A parametric method for computing the mean water budget of the atmosphere

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ABSTRACT

Using the equation of conservation of water vapor in the troposphere, the difference, evaporation at the surface minus precipitation, is expressed as a function of total cloudiness (or mean tropospheric relative humidity) and of temperature and horizontal wind at the surface and in mid-troposphere.

Computations of the annual normal values of evaporation minus precipitation using the resulting formulas together with monthly averages of the required parameters for the Northern Hemisphere, agree well in the general patterns with estimates made by Starr, Peixoto and Crisi, as well as with climatological values. However, there are large small-scale discrepancies in the three estimates.

The computations have a strong dependence on the divergence of the horizontal wind, and some of the discrepancies must be attributed to errors in evaluating the latter.

To determine the importance of anomalies of the difference, evaporation minus precipitation, computations for February 1964 and February 1962 are carried out, showing that the values of the difference between these two months are of the same order of magnitude as the monthly values themselves.

If positive values indicate wind from the south, the values of evaporation minus *precipitation are inversely correlated to the meridional wind component on a broad* scale. Furthermore the computations show that the anomalies in evaporation minus precipitation depend strongly not only on the anomalies of the meridional wind component but also on the anomalies of precipitable water, which in turn depend on the anomalies of surface temperature and cloudiness (or mean tropospheric relative humidity).

1. Introduction

Several authors have used the equation of conservation of water vapor in the atmosphere to make estimates of the difference, evaporation minus precipitation. Among them Benton and Estoque (1954), who made computations for North America, and Starr, Peixoto and Crisi (1965) who made estimates for the whole Northern Hemisphere.

The primary purpose of this paper is somewhat different from that of the afore-mentioned authors. Instead of aiming at making the best possible estimates from the most complete available data at different levels of the atmosphere, as they do, we introduce simplifying assumptions, so that it may be possible to compute the mean water budget from a few parameters which appear to be the essential ones.

This parametric method of computing the mean water budget may find applications in general circulation studies as well as in numerical prediction. In fact, the author's interest in this subject arose from his modeling of the troposphere-surface of the earth system, applying the equation of conservation of thermal energy to long-range forecasting (Adem, 1964, 1965). This model included a method for the parameterization of heat lost by the surface of the earth due to evaporation, as well as the heat gained by the atmosphere due to condensation of water vapor in the clouds, which give rise to quasi-independent formulas which may violate the conservation of water vapor in the atmosphere. By introducing an equation for conservation of water vapor, we not only avoid violating this conservation law, but also introduce a new way to generate the addition of thermal energy in the Troposphere-Ocean-Continent System due to evaporation at the surface and condensation of water vapor in the clouds.

2. The equation of water balance in the atmosphere

The conservation of water vapor in the atmosphere can be written as

$$\frac{\partial}{\partial t}(\overline{\varrho^* q}) + \nabla_T \cdot (\bar{\mathbf{v}}_T \overline{\varrho^* q} + \bar{\varrho}^* \overline{\mathbf{v}_T' q'}) = \bar{g}_5 \qquad (1)$$

where the variables with a bar are average values over a given time interval, and those with prime are the departures from the average values; $\varrho^* q$ is the water vapor per unit volume, ϱ^* is the density of the humid air, q the specific humidity, $\mathbf{v}_T$ is the wind velocity, g_5 is the rate at which water vapor is added by evaporation per unit volume, t is the time, and Δ_T is three-dimensional gradient operator.

The water balance equation for a vertical column of height H_1 and unit area is obtained by integrating equation (1) from the surface to H_1. The resulting equation is the following:

$$\int_h^{H_1} \left[\frac{\partial}{\partial t}(\overline{\varrho^* q}) + \nabla \cdot (\bar{\mathbf{v}} \overline{\varrho^* q})\right] dz + (\bar{\mathbf{w}} \overline{\varrho^* q})_{z=H_1}$$

$$-(\bar{w}\overline{\varrho^* q})_{z=h} + \int_h^{H_1} \nabla \cdot (\bar{\varrho}^* \overline{\mathbf{v}' q'})\, dz$$

$$+(\bar{\varrho}\overline{w' q'})_{z=H_1} - (\bar{\varrho}^* \overline{w' q'})_{z=h} = \int_h^{H_1} \bar{g}_5 dz \qquad (2)$$

where ∇ is the two-dimensional horizontal gradient operator, h is the height of the earth's surface (used instead of the top of the surface boundary layer) $\mathbf{v}$ is the horizontal component of the wind and w is the vertical component.

The terms

$$(\bar{w}\overline{\varrho^* q})_{z=H_1}, \quad (\bar{w}\overline{\varrho^* q})_{z=h} \quad \text{and} \quad (\bar{\varrho}^* \overline{w' q'})_{z=H_1}$$

are presumably small where the surface of the ground is flat and when H_1 is the 500 mb level or higher. These terms will be neglected over all areas of the Northern Hemisphere since the second term is locally important only in restricted mountain regions.

The term $(\bar{\varrho}^* \overline{w' q'})_{z=h}$ which is the water vapor added to the atmosphere by evaporation at the surface is equal to $\bar{G}_3/L$ where $\bar{G}_3$ is the heat lost at the surface due to the evaporation and L is the latent heat of vaporization, taken as a constant.

The term $\int_h^{H_1} \bar{g}_5 dz$ is equal to $-\bar{G}_5/L$ where $\bar{G}_5$ is the heat gained by condensation of water vapor in the clouds.

