frac{\alpha' H'}{\overline{\sigma} p} dS \, dp \tag{2}$$

where
$$A = \frac{1}{2g} \int_{p_t}^{p_b} \int_S \frac{\alpha'^2}{\overline{\sigma}} dS \, dp \tag{3}$$

will be referred to as the available potential energy of the region (c f. Eq (10) of Lorenz (1955) or Eq (2.2) of Wiin-Nielsen (1964)). In (2) and (3), p_t and p_b denote the pressures at the top and bottom of the region of interest.

If a quasi-geostrophic balance exists, ω' will be related to H'. However, attention will be restricted here simply to the final term in (2), representing the generation and designated as G:

$$G = \frac{R}{C_p g} \int_{p_t}^{p_b} \int_s \frac{\alpha' H'}{\bar{\sigma} p} \, dS \, dp \tag{4}$$

In (4), H' will be identified with long-wave radiative effects computed in Section 6 of Sci. Rep. No. 1. The area S is bounded by the latitudes 28.7N and 51.3N and by the longitudes 70W and 115W. Computations are performed for three layers: 900-700 mb, 700-500 mb, and 500-300 mb. Results are given in Table 1.

In the lower layers, G has small, positive values. This is due in large part to the absence of cooling in the lower portion of the main cloud mass (see Figs. 8 and 9 of Sci. Rep. No. 1.) In the lowest layer, there is also the effect of cooling from the top of stratocumulus cloud in the cold air over Kansas and Oklahoma (see Fig. 9 of Sci. Rep. No. 1). However, in the uppermost layer, G is negative and fairly large in magnitude. This is due to radiation from cloud-top in the warm air. Thus long-wave radiation tends to reduce the temperature contrast in the upper troposphere, a not unexpected result. Assuming G constant and neglecting the other terms on the right side of (2), the quotient $A/|G|$ represents the time required to effect a change in A equal in magnitude to the initial value. Whereas this time is of the order of weeks or longer in the lower layers, it is of the order of days in the highest layer. This suggests that long wave radiation from cloud-top may be important in the upper troposphere even for short time periods.

Table 1. Comparison of terms A and G . See Eqs (2) - (4).

Term	Units	900-700 mb	700-500	500-300	900-300
A	10^{25} ergs	6.04	7.36	4.97	18.4
G	10^{20} ergs sec^{-1}	0.26	0.08	-1.66	-1.32
$A/\|G\|$	days	27	104	3.5	16

3. Effects of long-wave radiation from cloud-top on precipitation rates

The basic principles in computing the convective precipitation are similar to those employed by Manabe et al (1965). Consider an unstable atmosphere initially at rest. If the motions arising from convective overturning are dissipated as heat, it may be shown that

$$\delta \int C_p T \, dm = \int \delta Q \, dm \tag{5}$$

where δ refers to a change over a finite time interval, dm is an element of mass and δQ is the heat added per unit mass from all sources other than friction. In (5), the integration extends over the entire atmosphere. Eq (5) is derived in the Appendix. In a dry adiabatic transformation the sum of the potential and internal energies (represented by the integral on the left side of (5)) is conserved.

Eq (5) is applied to a layer cloud extending from 1000 to 400 mb. The lapse rate is assumed initially pseudo-adiabatic. To simulate conditions during the night, the cloud top radiates for 12 hours. As this proceeds, the cloud becomes unstable. The convective overturning then produces precipitation and returns the lapse rate to the pseudo-adiabatic value. The liquid water contents of the initial and final states are assumed equal (i.e., all water which condenses during this process falls out as precipitation).

Applied to the above problem (5) may be written

$$- \frac{C_p}{g} \int_{p_4}^{p_o} \delta T \, dp = \Delta Q + \frac{L}{g} \int_{p_4}^{p_o} \delta r \, dp \tag{6}$$

where $p_4 = 400$ mb, $p_o = 1000$ mb, $\delta T (< 0)$ and $\delta r (< 0)$ are the differences between final and initial temperatures and mixing ratios, $\Delta Q (> 0)$ is the heat lost by radiation, and L is the latent heat of vaporization. The total precipitation is

$$P = - \int_{p_4}^{p_o} \delta r \, \frac{dp}{g} \tag{7}$$

The interval between p_4 and p_o is divided into six layers each 100 mb thick. The integrals are then expressed as sums of contributions from these layers. With a known initial state, Eq (6) is thus an equation in 12 unknowns, the final values of T and r at the mid-points of the six layers. Eleven more equations are needed. Five are obtained from the final condition that

$$\frac{\partial \theta_e}{\partial p} = 0 \tag{8}$$

where

$$\theta_e = \left(T + \frac{Lr}{C_p} \right) \left(\frac{P_o}{p} \right)^{R/C_p} \tag{9}$$

is the equivalent potential temperature. The remaining six equations are obtained from the Clausius-Clapeyron equation. If r_i, e_i and T_i denote initial mixing ratio, vapor pressure and temperature, one obtains

$$\delta r = \frac{0.622 \, r_i L}{R T_i^2} \; \frac{p}{p - e_i} \; \delta T \tag{10}$$

Eq (10) is applied at the six levels.

