The relative importance of variables in initial conditions for dynamical weather prediction[1]

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(Manuscript received December 17, 1969)

ABSTRACT

Tests were made to evaluate the relative importance of large errors in some of the variables formally required to specify initial conditions for the time integration of the hydro-thermodynamic equations for the atmospheric motion. The prime purpose was to analyze the adjustment process immediately following the initial disturbance. The variables selected were the surface pressure, the water vapor, the boundary layer temperature, and the boundary layer wind. It is likely that these quantities are not individually nor possibly even collectively essential to the definition of initial conditions because of dynamic coupling. The experiments showed that the predictions appear almost the same after about six hours irrespective of the initial values of these quantities. But after several days, the subsequent error behaves similar to the growth noted in predictability experiments.

1. Introduction

Several years ago, there was discussion on the feasibility of remotely measuring surface atmospheric pressure from a weather satellite, a difficult, if not impossible task. At that time it raised the question as to whether a knowledge of non-orographic surface pressure variations is essential input for a prediction computation.

It has been customary to regard the surface pressure as a fundamental quantity in characterizing weather conditions and that it is therefore an indispensable variable for initial conditions. But its necessity is not obvious. There is a possibility that the surface pressure could be derived from other variables within a certain tolerable accuracy. This suspicion was one of the motivations for initiating the series of the experiments to be described here.

Y. Mintz & T. N. Krishnamurti (personal communication), based on the original ideas of Mintz (1963), conducted an interesting experiment concerning the effect of the initial divergent component of wind on the subsequent evolution of the overall atmospheric motion. They interrupted a long-term general circulation integration, using a global, primitive equation model. Next, they took the geopotential heights at a particular time from the data of the experiment and computed the geostrophic wind with a variable and a constant coriolis parameter, and also determined the linearly balanced wind. Using each of these wind fields as initial conditions, numerical integrations were made with the same model. They found that there was remarkably close agreement between the original experiment and each of the three integrations with the modified initial winds for about five days. This was noted in the behavior of the surface pressure as well as in that of the upper level wind and temperature fields. However, the solutions began to deviate from each other after five days.

Smagorinsky *et al.* (1967) had a similar experience. They carried out two marching computations with a moist general circulation model in which the initial vertical velocities were different. For one, the vertical velocity was obtained from the solution of the ω-equation, and in the other it was zero. In both cases the precipitation computed in the experiments was quite similar in horizontal distribution and in magnitude, except for the first six hours, even

[1] Some of these results were presented at the WMO/IUGG Symposium on Numerical Weather Prediction, Tokyo, 1968, under the title, "The relative importance of variables in initial conditions for numerical predictions", by J. Smagorinsky and K. Miyakoda.

though precipitation amount would be expected to be a very sensitive measure.

It thus appears that of the dependent variables formally necessary to specify the large-scale state of the atmosphere, some are more fundamental than others. As we know, they are all physically coupled but some may have greater dynamical inertia than others which will be reflected in their relative inherent information content. The above cited example suggests that the vertical velocity is secondary in an ordering of importance—a fact that has been appreciated as a dynamical characteristic of the large-scale atmospheric motions. As will be seen, this motion can be generalized to include a spatial ordering as well as an ordering of variables—an obvious consequence of the fact that the dynamical and energetic coupling in the extra-tropical troposphere and the equatorial troposphere and the boundary layer are essentially different. Subsequent studies (to be published later) even show distinct and useful differences between data in the troposphere and stratosphere and even between the upper and lower free troposphere, as one might suspect. The question then is which quantities are dominant and which are secondary and, therefore, redundant.

The objective of this work is to begin to identify the redundant variables. Ultimately, it will be our objective to order their degree of redundancy as a basis for assessing information content and its impact in designing observational systems. In a sense, it is to examine the nature of the variable's adjustment capability within a given system of equations, and to investigate how it responds to baroclinic instability. The adjustment of mass and motion fields has been extensively studied for barotropic fluids by Rossby (1938), Cahn (1945), Obukhov (1949), Raetjen (1950), Washington (1964), and for baroclinic fluids by Bolin (1953), Kibel (1955), Veronis (1956), and Monin & Obukhov (1959).