Using the above assumptions and notation equation (2) becomes:

$$E_1 + E_2 + E_3 = (\bar{G}_3 - \bar{G}_5)/L \qquad (3)$$

where

$$E_1 = \int_h^{H_1} \frac{\partial}{\partial t}(\overline{\varrho^* q})\, dz,$$

$$E_2 = \int_h^{H_1} \nabla \cdot (\bar{\mathbf{v}} \overline{\varrho^* q})\, dz \qquad \text{and}$$

$$E_3 = \int_h^{H_1} \nabla \cdot \bar{\varrho}^* \overline{\mathbf{v}' q'}\, dz.$$

The term E_1 is generally assumed to be negligible when annual averages are considered. However, as pointed out by Rasmusson (1966, p. 19) it is not negligible for monthly averages, particularly during the spring and fall. E_2 is the flux divergence due to the mean wind and E_3 the flux divergence due to the transient eddies.

To determine the relative importance of the meridional components of E_2 and E_3 in formula (3), we shall use the values published by Peixoto (1958), and Peixoto and Crisi (1965) in their studies of the hemispheric humidity conditions for the years 1950 and 1958. We shall use only their zonally-averaged values of the total and transient eddy transports at every 10 degrees of latitude.

Fig. 1 shows the water vapor flux across the specified latitude in units of 10^{11} grams per second, where a linear interpolation has been used between the ten-degree latitude values taken from the above reference. The continuous line represents the total flux and the dashed line the transient-eddy flux. The dotted line is the standing-eddy flux obtained by subtracting the transient-eddy flux from the total. Parts A and C show the values for 1950, for "Winter" (six months average from October to March) and "Summer" (six months average from April

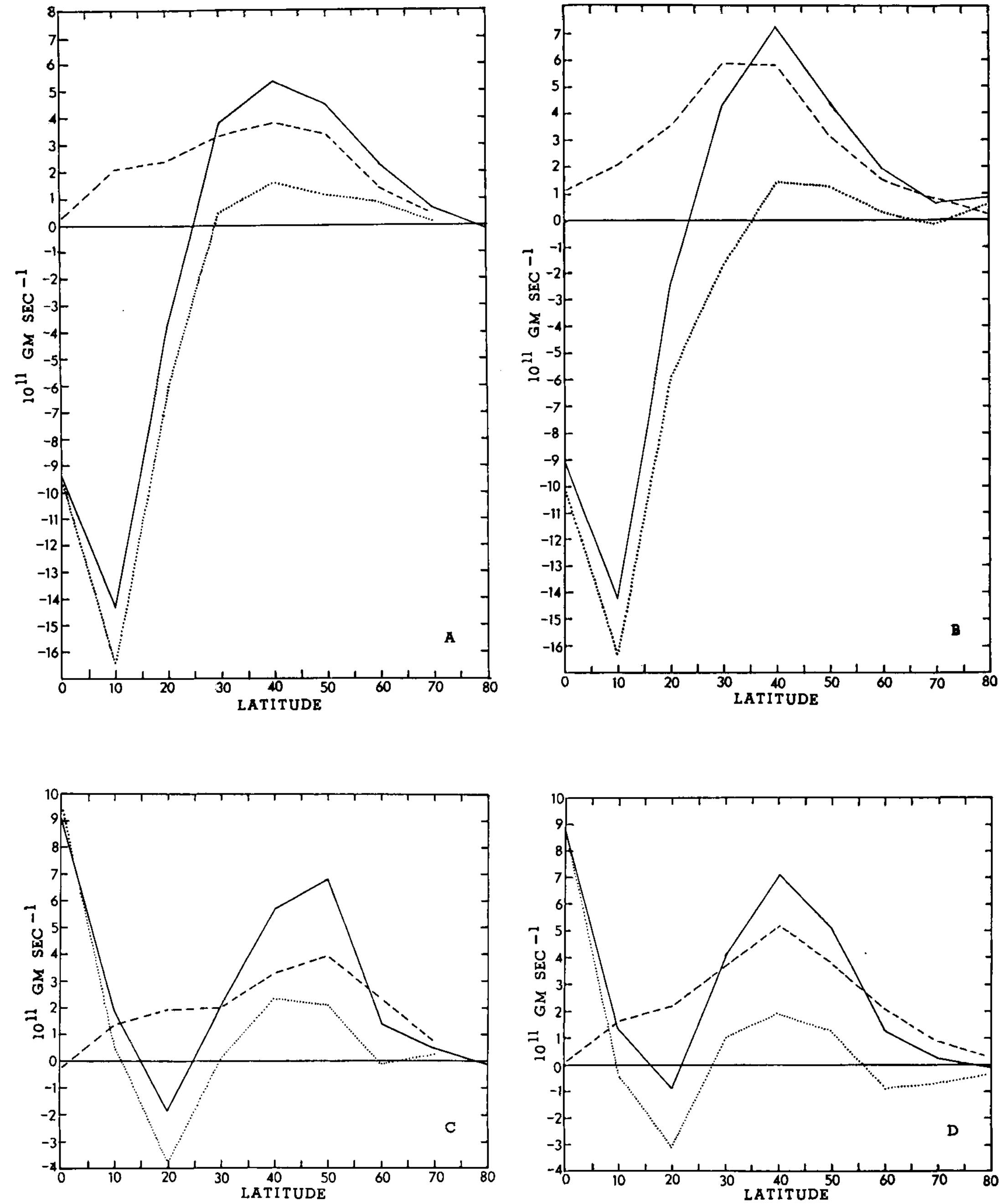

Fig. 1. Zonally averaged values of the mean meridional transport of water vapor, in 10^{11} grams per second: The continuous line is the total transport, the dashed line is the transport by transient eddies, and the dotted line the transport by standing eddies. (*A*) and (*C*) are the Winter and Summer values, respectively, for 1950, and (*B*) and (*D*) the corresponding values for 1958.

