# The Release of Latent Heat of Condensation in a Simple Precipitation Forecast Model[1]

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ABSTRACT

A method is proposed for deriving a useful quantitative precipitation forecast from a physical model in which no account is taken of the release of latent heat of condensation. The approach is to calculate, on the basis of a simple theory, the ratio between the large-scale updraft speed when the latent heat is incorporated and the smaller updraft speed when it is not. The resulting ratio is then assumed to be equivalent to the ratio of the storm-average precipitation to the comparable smaller average computed from the thermodynamically dry model. An empirical test supports the assumption reasonably well for large winter storms in the central and eastern United States. For operational purposes it is suggested that the theoretical ratio be estimated statistically from the mean temperature of the layer from 1000 mb to 500 mb. Evidence is provided that this estimate can be made fairly successfully. Finally, some inferences are drawn concerning the role of cumulus convection in this type of storm.

## 1. Introduction

An objective procedure for prediction of large-scale cloudiness and precipitation, based on a physical model put forth by Younkin *et al.* (1965) has been in recent operational use at the National Meteorological Center. This model (hereinafter called the SLYH[3] model) uses as input, in addition to the initial moisture field, operational prognoses at 500 and at 1000 mb. The thermodynamics of these prognostic circulation models are assumed to be adiabatic, so that no account is taken of the release of latent heat of condensation of water vapor. The SLYH model, in fact, predicts supersaturation, which is removed at the end of each 12-hr interval and is regarded as a quantitative precipitation forecast.

According to Saylor,[4] the model has been strikingly successful in prediction of the areal extent of large-scale precipitation out to 48 hr in advance over level ground, but the predicted amounts are substantially smaller than the observed amounts. A primary reason for this deficiency is doubtless the neglect of release of latent heat. We may expect *a priori* that this neglect will result in an underestimate of the intensity of the updraft in rainstorms.

In this paper we shall propose a method for estimating the enhancement of the updraft due to this source of diabatic heating. On the assumption that, other things being equal, the condensation rate is proportional to the updraft speed and similarly that the precipitation rate is proportional to the condensation rate, we then have a method for making a realistic quantitative precipitation forecast based upon the "thermodynamically dry" SLYH prediction. The method is tested on a sample of the 16 largest rainstorms in the central and eastern United States from December 1964 through April 1965.

## 2. Theoretical development

The SLYH model prediction equation, the derivation of which is discussed in detail by Younkin *et al.* (1965), is the conservation equation

$$\frac{\partial}{\partial t}(h_d-2h_5)=-\mathbf{V}_*\cdot\nabla(h_d-2h_5), \tag{1}$$

where $h_d$ is the saturation deficit, a measure of the mean relative humidity of the layer from 1000 to 500 mb, $h_5$ is the thickness, which is related to the mean temperature of the layer, and $\mathbf{V}_*$ is the "moisture-steering wind," a linear combination of the winds at the bottom and top of the layer. The coefficients of these winds and the coefficient "2" in (1) are determined from appropriate climatological vertical profiles of horizontal wind, specific humidity and stability, and from an assumed quasi-linear vertical profile of horizontal divergence. The vertical motions do not appear explicitly but they are implicit in the changes of $h_5$ experienced along the trajectory in the moisture-steering flow. When the predicted values of $h_d$ are negative the thermo-

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[4] Personal communication.

dynamically dry quantitative precipitation forecast is obtained from Table 5 in Younkin *et al.* (1965).

We shall now examine in detail the problem of incorporating the effect of release of latent heat. We assume that this heating does not affect the area of the precipitation forecast. This assumption is supported by operational experience with the SLYH model and by computations by Danard (1964), who found that only a slight shrinkage of the rainfall area results from incorporation of latent heat release.

The approach is suggested by the work of Sumner (1950) in a somewhat different context. We start with an appropriate form of the first law of thermodynamics,

$$\frac{d\theta}{dt}=\frac{\theta}{c_pT}\frac{dQ}{dt}=\frac{\partial\theta}{\partial t}+\mathbf{V}\cdot\nabla\theta+\omega\frac{\partial\theta}{\partial p},\tag{2}$$

where $c_p$ is the specific heat of air at constant pressure, $\theta$ potential temperature, $dQ/dt$ the rate of diabatic heating, and $\omega\equiv dp/dt$ will be regarded as the vertical motion. In the case of heating due to the release of latent heat of condensation,

$$\frac{dQ}{dt}=L\left(\frac{dq_s}{dp}\right)_{ma}\omega,\tag{3}$$

where $L$ is the latent heat of condensation and $(dq_s/dp)_{ma}$ is the rate of change of saturation specific humidity with pressure along a moist adiabat. We substitute (3) into (2), divide the result by $\theta$, and apply the equation of state, to obtain