Relation to predictability decay

In actuality, the dynamic predictability experiments (National Academy of Sciences, 1966; Smagorinsky, 1969) were also examples of the adjustment process. Although the initial vertical temperature errors in those experiments were small, they did in fact upset the thermal wind balance and one did observe a decrease in the temperature error within the first day or two, while the wind field, which was initially undisturbed, monotonically degraded. To the writers' knowledge, however, the questions as to which quantities adjust more easily and which quantities are instrumental in subsequent error growth have not been discussed.

In the predictability studies, a *control run* is compared with a *perturbation run*, in which the initial temperature fields are disturbed randomly at all levels for the entire hemisphere (Lorenz, 1963, 1965; National Academy of Sciences, 1966; Smagorinsky, 1969). Beyond the second day, the temperature error grows more rapidly in the lower troposphere at the zonal wave numbers six to ten, suggesting that baroclinic instability is the primary cause (Eady, 1949; Smagorinsky, 1969). At the same time, the error spreads to a wider range of the spectrum.

Presumably, an initial disturbance, even on the turbulence scale, may eventually develop to a sizeable magnitude in the synoptic or even hemispheric scales. The period for the development of the error to a "fatal" level is tentatively considered to be beyond three weeks (Smagorinsky, 1969). A loss of predictability in synoptic scales is therefore inevitable, since atmospheric measurements cannot, in practice, be made in the turbulent range for the entire globe This period supposedly is the "ultimate limit of predictability" of large scale weather. An accurate prediction by the hydro-thermodynamical method can never be achieved *deterministically* beyond this limit. This error growth may be called "predictability decay".

In these tests, the initial adjustment is clearly reflected in a drop of the standard deviation at the beginning (see Fig. 4). The distribution of surface pressure was restored for the most part within several hours. It is clear that this process is not due to frictional dissipation but is primarily due to other dynamical characteristics of a fluid in a rotating frame.

The problem of adjustment (or adaptation) has been one of the most interesting fundamental subjects in oceanography and meteorology (see the review paper by Phillips, 1963). In many of the previous studies, discussion has been focused on the nature of the final state of balance and on how the inertia-gravity motion suddenly generated decays in the adjustment of the system toward this balance.

Veronis (1956) studied the partition of energy between geostrophic and non-geostrophic motions for a two layer ocean model driven by wind stress. One conclusion, among others, is that if the stress is impulsive (less than six pendulum hours), 74 percent of energy goes into inertia-gravity motion, and that for longer period wind stress variations, the partitioning to the non-geostrophic modes is considerably decreased.

The phenomenon described above is essentially the same as that in the present study. After the adjustment period, the behavior of the solution enters into a second stage in which the error starts to grow. This is the inherent error amplification due to baroclinic instability, where the source of error is the remnant left after damping, even if small. It is presumed that before the error is completely dissipated by friction, some portion is converted to the geostrophic component, and at the same time, is amplified by the ever existing instability.

In general, however, any arbitrary variable may not always have the above error behavior characteristics. Some variables are so essential that their errors immediately grow at a sizeable rate without any apparent damping. On the other hand, some variables may quickly adjust, so that there is very little time for baroclinic instability to act. The question therefore is: what is the ordering of variables according to their susceptibility to error growth?

2. The method of the test

As a control run, a time integration has been made with a prediction model using a complete set of realistic initial conditions. This will be denoted as Experiment 53G2. Next, the integration is repeated, but the structure of one of the variables in the initial conditions is deliberately degraded in some way. This is the modification (or perturbation) run. The results of both integrations are then compared. If the solutions are found to be essentially the same, it is concluded that the variable in question is redundant.