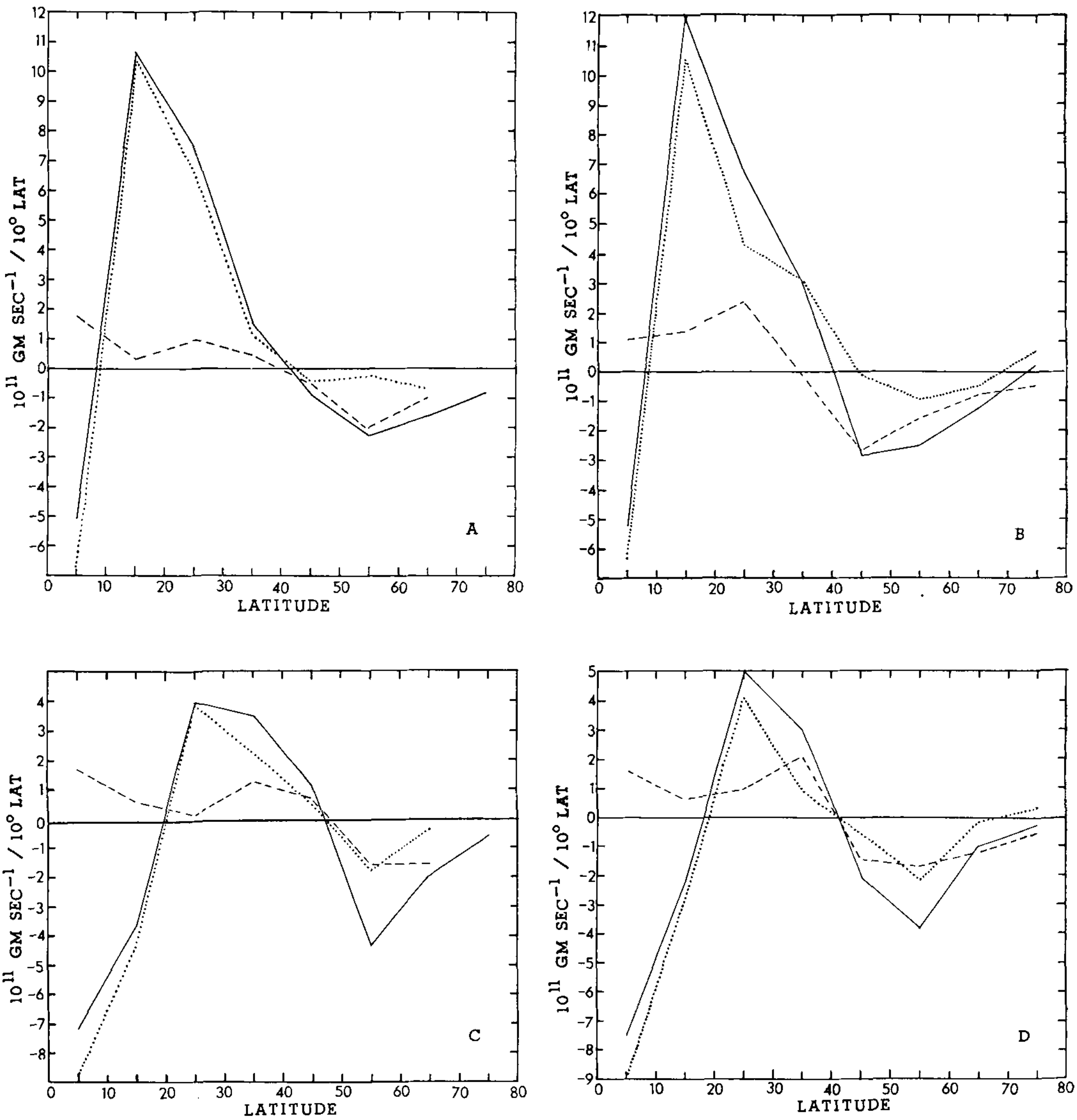

Fig. 2. Zonally averaged values of the mean meridional flux divergence of water vapor in 10^{11} grams per second per ten degrees of latitude: The continuous line is the total flux divergence, the dashed line is the flux divergence due to transient eddies, and the dotted line the flux divergence due to standing eddies. (*A*) and (*C*) are the Winter and Summer values, respectively, for 1950, and (*B*) and (*D*) the corresponding values for 1958.

to September) respectively; and B and D the corresponding values for 1958.

From Fig. 1 it is evident that both the transient and the standing eddy meridional transports are far from being negligible, the main contribution being in middle latitudes for the transient eddies and in lower latitudes for the standing eddies. However, the latitudinal variation of the standing-eddy flux resembles much more that of the total flux than does that of the transient-eddy flux. This suggests that the flux divergence due to standing eddies (which is proportional to the slope of the curves in Fig. 1) is more important than one might conclude by comparing the flux values.

To substantiate this hint, and in order to be able to draw more specific conclusions, we have computed the flux divergence in the region between each two consecutive ten-degree latitude circles, using the flux values shown in Fig. 1.