$$\frac{\partial}{\partial p}\left(\frac{\partial\Phi}{\partial t}\right)=-\mathbf{V}\cdot\nabla\left(\frac{\partial\Phi}{\partial p}\right)-\omega\left[\frac{\partial\ln\theta}{\partial p}\frac{\partial\Phi}{\partial p}-\frac{RL}{c_pp}\left(\frac{dq_s}{dp}\right)_{ma}\right],\tag{4}$$

where $\Phi$ is the geopotential and $R$ is the gas constant for dry air. When (4) is combined with the vorticity equation for quasi-geostrophic motion so as to eliminate the geopotential tendency, the diagnostic equation for vertical motion is

$$\left\{\left[\sigma_d-\frac{RL}{c_pp}\left(\frac{dq_s}{dp}\right)_{ma}\right]\nabla^2+f_0^2\frac{\partial^2}{\partial p^2}\right\}\omega_m=-f_0\frac{\partial}{\partial p}(-\mathbf{V}\cdot\nabla\eta)+\nabla^2\left(-\mathbf{V}\cdot\nabla\frac{\partial\Phi}{\partial p}\right),\tag{5}$$

where $\sigma_d\equiv(\partial\ln\theta/\partial p)(\partial\Phi/\partial p)$, $\nabla^2$ is the horizontal Laplacian, $f_0$ is a constant value of the Coriolis parameter, and $\eta$ is the geostrophic absolute vorticity. It is convenient to denote $\omega_m$ the "moist" vertical motion. For the dry case we obtain the "dry" vertical motion $\omega_d$ from an equation which is identical to (5) except that the coefficient of the $\nabla^2$-operator is simply $\sigma_d$. For simplicity, denote the right side of (5) by $F$ and define $\sigma_m\equiv\sigma_d-RL(dq_s/dp)_{ma}/c_pp$. Thus, we account for release of latent heat by adjusting the effective stability. For a moist-adiabatic lapse rate $\sigma_m=0$, just as $\sigma_d=0$ for a dry-adiabatic lapse rate.

Now we assume that the field of vertical motion at a given time can be adequately represented by the simple harmonic function

$$\omega(x,y,p)=\Omega\sin\frac{2\pi x}{A}\sin\frac{2\pi y}{B}\sin\frac{2\pi p}{C},\tag{6}$$

where $\Omega$ is a constant and $A$, $B$, and $C$ are wavelengths along the $x$, $y$, $p$ axes. (The verisimilitude of the final result will be a test of the realism of this assumption, among other things.) Then

$$\nabla^2\omega=-\left[\left(\frac{2\pi}{A}\right)^2+\left(\frac{2\pi}{B}\right)^2\right]\omega,\tag{7}$$

and

$$\frac{\partial^2\omega}{\partial p^2}=-\left(\frac{2\pi}{C}\right)^2\omega.\tag{8}$$

Substitution of these relationships in (5) then gives

$$\omega_{a,m}=-\frac{F}{\left(\frac{2\pi f_0}{C}\right)^2\left[1+\sigma_{d,m}\frac{\left(\frac{1}{A^2}+\frac{1}{B^2}\right)}{(f_0/C)^2}\right]}.\tag{9}$$

The desired ratio between the moist and dry vertical motions is then

$$\frac{\omega_m}{\omega_d}=\frac{1+\dfrac{\sigma_d\left(\frac{1}{A^2}+\frac{1}{B^2}\right)C^2}{f_0^2}}{1+\dfrac{\sigma_m\left(\frac{1}{A^2}+\frac{1}{B^2}\right)C^2}{f_0^2}}.\tag{10}$$

This ratio, or "enhancement factor," is thus seen to depend upon the horizontal and vertical scales, as defined by $A$, $B$ and $C$, and upon the moist and dry stabilities, $\sigma_m$ and $\sigma_d$, which depend in turn principally upon the static stability $(\partial\theta/\partial p)$ and the moisture content $(dq_s/dp)_{ma}$.

The relative importance of the various factors in determining the enhancement factor is illustrated by some sample calculations based on (10). To obtain a realistic idea of the stability and moisture distributions characteristic of heavy large-scale rainstorm situations, 90 soundings were selected at random from times and places at which precipitation was occurring during the selected sample of 16 storms during the 1964–65 winter season in the central and eastern United States. The average of these data is displayed in Fig. 1. Interestingly, in view of the oft-held belief that such situations are characterized by nearly moist-adiabatic conditions, the average sounding in Fig. 1 is distinctly more stable than the moist-adiabatic lapse rate.

To test the effect of horizontal scale, (10) was evaluated for this mean sounding ($\sigma_d = 2.5\times10^{-2}$ m$^2$ sec$^{-2}$ mb$^{-2}$, $\sigma_m = 0.8\times10^{-2}$ m$^2$ sec$^{-2}$ mb$^{-2}$) and for a moist-adiabatic lapse rate with temperature 293K at the 1000-mb level ($\sigma_d = 2.0\times10^{-2}$ m$^2$ sec$^{-2}$ mb$^{-2}$, $\sigma_m = 0$). Results are given in Table 1. It is clear from (10) that as the scale becomes smaller, the enhancement factor $\omega_m/\omega_d$ will approach $\sigma_d/\sigma_m$. For the moist-adiabatic lapse rate, the factor approaches infinity as the scale of the storm approaches zero, and Table 1 shows that it is highly sensitive to scale in the range from 4000 km to 200 km. In the mean sounding, on the other hand, the limiting ratio is 3.125, and Table 1 indicates relatively little sensitivity to scale, with the limit nearly reached at a scale of 400 km.