The prediction model used was that of the Geophysical Fluid Dynamics Laboratory (Smagorinsky, Manabe & Holloway, 1965; Manabe, Smagorinsky & Strickler, 1965). The general characteristics of the model are: nine-level; hemispheric; primitive equations; stereographic mapping with 40 gridpoints between the pole and the equator ("N = 40"), i.e., the grid size is about 270 km at mid-latitudes; the model admits moist adiabatic processes with condensation criterion of 100 percent relative humidity; a moist adiabatic convective temperature and water vapor adjustment; radiation; mountains; land-sea contrast; a non-linear lateral viscosity with "Karman" constant of 0.4; cloud cover specified as a function of height and latitude only; sea surface temperature as prepared by the United States Navy Hydrographic Office; continental surface temperature as determined from a local heat balance at the surface; availability of water for evaporation over land and sea is 0.5 and 1.0, respectively; water vapor and ozone as absorbers of radiant energy are specified as functions of height and latitude and are constant with time; the surface drag over land and sea is assumed equal.

The initial data of wind, temperature, and water vapor from 1,000 mb to 10 mb level for 1,200 GMT, January 9, 1964, were used. This is the same case as that treated in Miyakoda *et al.* (1969).

As a measure of dissimilarity of the two maps, the standard deviation of the difference in the variables (the "error") between two runs was taken. Let us denote χ_c as the variable in the control run and χ_m as that in the modification run. Definitions of the various quantities are as follows:

Error of χ: $\Delta\chi = \chi_m - \chi_c$

Mean of error: $\overline{\Delta\chi} = (\Sigma\Delta\chi)/n$

where the summation is over all gridpoints poleward of 20° N and n is the number of gridpoints.

Standard deviation:

$$S(\Delta\chi) = \sqrt{\Sigma(\Delta\chi - \overline{\Delta\chi})^2/n}$$

Denoting by $\chi(t)$ the variable for the day t and $\chi(0)$ that for the initial time, we have

Error for a "persistence" forecast:

$$\delta\chi = \chi(t) - \chi(0)$$

Standard deviation of error for persistence forecast:

$$\sqrt{\Sigma(\delta\chi_c - \overline{\delta\chi_c})^2/n}$$

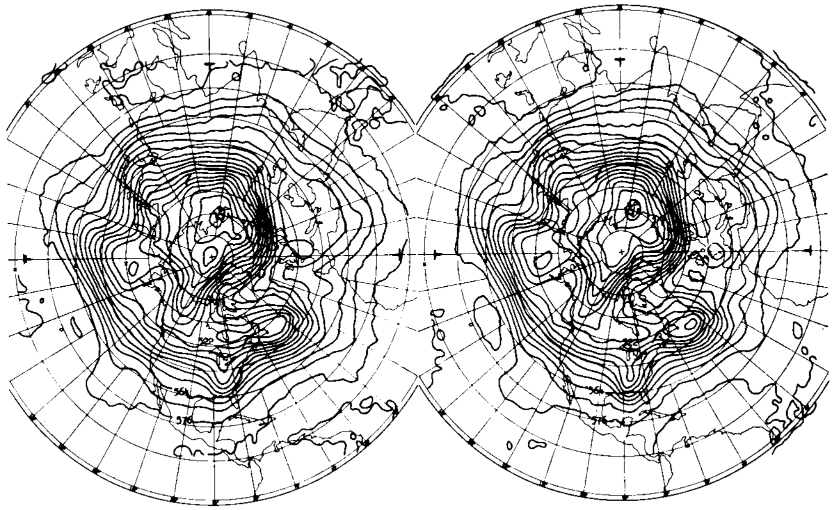

Fig. 1. Comparison of the 500 mb geopotential height for the control run (left) and the modification run of the surface pressure (Experiment 53M) (right) for fourth day. The contour interval is 60 m.

3. Choice of the variables

We have already discussed earlier tests on the geostrophic adjustment and on the generation of the vertical motion fields.

The variables for the test of possible redundancy were selected neither randomly nor systematically, but rather intuitively. They are: surface pressure; relative humidity; boundary layer temperature; boundary layer wind.

These variables are relatively difficult to measure or to determine with high accuracy either for technical reasons or because of the variability of the quantity. It is therefore natural to inquire as to the degree of accuracy needed for input to a hydrodynamical prediction.

It was also of interest to perform experiments constraining the initial stratospheric structure. But since such experiments require special treatment in analysis and interpretation, those results will be reported in a separate paper.