Fig. 2 shows the result of this computation. The continuous line is the total flux divergence, the dotted line the standing-eddy flux divergence and the dashed line the transient-eddy flux divergence. The units are 10^{11} grams per second per ten degrees latitude. A and C show the values for 1950, for "Winter" and "Summer" respectively; and B and D the corresponding values for 1958.

It is evident from Fig. 2 that a fair approximation to the meridional flux divergence is obtained when one neglects the part associated with the transient eddies, especially in lower latitudes.

It can be shown, with a similar computation, that the contribution of the flux divergence due to the zonal component of E_3 is negligible compared with that due to the corresponding component of E_2. Therefore, since at present we lack a way of expressing the transient-eddy flux divergence as a function of the mean variables, it seems reasonable to start by parameterizing and evaluating the other terms that appear on the left side of equation (3) and consider that E_3 is equal to zero. A comparison of the results of the computation using this drastic assumption with other estimates of $\bar{G}_3 - \bar{G}_5$ will show the importance of the transient eddies as a function of the geographic locality.

3. Parameterization of evaporation minus precipitation

In a previous paper (Adem, 1967), which from here on will be identified by (I), we have shown that the average water vapor per unit volume can be expressed as

$$\overline{\varrho^* q} = \frac{0.622}{R} P_1 \bar{f} \tag{4}$$

where R is the gas constant, $\bar{f}$ the average relative humidity; P_1 is a function of the average temperature and is equal to $e_s(T^*)/T^*$, where $e_s(T^*)$ is the saturation water vapor corresponding to the average temperature and T^* is the average temperature.

To evaluate $e_s(T^*)$ we will use the same formula given in (I), which is based on the assumption that the water contained in the layer of height H_1 is supercooled where the temperature is below freezing. This condition is satisfied if we take H_1 sufficiently small so that the temperatures are above say $-20°C$.

Since the total amount of water vapor contained above 500 mb is assumed to be negligible compared with the amount below that surface, we will take $H_1 = 5.5$ km. Therefore, except at higher latitudes, the assumption that there is supercooled water in the clouds is probably justified.

In (I) we have shown that the average relative humidity $\bar{f}$ at a height z can be expressed as

$$\bar{f} = \bar{f}_m + f_1(z) \tag{5}$$

where $\bar{f}_m$ is the average tropospheric relative humidity and $f_1(z)$ is an empirical function of the vertical coordinate, given by

$$f_1 = A_0 + A_1 z + A_2 z^2$$

where A_0, A_1 and A_2 are constant parameters that vary with the seasons.

Furthermore, we have shown in (I) that the mean tropospheric relative humidity, $\bar{f}_m$, can be expressed as a function of the total cloudiness, $\bar{\varepsilon}$, by the formula

$$\bar{f}_m = B_1 + B_2 \bar{\varepsilon} \tag{6}$$

where B_1 and B_2 are constant parameters that vary with the seasons.

We will assume a generalized lapse rate given by

$$\bar{\beta} = \frac{T_s - T}{H_1 - h} \tag{7}$$

where β is the lapse rate, T_s the surface temperature, T the temperature at height H_1 and h the surface topography.

Therefore

$$T^* = \bar{\beta}(H_1 - z) + T \tag{8}$$

For the sake of simplicity we shall assume that the wind field is given by

$$\bar{\mathbf{v}} = \frac{z-h}{H_1 - h} \bar{\mathbf{v}}_1 + \frac{H_1 - z}{H_1 - h} \bar{\mathbf{v}}_s \tag{9}$$

where $\bar{\mathbf{v}}_s$ is the surface wind and $\bar{\mathbf{v}}_1$ is the wind at 500 mb (which is a very good approximation to the wind at $H_1 = 5.5$ km).

Using (7) and (8), formula (9) can be written

$$\bar{V}^* = \frac{T_s - T^*}{T_s - T}\bar{V}_1 + \frac{T^* - T}{T_s - T}\bar{V}_s \qquad (10)$$

Admittedly formula (10) is a very crude approximation. However, since the results are vertically integrated the overall horizontal patterns should be essentially correct.

Substituting (4), (5), (8), and (10) in (3), we obtain the following equation:

$$\frac{\bar{G}_3 - \bar{G}_5}{a'} = (I_{4s}^* \nabla T + J_{4s}^* \nabla \bar{\beta} + J_5^* \nabla h + I_{6s} \nabla \bar{f}_m) \cdot \bar{V}_s$$

$$+ (I_4^* \nabla T + J_4^* \nabla \bar{\beta} - J_5^* \nabla h + I_6 \nabla \bar{f}_m) \cdot \bar{V}_1$$

$$+ I_{6s}^* \nabla \cdot \bar{V}_s + I_6^* \nabla \cdot \bar{V}_1 + I_1^* \frac{\partial T}{\partial t}$$

$$+ I_{1\beta}^* \frac{\partial \bar{\beta}}{\partial t} + I_7 \frac{\partial \bar{f}_m}{\partial t} + E_3 \qquad (11)$$

where $$a' = \frac{0.622L}{R},$$

$$I_1^* = \frac{1}{\bar{\beta}} \int_T^{T_s} (\bar{f}_m + f_1) \frac{dP_1}{dT^*} dT^*,$$

$$I_4^* = \frac{1}{\bar{\beta}(T_s - T)} \int_T^{T_s} (\bar{f}_m + f_1) \frac{dP_1}{dT^*} (T_s - T^*)\, dT^*$$

$$I_{1\beta}^* = \frac{T_s - T}{\bar{\beta}} (I_1^* - I_4^*), \quad I_{4s}^* = I_1^* - I_4^*,$$