## 3. An empirical test

As a test of (10) a study was made of the 16 largest rainstorms in the central and eastern United States from December 1964 through April 1965. Unavailability of SLYH forecasts precluded use of cases in October and November 1964, and no cases were selected during summer because is was felt that these storms would prove to be essentially convective and thus would be unsuitable for consideration.

For each case the facsimile chart of 24-hr rainfall amounts for the period ending at 1200 GMT was analyzed. Results are shown in Fig. 2. An average observed precipitation depth for each storm was obtained by averaging the values read from the analysis at the NMC grid points in the storm area. In view of the wide spacing of the points (approximately 400 km) relative to the scale of important detail in the observed data, some lack of confidence was felt in the averaged values. Accordingly, for two of the cases (randomly selected) a more refined average was obtained from a grid of points spaced at one-half the NMC distance. For the first case the average depth for the NMC grid was 0.57 inch while for the finer grid the value was 0.585 inch. For the second the values were 0.65 and 0.67. This result was felt to provide justification for employing the NMC grid for further work with most of these large winter storms. The reduced mesh length

Table 1. Enhancement factor $\omega_m/\omega_d$ for a variable horizontal scale and two different lapse rates. (See text for values.)

| Scale = $A$ = $B$ (km) | Moist adiabatic | Mean sounding |
|---|---|---|
| 4000 | 2.0 | 1.6 |
| 2000 | 5.0 | 2.3 |
| 1000 | 17.0 | 2.8 |
| 400 | 101.0 | 3.1 |
| 200 | 401.0 | 3.1 |

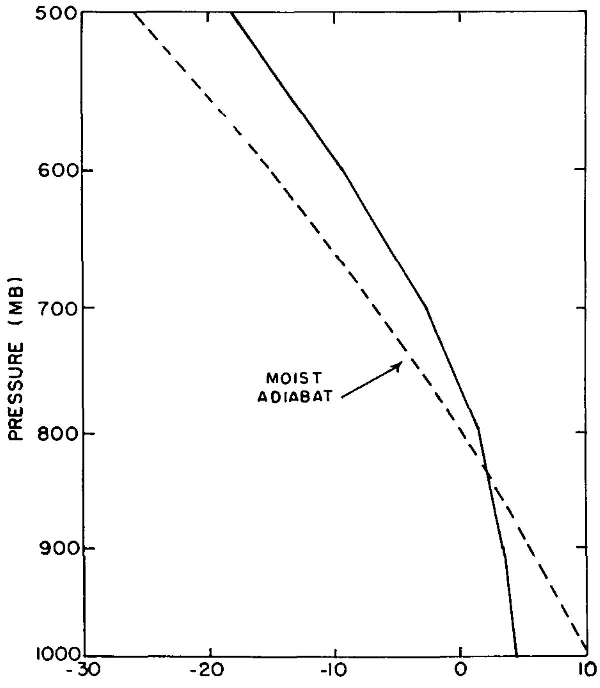


Fig. 1. Mean lapse rate of temperature for 90 randomly selected soundings during large-scale winter rainstorms in the central and eastern United States during the winter season 1964–65.

was used, however, in some of the cases in which the area of the rainstorm was relatively small. Nevertheless, the results must be accepted with some reservation in cases where the precipitation is particularly variable; in such cases, in fact, it is doubtful whether all the raingage data themselves contain sufficient information to provide an entirely reliable estimate of the mean.

To obtain the dry-predicted 24-hr precipitation, the first 12 hr of two consecutive SLYH forecasts were used, one made from observed initial data at the beginning of the period and the other from observed data halfway through the period. For each SLYH forecast the area of predicted negative saturation deficit was identified. At each NMC grid point within the area, this value was used together with the observed thickness and Table 5 from Younkin *et al.* (1965) to obtain a predicted precipitation amount. The contributions from the two 12-hr periods were then added to yield the dry-predicted 24-hr rainstorm. Results are shown in Fig. 2.