4. The results of the experiments

(a) *Surface pressure (Experiment 53M)*

The modification run in this case starts with the surface pressure (p_*) purposely aberrated. We took the initial surface pressure field such that the geopotential height at 1,000 mb level is a constant for the entire domain. This avoids the excessively drastic imbalance which would occur over terrain had we taken $p_* = \text{const}$.

It may be useful here to mention how the surface pressure is computed in the model, though it is only a part of the system of prediction equations. The tendency equation is

$$\frac{\partial p_*}{\partial t} + \nabla(\mathbf{v}p_*) + \frac{\partial \tilde{\omega} p_*}{\partial Q} = 0 \tag{4.1}$$

where

$$p = p_* \cdot Q$$

$$\tilde{\omega} \equiv \frac{dQ}{dt} \tag{4.2}$$

p and p_* are the pressures at an arbitrary height and at the ground surface, and $\mathbf{v}$ is the horizontal wind vector. With the vertical boundary conditions,

$$\tilde{\omega} = 0 \quad \text{at} \quad Q = 0 \text{ and } 1, \tag{4.3}$$

equation (4.1) yields upon integration over Q from 0 to 1,

$$\frac{\partial p_*}{\partial t} = -\int_0^1 \nabla(\mathbf{v}p_*)\, dQ \tag{4.4}$$

Hence, if the initial state of p_* is given, the surface pressure for any time can be obtained, provided that $\mathbf{v}$ is known. The total mass of air is conserved for a closed domain.

Figs. 1 and 2 show the time evolution of geopotential fields for the control run (Experiment 53G2) and the test run with degraded

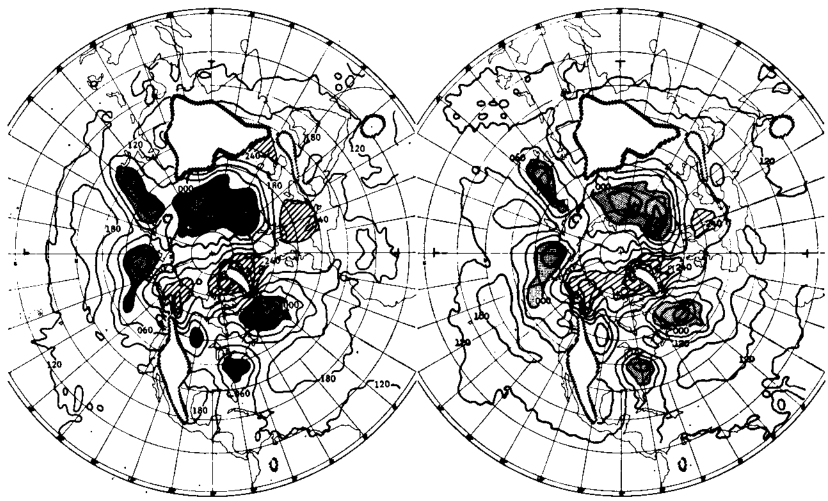

Fig. 2. The same as in Fig. 1, but for the 1,000 mb geopotential height. The contour interval is 60 m. The anticyclone areas with the geopotential value greater than 240 m are hatched, and the cyclone areas with the geopotential value less than zero meter are stippled. The mountains above 1,000 mb level are blank areas enclosed by small segmented lines.