$$I_6^* = \frac{1}{\bar{\beta}(T_s - T)} \int_T^{T_s} (\bar{f}_m + f_1) P_1 (T_s - T^*)\, dT^*,$$

$$I_6 = \frac{1}{\bar{\beta}(T_s - T)} \int_T^{T_s} P_1 (T_s - T^*)\, dT^*,$$

$$I_{6s}^* = \frac{1}{\bar{\beta}(T_s - T)} \int_T^{T_s} (\bar{f}_m + f_1) P_1 (T^* - T)\, dT^*,$$

$$I_7 = \frac{1}{\bar{\beta}} \int_T^{T_s} P_1\, dT^*, \quad J_5^* = \frac{\bar{\beta}}{(T_s - T)} I_{6s}^*, I_{6s} = I_7 - I_6,$$

$$J_4^* = \frac{1}{\bar{\beta}^2 (T_s - T)} \int_T^{T_s} (\bar{f}_m + f_1) \frac{dP_1}{dT^*} (T_s - T^*) \times (T^* - T)\, dT^*,$$

$$J_{4s}^* = \frac{1}{\bar{\beta}^2 (T_s - T)} \int_T^{T_s} (\bar{f}_m + f_1) \frac{dP_1}{dT^*} (T^* - T)^2\, dT^*.$$

In the evaluation of the integrals we have used T^* as the variable of integration instead of z, in order to show their dependence on the surface temperature and the 500-mb temperature.

Expressing e_s as a fourth degree polynomial in T^* as was done in (I), the above integrals may be evaluated analytically.

Formula (11) together with (7) and (8), expresses the average over a given time interval of the difference, evaporation minus precipitation $(\bar{G}_3 - \bar{G}_5)/L$, as a function of the mean tropospheric relative humidity $\bar{f}_m$, the surface temperature T_s, the 500-mb temperature T, the surface wind $\bar{V}_s$ and 500-mb wind $\bar{V}_1$, averaged over the same interval. The local rates of change of the mean relative humidity $\partial \bar{f}_m / \partial t$, of the surface temperature $\partial T_s / \partial t$, and of the 500-mb temperature $\partial T / \partial t$, also appear in (11).

Substituting (6) in formula (11), we can use as a variable the total cloudiness instead of the mean tropospheric relative humidity. This substitution in (11) will replace $\bar{f}_m$, $\Delta \bar{f}_m$, and $\partial \bar{f}_m / \partial t$ by $B_1 + B_2 \bar{\varepsilon}$, $B_2 \Delta \bar{\varepsilon}$ and $B_2 \partial \bar{\varepsilon} / \partial t$ respectively. Since $\bar{G}_3$ is the heat lost at the surface by evaporation and $\bar{G}_5$ is the heat gained by condensation of water vapor in the clouds, (11) also represents a parameterization of the difference $\bar{G}_3 - \bar{G}_5$ which may be used in heat budget studies.

There are 3 types of terms in the right side of (11):

(1) The sum of the terms containing $\bar{V}_s$, $\bar{V}_1$, $\nabla \cdot \bar{V}_s$, and $\nabla \cdot \bar{V}_1$ is the horizontal divergence of the transport of water vapor due to the mean horizontal wind.

(2) The sum $I_1^* \partial T / \partial t + I_{1\beta}^* \partial \bar{\beta} / \partial t + I_1 \partial \bar{f}_m / \partial t$ which represent the storage of water vapor in the atmosphere.

(3) The flux divergence due to transient eddies which is represented by E_3, and which in accordance with the discussion given in the previous section, will be neglected.

4. Annual normal values of evaporation minus precipitation in the northern hemispere

We shall carry out computations with formula (11) using monthly normal values of the required independent parameters.

The normal surface air temperatures used

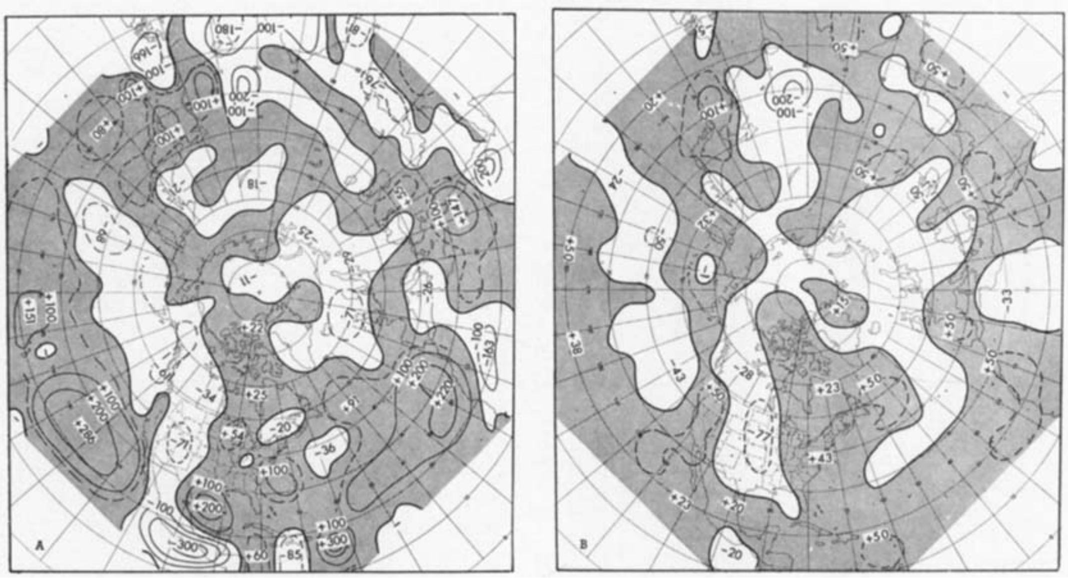

Fig. 3. Annual values of evaporation minus precipitation, in cm per year: (*A*) shows the computed normal values from formula (9); and (*B*) the computed values when the divergence of the horizontal wind is neglected.