Next, for each rainstorm, the predicted and observed rainfall was obtained from an average of the grid point values over an area representing the envelope of the observed and predicted rainstorms. When a point fell in one but not in both rainstorms the value zero was used for the other. If many such points are included, the two averages and probably their ratio are spuriously small; in the present instance, however, the predicted precipitation areas seem sufficiently accurate that this source of error in the ratio is almost certainly smaller

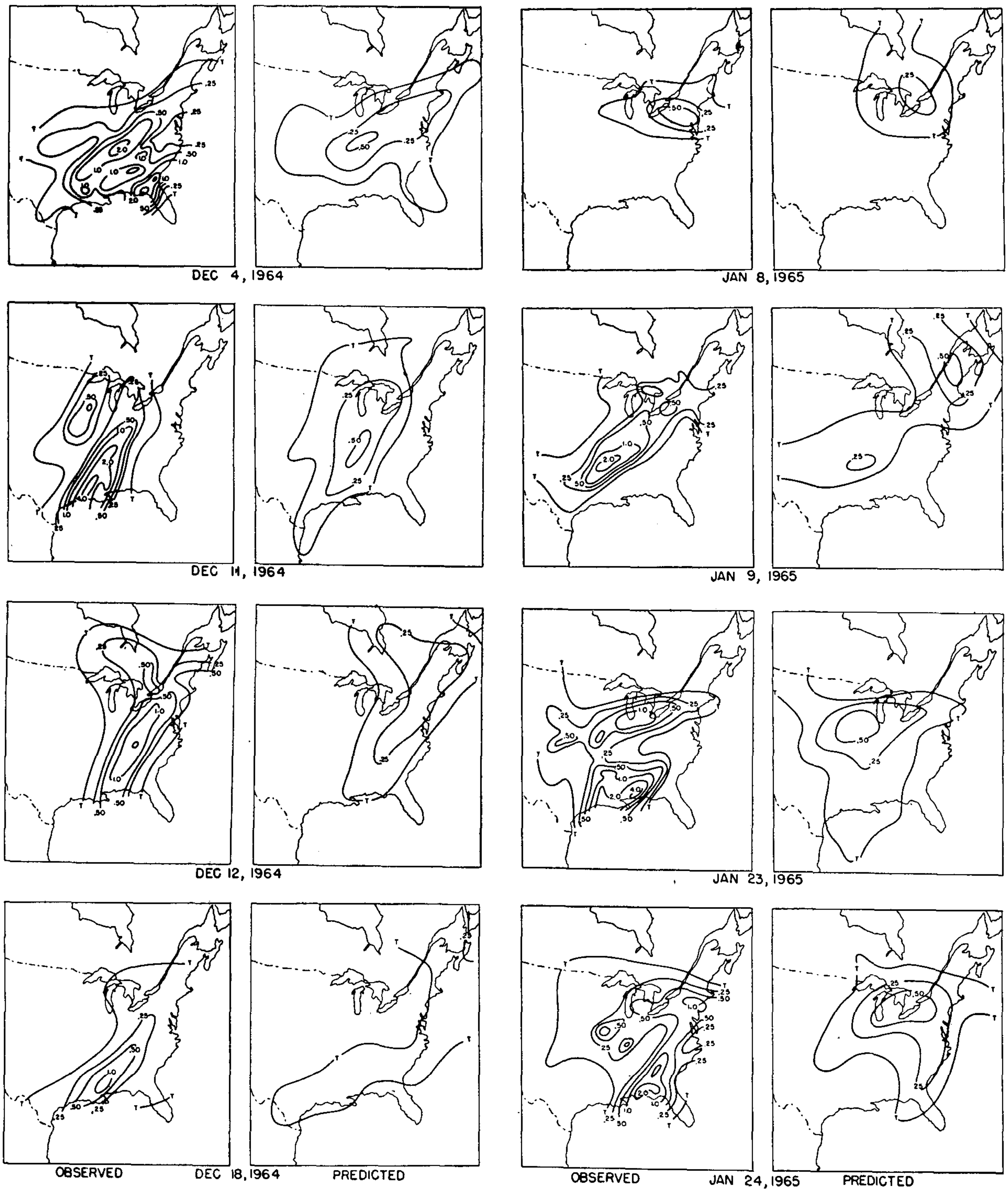


FIG. 2. Twenty-four hour precipitation amounts for periods ending at 1200 GMT of the date indicated. In each pair of maps the observed pattern is to the left and the dry SLYH composite forecast is to the right. Isohyets are drawn for trace, 0.25 inch, 0.50 inch, 1.00 inch and 1-inch intervals thereafter.

than other uncertainties. The average predicted and observed values, and their ratio $\bar{P}_0/\bar{P}_f$, are given for each storm in Table 2. Each 24-hr storm period ended at 1200 GMT of the date given in the table. The ratios range from 1.9 to 3.6, except for the cases of 18 December and 8 January. As seen in Fig. 2, the former presents an instance of heavy observed precipitation centered near the Gulf Coast downwind from an area in which