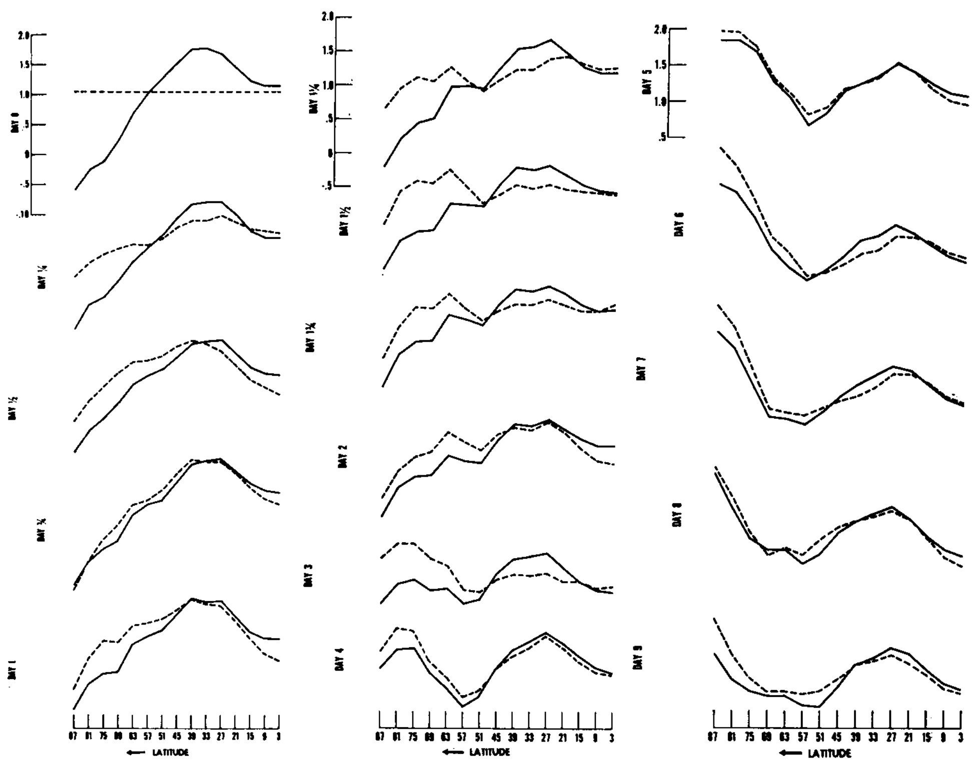

Fig. 3. Latitudinal distribution of 1,000 mb zonally averaged geopotential height for the control run (solid lines) and the modification run (Experiment 53M) (dashed lines).

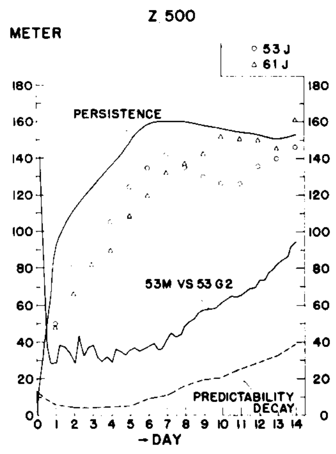

Fig. 4. Standard deviation of error in 500 mb geopotential height between the control run (53G2) and the modification run (53M). For reference, the standard deviations of error of the two real predictions compared against the observed are given. A "predictability decay" curve is also shown.

initial surface pressure (Experiment 53M). The maps are for the 500 mb and 1,000 mb levels at the fourth day after initial time.

The height contours for the two runs are quite similar. However, the 1,000 mb map at the twelfth day begins to show some discrepancy (not shown).

Fig. 3 displays another aspect of the result. The zonally averaged geopotential height at the 1,000 mb level is shown as a function of latitude. For the first two days, the curves are shown at six-hour intervals, and thereafter at 24-hour intervals. The profile in the test run is almost latitudinally constant initially. Eighteen hours later, the two curves are very close to each other, then start to deviate slightly from the first to third days, are close again on the seventh day, and start to depart on the ninth day. A kind of tidal oscillation is taking place. Apparently the local, internal redistribution of mass is fairly rapid, but the adjustment on the hemispheric scale is much slower.

Fig. 4 is a graph of the standard deviation and of the persistence forecast of the 500 mb geopotential height. The error of the persistence forecast at its asymptotic level gives a measure of the natural variability of a quantity. The figure also includes the standard deviation of error taken from the corresponding predictability study (Smagorinsky, 1969). Shown also are the standard deviations of the predictions for two January cases, which were reported by Miyakoda *et al.* (1969) (Experiment 53J for January, 1964, and Experiment 61J for January 1966). They are the real errors of the predictions compared with observed verification data, and are shown by small circles and triangles. It is to be noted that the differences between the control and test runs in the present study are much lower than the real errors.