In computing the radiation lost from cloud-top, equations similar to (2) and (13) of Sci. Rep. No. 1 were derived. The level p_u was taken as 400 mb (cloud-top) and q_u was set equal to the saturation value corresponding to the temperature there. The upward flux at cloud-top is thus

$$F_u = F_{bu} (1 - \epsilon_u) \tag{11}$$

where F_{bu} is the black body flux corresponding to cloud-top temperature, and ϵ_u is the water vapor emissivity for the layer above the cloud. The heat lost by radiation is, per unit area,

$$\Delta Q = F_u \Delta t \tag{12}$$

where $\Delta t = 12$ hrs.

The method described above is applied to three cases where the initial state is saturated with a constant wet-bulb potential temperature θ_w of 0, 10 and 20 C. These would represent conditions over polar, mid-latitude and sub-tropical oceans, respectively. Results are shown in Table 2.

The values of ΔQ decrease with increasing initial θ_w. This is because the decrease in $(1 - \epsilon_u)$ in Eq (11) over-compensates the increase in F_{bu}. Release of latent heat tends to offset the cooling due to long-wave radiation. This results in fairly small values of $\delta \theta_w$. Precipitation amounts in Table 2 are to be regarded as averages over large areas, and not as values associated with individual convective

Table 2. Net cooling $(\delta\theta_w)$, heat lost by long-wave radiation (ΔQ) and total precipitation (P) during 12-hr period for cloud extending from 1000 to 400 mb.

Initial θ_w (deg C)	$\delta\theta_w$ (deg C)	ΔQ (10^9 ergs cm^{-2})	P (mm)
0	1.2	11.2	0.9
10	0.8	9.8	1.4
20	0.3	6.1	1.4

clouds. The amounts increase slightly with initial θ_w. Note that a cooling of only 0.3 C for $\theta_w = 20$ C produces more precipitation than does a cooling of 1.2 C for $\theta_w = 0$ C.

Precipitation amounts given in Table 2 are fairly modest. However, in a quasigeostrophic numerical prediction model, the large-scale vertical motion varies inversely as the value assigned to the static stability. Thus the cloud destabilization may lead to an increase in the large-scale vertical velocity also. This would contribute further to the amount of precipitation.

Appendix - Derivation of Eq (5)

Consider the equation for 3-dimensional motion

$$\frac{d\vec{V}}{dt} = -\alpha \nabla p - 2\vec{\Omega} \times \vec{V} + \vec{g} + \vec{F} \tag{A1}$$

where $\vec{V} = u\vec{i} + v\vec{j} + w\vec{j}$, $\nabla = \vec{i}\ \partial/\partial x + \vec{j}\ \partial/\partial y + \vec{k}\ \partial/\partial z$ and $\vec{F}$ is the frictional force associated with motions arising from convective overturning. Multiplying Eq (A1) scalarly by $\vec{V}$ gives

$$\frac{d}{dt}\left(\frac{V^2}{2} + gz\right) = -\alpha \vec{V} \cdot \nabla p + \vec{V} \cdot \vec{F} \tag{A2}$$

The first law of thermodynamics may be expressed as

$$C_v \frac{dT}{dt} = \frac{dQ}{dt} - p\frac{d\alpha}{dt} \tag{A3}$$

Adding (A2) and (A3) and using the equation of continuity

$$\frac{d\alpha}{dt} = \alpha \nabla \cdot \vec{V} \tag{A4}$$

yields the equation

$$\frac{d}{dt}\left(\frac{V^2}{2} + gZ + C_v T\right) = -\alpha \nabla \cdot p\vec{V} + \frac{dQ}{dt} + \vec{V} \cdot \vec{F} \tag{A5}$$

Now if F is any function, use of (A4) gives

$$\rho \frac{dF}{dt} = \frac{\partial}{\partial t} \rho F + \nabla \cdot \rho F \vec{V} \tag{A6}$$

Integrating (A5) with respect to mass over the entire atmosphere, and making use of (A6) and the relation

$$\int_0^\infty \rho (gZ + C_v T)\, dz = \int_0^\infty \rho C_p T\, dz \tag{A7}$$

one obtains

$$\frac{\partial}{\partial t}\int \left(\frac{V^2}{2} + C_pT\right) dm = \int \frac{dQ}{dt} dm + \int \vec{V} \cdot \vec{F}\, dm \qquad \text{(A8)}$$

where $dm = \rho dz\, dA$ is an element of mass.

Now if the motions arising from convective overturning are dissipated as heat,

$$\delta \int \frac{V^2}{2}\, dm = 0 \qquad \text{(A9)}$$

Furthermore,

$$\int \delta Q_f dm = -\int \left[\int \vec{V} \cdot \vec{F}\, dm \right] dt \qquad \text{(A10)}$$

where δQ_f is the heat added per unit mass due to friction and the time integration extends over the interval to which δQ_f refers. Thus (A8) becomes

$$\delta \int C_pT\, dm = \int \delta Q\, dm \qquad (5)$$

where δQ refers to heat added from sources other than friction.

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