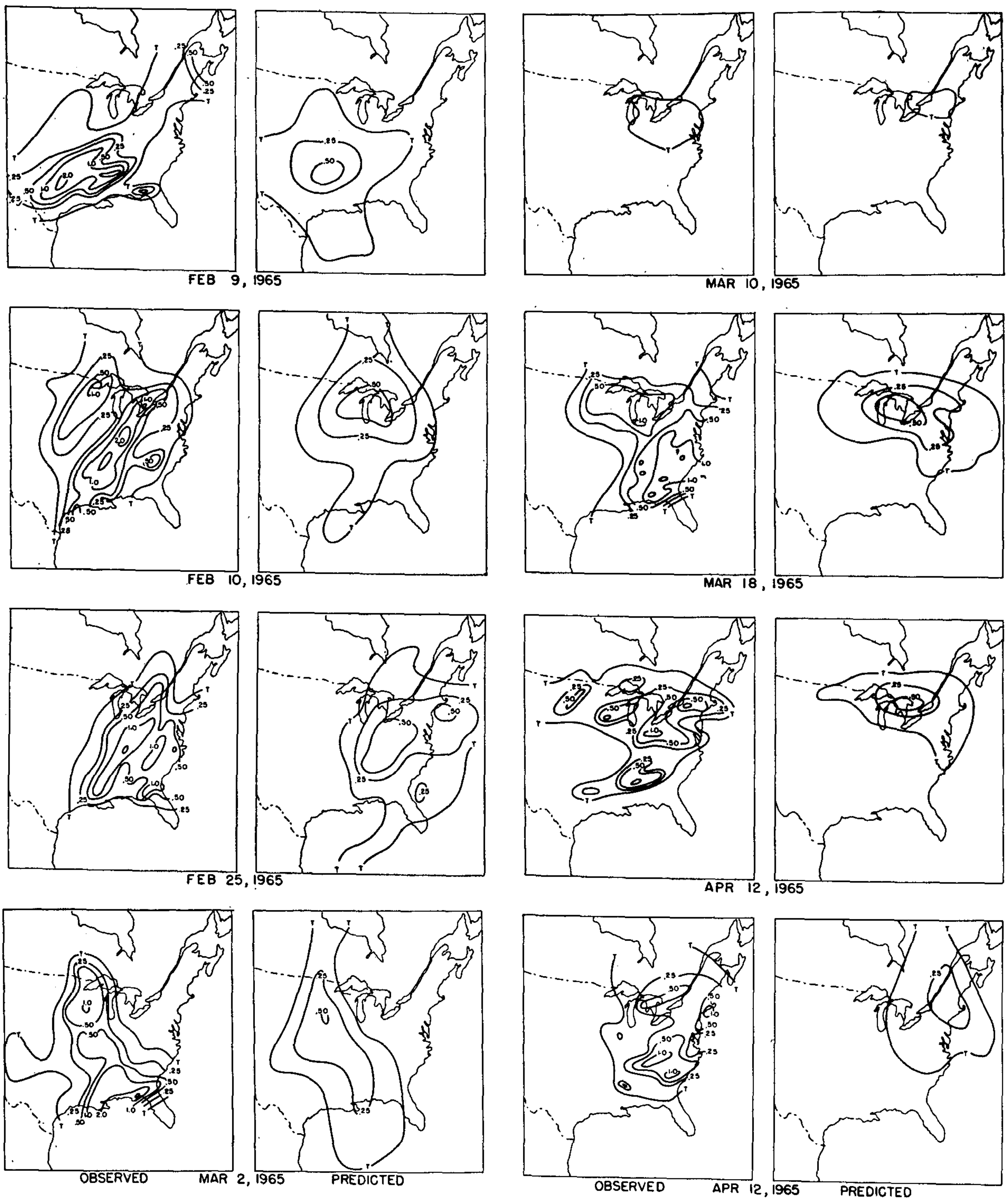


FIG. 2. (Continued).

the initial moisture information may have been seriously deficient. The latter, by contrast, is an instance of substantial predicted precipitation north of the Great Lakes which failed to occur at all. Since both of these forecasts likely stemmed from errors other than the thermodynamic deficiency of the SLYH model under examination here, they are excluded from further consideration.

The specific object of the test is to compare the above empirical ratio between mean observed precipitation

TABLE 2. Observed and dry predicted precipitation amounts, moist stabilities, and theoretical enhancement factors for selected major storms, 1964–65 winter season.