(b) *Humidity* (*Experiment 53P*)

In the test for water vapor, the modification of initial conditions was made such that the relative humidity was specified as the zonal mean for the respective latitude and height. The two week integration was then carried out.

The time evolution of humidity for the control run (Experiment 53G2) and the modification run (Experiment 53P) is shown in Fig. 5. It is a time-longitude diagram of the relative humidity at level 8 (about 926 mb) in the belt between 35° and 45° N latitude. In the test run, the humidity at the initial time is constant, i.e., about 70 percent. The longitudinal structure starts to develop promptly approaching that of the control run as the prediction progresses. The relaxation time is less than one day.

It is clear therefore that the initial water vapor data in the extra-tropics is largely redundant. Had another forecast been made with different initial water vapor data, essentially the same result should have been expected. It follows, then, that if the initial water vapor data is not consistent nor compatible with the prediction model, the distribution as specified may not remain intact, and the information not utilized. Instead, it may be washed out as a consequence of adaptation despite the great effort which may have been required to construct the data originally, irrespective of its accuracy. This conclusion may not be entirely surprising to those who are acquainted with the large-scale correlation between relative humidity and vertical velocity. This is quite analogous to the futility of trying to use a directly measured vertical motion field, if it were possible. In fact, the two are very closely linked dynamically. A three-dimensional specification of the mass field in the extra-tropics, or preferably the horizontal wind field, will determine the vertical velocity distribution and the humidity distribution.