were based largely on recent WMO normals (1962). The normal 500-mb temperatures were computed from upper-air height data prepared by Hennig (1958). The normal cloud cover was obtained from London (1957), and the mean surface topography from Berkofsky and Bertoni (1965). The normal 1000-mb surface winds were taken from the recently published charts by the U.S. Navy (1966), and the normal 500-mb winds from the geostrophic wind charts published by Lahey *et al.* (1958).

In order to compare the results of the computations with values of Starr *et al.* (1965) for 1958, we computed annual values. This was done by averaging the computed values for January, April, July and October. The storage terms were neglected.

The annual values of the difference, evaporation minus precipitation, are shown in Figs. 3 and 4. Fig. 3*A* shows the computed normal values using formula (11); 4*A* the annual values obtained by Starr *et al.* for 1958; 4*B*, "climatological" values, obtained by evaluating the evaporation using Budyko's (1963) monthly values over oceans and Thornthwaite's method (1948) applied to monthly normal data over continents (Taubensee, 1966), and using for precipitation Möller's (1951) seasonal charts. The annual climatological values were also obtained as an average of the values for January, April, July and October (average of the four seasons in the case of precipitation).

Comparison of 3*A* with 4*A* and 4*B* shows generally good agreement in the overall patterns. However, there are very large small-scale discrepancies in the three estimates. Some of the discrepancies with Starr's *et al.* values may be attributed to the fact that the latter are for 1958, while ours are normal values.

More recent computations by Rasmusson (1966) and by Bock, Frazier and Welsh (1966) for North America show similar overall patterns as 3*A* and 4*A*. However, all these computations show substantial small-scale discrepancies among them, which are as large as those with ours.

Since we have neglected the transient eddies, it is evident from this agreement in the overall patterns that the most important contribution is the one due to the flux divergence associated with the monthly mean wind, in support of the conclusion obtained in section 2.

Figure 3*B* shows the result of the computations when we neglect the term containing the divergence of the wind. Its comparison with 3*A* shows that the most important term in the

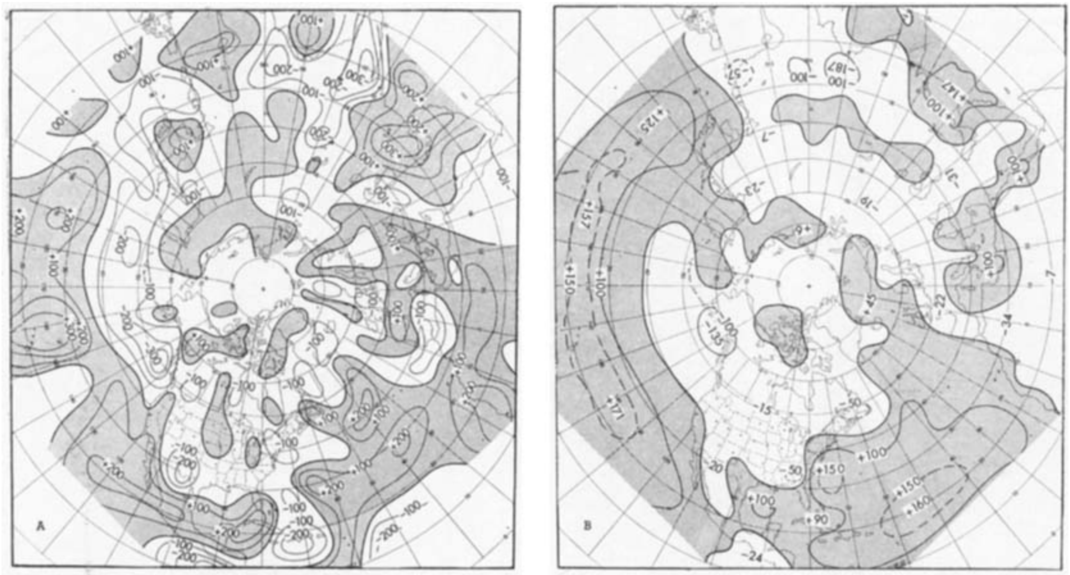

Fig. 4. Annual values of evaporation minus precipitation, in cm per year: (A) shows the values for 1958 after Starr *et al.*, and (B) the climatological normal values.

evaluation of $\bar{G}_3 - \bar{G}_5$ is the term containing the divergence of the wind. Since the contribution of the divergence of the 500 mb winds is small, it follows that the most important term is the one that depends on the divergence of the lower winds (1000 mb). Therefore, some of the discrepancies that appear in all the computations can be attributed to errors in the evaluation of the divergence of the wind.

5. Main variables on which the difference, evaporation minus precipitation, depends

In this section we shall consider only computations for winter months, and will neglect the storage term in addition to the eddy term E_3.

Fig. 5A shows the computed January values of evaporation minus precipitation multiplied by the latent heat of evaporation to get energy units, in ly per day; and Fig. 5B the meridional component of the 1000 mb winds, where positive values indicate wind from the south. Comparison of 5A and 5B shows a striking inverse relationship between the difference, evaporation minus precipitation, and the meridional wind.