| Case | $\bar{P}_f$ (inches) | $\bar{P}_0$ (inches) | $\bar{P}_0/\bar{P}_f$ | $\sigma_m(10^{-2}\text{m}^2\ \text{sec}^{-2}\text{mb}^{-2})$ | $\omega_m/\omega_d$ | $\bar{\sigma}_m$(adj.) $(10^{-2}\text{m}^2\text{sec}^{-2}\text{mb}^{-2})$ | $\omega_m/\omega_d$(adj.) |
|---|---|---|---|---|---|---|---|
| Dec. 4 | 0.15 | 0.50 | 3.3 | 0.60 | 2.94 | 0.50 | 3.37 |
| Dec. 11 | 0.19 | 0.56 | 2.9 | 0.76 | 2.43 | 0.76 | 2.43 |
| Dec. 12 | 0.16 | 0.30 | 1.9 | 0.90 | 2.12 | 0.90 | 2.12 |
| Dec. 18 | 0.03 | 0.18 | 6.0 | 0.68 | 2.91 | — | — |
| Jan. 8 | 0.11 | 0.10 | 0.9 | 0.73 | 2.60 | — | — |
| Jan. 9 | 0.12 | 0.30 | 2.5 | 0.88 | 2.25 | 0.73 | 2.66 |
| Jan. 23 | 0.19 | 0.51 | 2.7 | 0.81 | 2.25 | 0.73 | 2.66 |
| Jan. 24 | 0.19 | 0.40 | 2.1 | 1.25 | 1.56 | 1.00 | 1.91 |
| Feb. 9 | 0.14 | 0.29 | 2.1 | 0.79 | 2.36 | 0.79 | 2.36 |
| Feb. 10 | 0.19 | 0.47 | 2.5 | 0.84 | 2.27 | 0.76 | 2.66 |
| Feb. 25 | 0.23 | 0.43 | 1.9 | 0.90 | 2.06 | 0.80 | 2.27 |
| Mar. 2 | 0.15 | 0.45 | 3.0 | 0.72 | 2.71 | 0.60 | 3.17 |
| Mar. 10 | 0.027 | 0.071 | 2.6 | 0.78 | 2.53 | 0.78 | 2.53 |
| Mar. 18 | 0.14 | 0.51 | 3.6 | 0.78 | 2.36 | 0.56 | 3.15 |
| Apr. 12 | 0.09 | 0.27 | 3.0 | 0.54 | 3.39 | 0.54 | 3.39 |
| Apr. 16 | 0.10 | 0.30 | 3.0 | 0.68 | 2.69 | 0.68 | 2.69 |
| Average (excluding 18 Dec. and 8 Jan.) | | | 2.65 | 0.80 | 2.42 | 0.72 | 2.67 |

and dry predicted precipitation with an appropriate theoretical ratio computed from (10) between moist and dry vertical motions. To obtain this latter ratio it is necessary to measure the scale of the vertical motion system and the moist and dry stabilities for each case.

Measurement of the scale was accomplished by shading the area of current precipitation on each surface chart during the storm period and then subjectively fitting an ellipse to the area. The major and minor axes were then taken as half the wavelength in the two directions (probably a fair approximation for major rainstorms).

The Reed and SLYH forecasts use implicitly a mean winter value of $\partial\theta/\partial p$, from which we obtain approximately $\sigma_d=2.0\times10^{-2}$ m$^2$ sec$^{-2}$ mb$^{-2}$. For each case, the moist stability, $\sigma_m\equiv(\partial\ln\theta/\partial p)(\partial\Phi/\partial p)-RL(dq_s/dp)_{ma}/c_pp$, was obtained for each sounding taken during precipitation. Values of $\partial\theta/\partial p$ and $(dq_s/dp)_{ma}$ were approximated by finite differences taken between the 900-mb and 500-mb levels and 700 mb was used as $p$. Neglect of the layer below 900 mb seemed advisable because of the probable smallness of condensation rate near the ground. Only about 5 per cent of the soundings disclosed conditional instability, leading to negative values of $\sigma_m$. An average value was about $0.8\times10^{-2}$ m$^2$ sec$^{-2}$ mb$^{-2}$, except where frontal inversions extended above the 900-mb level, or in very cold air, where values near $1.5\times10^{-2}$ m$^2$ sec$^{-2}$ mb$^{-2}$ were the rule.

An average value of $\sigma_m$ was assigned to each rainstorm by computing a simple average of the values for each sounding in the precipitation area. Results are given in Table 2.

From the foregoing data the ratio $\omega_m/\omega_d$, as computed for each storm from (10), is also listed in Table 2. Since the theoretical value appears to be highly correlated with the storm-averaged moist stability, these storms are evidently large enough and of a sufficiently uniform size that variation of horizontal scale does not play an especially important role. A seasonal trend is evident. Note from the data in the table that the relatively small moist stabilities, due to the moisture-rich air masses present at the beginning and end of the winter season, produced relatively large theoretical and empirical ratios at these times. In midwinter, moist stabilities tend to be at a maximum and the ratios at a minimum. Finally, on the average but not in all instances, the theoretical ratio is smaller than the empirical ratio.

With reference to this last discrepancy, in a number of cases pronounced frontal inversions were present between 900 and 880 mb (but rarely above). On the assumption that the condensation was occurring mainly in the warm air, a more representative lapse rate was calculated from the top of the inversion to 500 mb for all affected soundings. A corresponding adjusted moist stability was thus obtained for each of those storms in which pronounced frontal structure was present. The resulting adjusted values of $\omega_m/\omega_d$, shown in Table 2, were in better agreement with the empirical ratio in all but one case than were the unadjusted theoretical values. The average of the theoretical ratios, adjusted where necessary, is in fact very close to the average of the empirical ratios between actual and dry SLYH predicted precipitation.

A linear regression analysis of $\bar{P}_0/\bar{P}_f$ and $\omega_m/\omega_d$ for the 14 storms remaining after elimination of the cases of 18 December and 8 January yielded a correlation coefficient of 0.83 and the regression equation

$$\bar{P}_0/\bar{P}_f=1.39(\omega_m/\omega_d)-0.92. \tag{11}$$

Visual inspection of the points, plotted in Fig. 3, suggests that the regression coefficients are not significantly different from one and zero, respectively, which would represent an optimum agreement between theory and observation.

## 4. A suggested operational procedure

To apply these results operationally the moist stability must be predicted, but to do this in a straightforward way would require models more complex than those which have been operationally popular. Since the thickness (or mean temperature) of the layer from 1000 to 500 mb is predicted by the current NMC models, however, we may avail ourselves of a presumed statistical relationship between this thickness and $\sigma_m$. We should expect that large values of the thickness, representing relatively warm air, would be associated in a precipitation area with relatively large lapse rates and large moisture contents, and thus with small values of $\sigma_m$.

The grid points used in obtaining the storm averages discussed above were then stratified according to 60-m ranges of thickness. For all the grid points in each thickness range, average values were obtained for $P_0/P_f$, adjusted $\sigma_m$ and $\omega_m/\omega_d$. For the last of these quantities, $\sigma_d$ was taken as $2.0\times10^{-2}$ m$^2$ sec$^{-2}$ mb$^{-2}$, $f_0$ as $10^{-4}$ sec$^{-1}$, $C$ as 2000 mb, and the updraft was assumed to cover an elliptical horizontal area with axes A and B equal to 750 and 1500 km, respectively.

Results are listed in Table 3 and plotted in Fig. 4, in which curves estimated by eye as a best fit are added. There is a reasonably close resemblance between the theoretical and empirical data through the middle range of thickness, where the ratios are between approximately 1.5 and 2.0. The scatter of the empirical data near both extremes of thickness is due in part to the small number of observations on which these averages are based. With extremely cold air, however, the empirical ratio is systematically larger than the theoretical ratio. This discrepancy is probably attributable to enhancement of the observed precipitation by heavy low-level convective snow showers in the lee of the Great Lakes, a process not represented at all in the SLYH forecasts. At the highest thickness both the theoretical and empirical ratios evidently rise sharply as the moist stability becomes small.

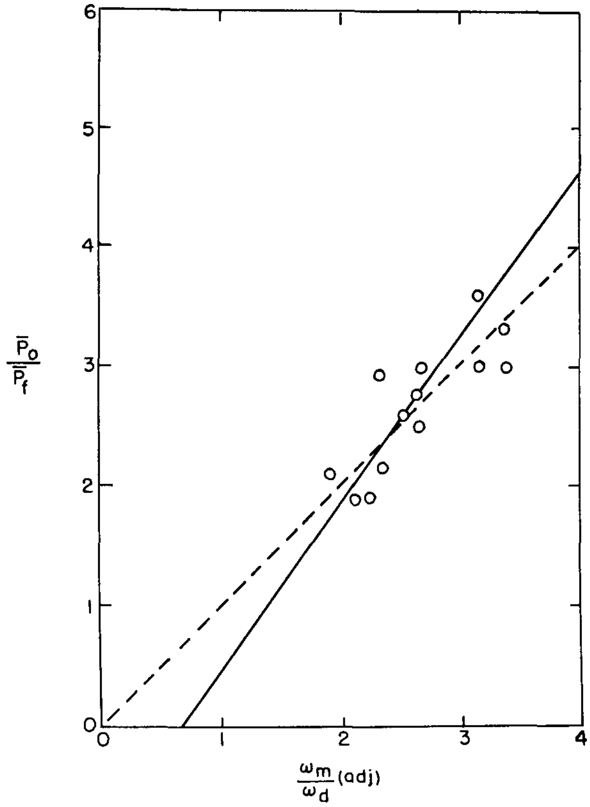


Fig. 3. Scatter diagram of observed ($\bar{P}_0/\bar{P}_f$) vs. theoretical ($\omega_m/\omega_d$) ratios for 14 major storms December 1964–April 1965. The solid line is the calculated regression line. The dashed line represents optimum agreement between observation and theory.