This relationship is due to the strong dependence of evaporation minus precipitation on the divergence of the lower wind, which in turn has a strong dependence on the geostrophic divergence which is equal to $-(\beta_1/f_1)\bar{v}_M$ where β_1 is the north–south gradient of the Coriolis parameter, f_1, and $\bar{v}_M$ the geostrophic meridional wind component. The departures from this inverse relationship are due to the dependence of evaporation minus precipitation on the moisture content of the atmosphere.

To explore in further detail this dependence of evaporation minus precipitation on the north–south wind component and on moisture content, one may compute the difference of the former quantity for the same month of two different years and compare the results with the corresponding differences of the other two variables.

We have carried out such a numerical experiment using February 1964 and February 1962. The wind at the surface and at 500 mb was computed geostrophically from the surface pressure and 500-mb heights respectively.

Since the observed cloudiness was not available, we used surface relative humidity together with (5) to compute the mean tropospheric relative humidity, which can be used directly in formula (11) instead of cloudiness. Furthermore this estimated mean tropospheric relative humidity value can be used together

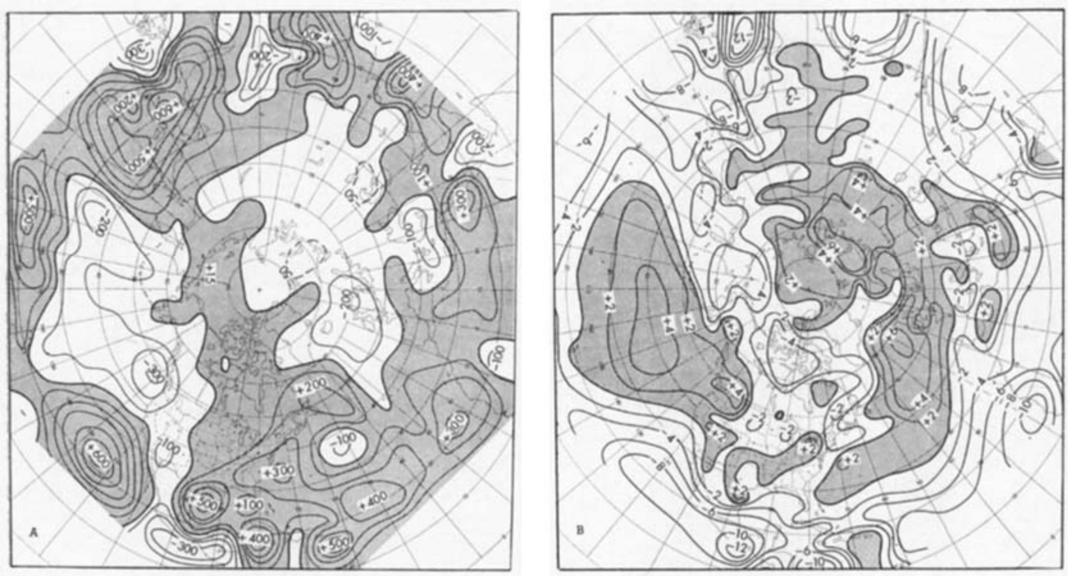

Fig. 5. (*A*) shows the computed normal January values of evaporation minus precipitation, in ly per day; and (*B*) the 1000 mb meridional wind component for the same month, in meters per second where the positive values indicate wind from the south.

with (6) to make a rough estimate of total cloudiness.

Figure 6 shows the difference February 1964 minus February 1962 of the computed values of evaporation minus precipitation. Comparison of these values with those of each of the considered February values shows that the difference of the values of the two months is of the same order of magnitude as the values of either of the months.

Figure 7 shows the difference, February 1964 minus February 1962, of the most important variables on which evaporation minus precipitation depends. Figure 7*A* is the difference of the meridional wind, and in agreement with what was shown above, reveals an inverse relationship with the difference of evaporation minus precipitation (Fig. 6). The departures from this relationship are partly due to the dependence of evaporation minus precipitation on the moisture content. This is clearly seen on comparing Figure 6 with the difference of the precipitable water (computed by integrating formula (4) from the surface to $z = H_1$) which is shown in figure 7*B*. As expected, evaporation minus precipitation depends strongly also on the precipitable water.

Finally Figs. 7*C* and 7*D* are, respectively, the differences (February 1964 minus February 1962) of the surface temperature and the com-

Fig. 6. Difference, February 1964 minus February 1962, of the computed values of evaporation minus precipitation, in ly per day.