## 5. Comments on the role of cumulus convection

A question of considerable current interest in meteorology is the interactive role of cumulus convection within the circulations of larger scale. In the present context

Table 3. Observed and dry-predicted precipitation amounts, moist stabilities, and theoretical enhancement factors for grid points stratified by thickness, winter season 1964–65.

| Thickness (m) | $P_f$ (inches) | $P_0$ (inches) | $P_0/P_f$ | $\sigma_m$(adj.) ($10^{-2}$m$^2$sec$^{-2}$mb$^{-2}$) | $\omega_m/\omega_d$(adj.) | Number of points |
|---|---|---|---|---|---|---|
| 5040 | 0.02 | 0.07 | 3.5 | — | — | 3 |
| 5100 | 0.06 | 0.15 | 2.5 | 1.45 | 1.4 | 2 |
| 5160 | 0.02 | 0.11 | 5.5 | 1.56 | 1.3 | 8 |
| 5220 | 0.08 | 0.19 | 2.3 | 1.05 | 1.8 | 23 |
| 5280 | 0.07 | 0.12 | 1.7 | 1.22 | 1.6 | 44 |
| 5340 | 0.15 | 0.13 | 0.9 | 1.22 | 1.6 | 63 |
| 5400 | 0.17 | 0.24 | 1.4 | 1.09 | 1.8 | 88 |
| 5460 | 0.21 | 0.33 | 1.6 | 0.87 | 2.15 | 85 |
| 5520 | 0.22 | 0.49 | 2.2 | 0.63 | 2.9 | 73 |
| 5580 | 0.16 | 0.59 | 3.7 | 0.51 | 3.4 | 94 |
| 5640 | 0.08 | 0.76 | 9.5 | 0.22 | 6.4 | 80 |
| 5700 | 0.04 | 0.21 | 5.2 | — | — | 21 |

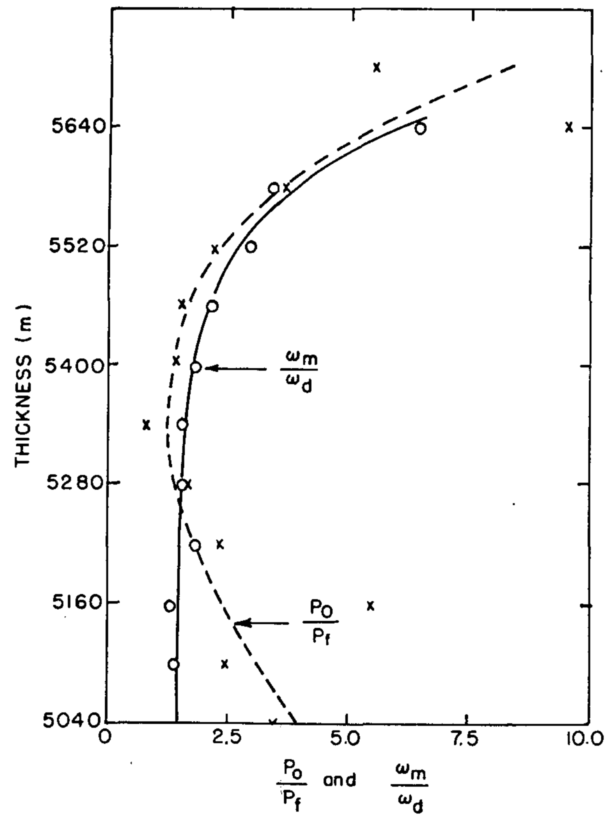


FIG. 4. $P_0/P_f$ and $\omega_m/\omega_d$ as a function of thickness of the layer from 1000 to 500 mb. The circles and the solid curve refer to $\omega_m/\omega_d$, the crosses and the dashed line to $P_0/P_f$.

this question may be specialized by asking whether the effect of cumulus activity embedded within the larger storm is to add to the total precipitation of the storm or simply to redistribute the precipitation within the storm area. The SLYH model is unaware of even the existence of such a thing as a cumulus cloud. Yet aside from the cold-air snow flurries mentioned above (which are relatively superficial though perhaps locally disastrous), the present results give little evidence of a deficiency in predicted precipitation in major winter storms over level terrain once the large-scale release of latent heat is taken into account. We are thus led, though not conclusively, to the finding that here the role of the embedded cumulus is distributive.

On the other hand it is surely not difficult to find summertime situations in which squall lines produce abundant precipitation in circumstances where a large-scale prognosis would have yielded none. Here the cumulus are obviously of crucial importance. Furthermore, an intermediate concept seems to prevail in contemporary thinking about hurricanes as exemplified by Kuo (1965). That is, the specific effects of cumulus are important to the larger circulation but can be adequately simulated statistically. It would be strange indeed if all three viewpoints were correct within their respective contexts, and if the importance of cumulus effects were thus in some sense proportional to the atmosphere's need for them. We leave this broader question as a puzzle for the reader.

## 6. Conclusions

We find that the quantitative deficiency in the SLYH precipitation forecasts for large winter storms can be removed, in effect, by replacing the computed large-scale thermodynamically dry updraft by a more intense one in which the effect of release of latent heat of condensation is taken into account. It does not appear to be necessary to consider the effects of cumulus convection either specifically or statistically except in cold-air snow flurries over the Great Lakes. This conclusion is tentative, however, because of the limited number of cases available for testing. In any event the cumulus must be dealt with if the details of the pattern are to be satisfactorily predicted.

We find that the ratio between the two updrafts for this type of major winter storm can be determined operationally by making simple reference to the thickness (or mean temperature) of the layer from 1000 to 500 mb.

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