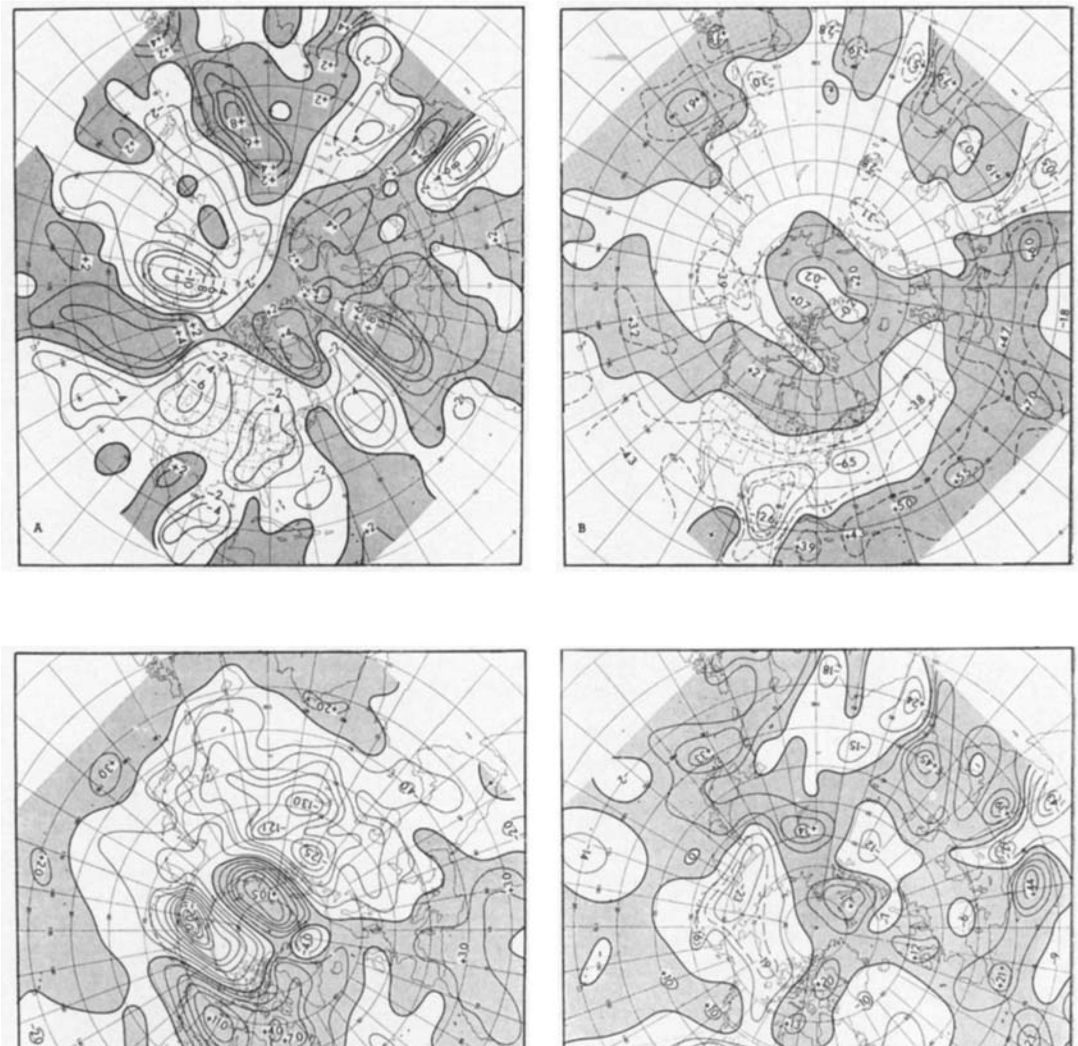

Fig. 7. Values of the difference, February 1964 minus February 1962, of the variables on which evaporation minus precipitation mainly depends: (*A*) Meridional wind component at 1000 mb, in meters per sec; (*B*) precipitable water, in gm cm^{-2}; (*C*) surface temperature in °C; and (*D*) total cloudiness, in percent.

puted cloudiness which are the factors on which the precipitable water (Fig. 7*B*) mainly depends, and which are variables that enter in the computation of the difference, evaporation minus precipitation.

6. Concluding remarks

The numerical computations have shown that formula (11) is a good parameterization of the difference, evaporation minus precipitation

The formula was applied to monthly averages; however, it is evident from its derivation that it may be possible to apply it to shorter or longer periods than a month.

Improvements in the formula or adjustments in its applicability to a particular time average, can be obtained by improving the vertical distribution of relative humidity, wind and temperature, as well as the relationship between cloudiness and relative humidity.

The most difficult component to be incorporated in the formula is the contribution due to the flux divergence of the transient eddies. Unless we obtain a realistic parameterization in terms of the time-averaged variables, the best possibility is to assume that it is equal to zero. This assumption, suggested by the discussion given in section 2, seems to be supported by the relative success of the computations.

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ПАРАМЕТРИЧЕСКИЙ МЕТОД ВЫЧИСЛЕНИЯ СРЕДНЕГО БАЛАНСА ВОДЫ В АТМОСФЕРЕ

Путем использования уравнения сохранения водяного пара в тропосфере разность между испарением с поверхности и осадками выражается в виде функции общей облачности (или средней тропосферной относительной влажности) и температуры и горизонтального ветра у поверхности и в средней тропосфере.

Вычисления годовых нормальных величин разности испарения и осадков с использованием окончательных формул вместе с среднемесячными величинами необходомых параметров для Северного полушария хорошо согласуются в основных чертах с оценками Старра, Пексото и Кризи, также как с климатологическими величинами. Однако существуют мелкомасштабные расхождения в этих трех оценках.

Вычисления сильно зависят от дивергенции горизонтального ветра, и некоторые расхождения могут быть объяснены ошибками в определении последнего.

Для определения важности аномалий разности испарения и осадков проделаны вычисления для февраля 1964 г. и февраля 1962 г., которые показывают, что величины разности между этими двумя месяцами того же порядка величины, что и сами месячные значения.

Если положительные значения указывают на ветер с юга на север, величины разности испарения и осадков обнаруживает обратную корреляцию с меридиональной компонентной ветра в широком диапазоне. Кроме того, вычисления показывают, что аномалии разности испарения и осадков строго зависят не только от аномалий меридиональной компоненты ветра, но и от аномалий воды в осадках, которая, в свою очередь зависит от аномалий поверхностной температуры и облачности /или средней тропосферной относительной влажности/.