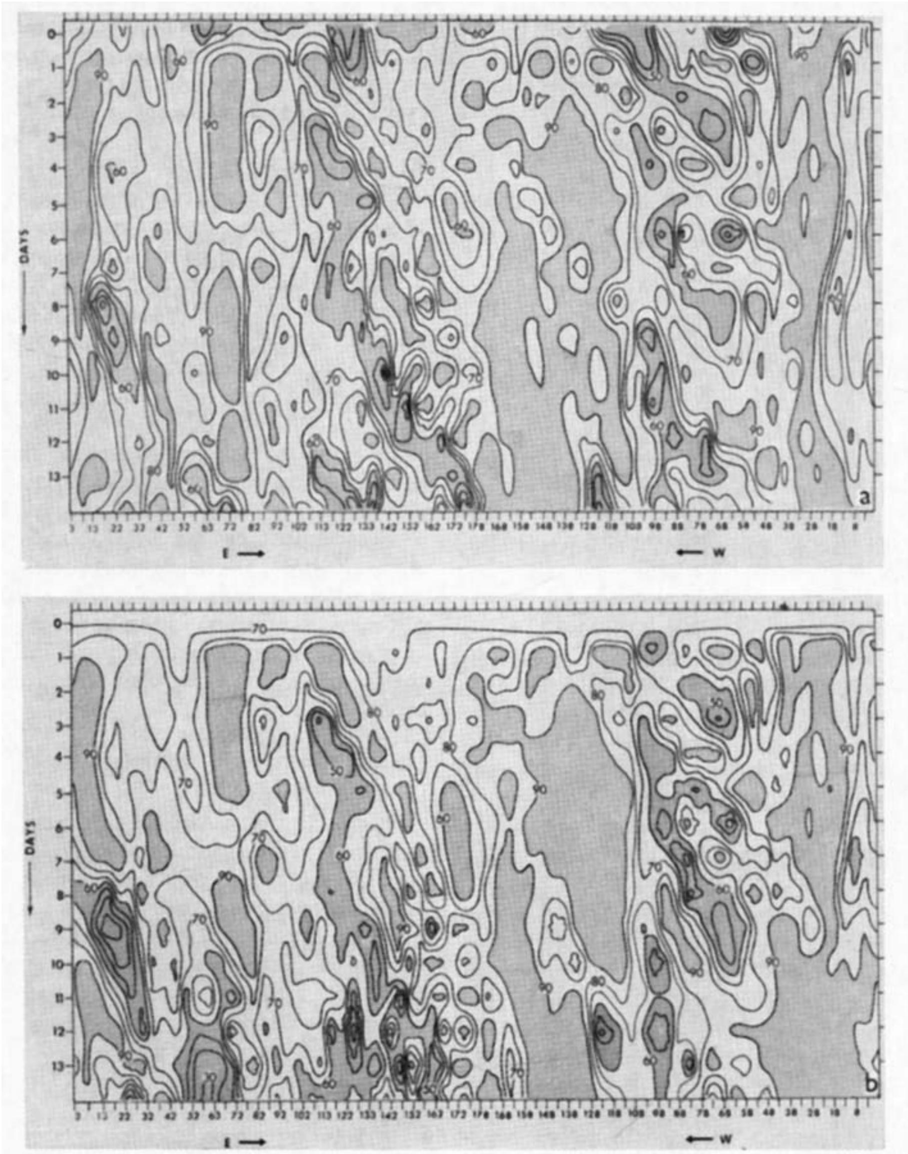

Fig. 5. Longitude-time diagram of the relative humidity at level 8 in a zonal belt between 35° and 45° N. (a) the control run, (b) the modification run (Experiment 53P). The contour interval is 10 percent.

Figs. 6 and 7 give measures of similarity between the humidity fields of the two runs. The first is the standard deviation of the difference in humidity at the 500 mb and 850 mb levels. The latter is the correlation coefficient of the change of relative humidity for the 500 mb, 700 mb, and 850 mb levels. The domain of the comparison is north of 20° N. If h_c is the humidity in the control run and h_m of the modification run, if $\Delta h = h_m - h_c$ and $\delta h = h(t) - h(0)$, then the correlation coefficient is given by

$$\frac{\Sigma(\delta h\delta h - \overline{\delta h_c})\,\chi(\delta h_m - \overline{\delta h_m})/n}{S(\delta h_c)\cdot S(\delta h_m)}$$

In Fig. 6 the two curves of the standard deviation for 500 mb and 850 mb levels start at large values, but they decrease and reach their

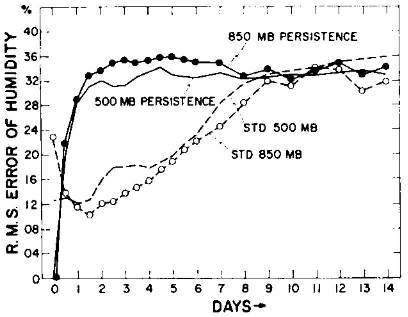

Fig. 6. Standard deviation of error in relative humidity between the control run and the modification run (Experiment 53P) at 500 and 850 mb.

minima at 1 or 1.5 days as the consequence of initial adjustment. The curves rise after the second day, and level off at about the eighth day, where they reach their asymptotic values. A similar tendency can be seen in the correlation coefficients in Fig. 7. It is unreasonable to expect extremely detailed agreement in the humidity patterns, since they normally have sharp horizontal variations and are usually quite erratic with time.

It is customarily assumed that accurate water vapor data is indispensable for forecasting smaller-scale weather phenomena. In this connection, it may be of interest to see how quickly the precipitation pattern is reconstructed in the modification run. Fig. 8 gives the six hourly amounts of precipitation over the United States and a part of Canada for the two runs at six-hour intervals during the first day. In the first six hours, the distribution of precipitation is quite different in the two runs, but by the second period, they have already become quite similar to each other. Of course, this is just one case, and, therefore, any generalized conclusion must be made cautiously. One might conclude, however, that the restoration of the rainfall pattern is rapid. It would be of interest to repeat this experiment for a case of excessively heavy precipitation.

Rapid adjustment of the humidity pattern does not occur in the stratosphere because the residence time is fairly long. Therefore, the total amount of water vapor given at initial time in the upper atmosphere must be carefully determined independently; otherwise, the original incorrectly specified water vapor could hardly be replaced by a more correct balanced water vapor distribution.

The degree of redundancy of water vapor in the tropics cannot be resolved in the present experiment. A separate study addressing this specific problem will be necessary.

(c) *Boundary layer wind* (*Experiment 53R*)

In this test, the initial condition for wind in the modification run was specified as zero at the lowest two levels in the model, approximately at 926 mb and 991 mb. A two-week integration was then performed (Experiment 53R). In Fig. 9, the latitude-time charts for the zonally averaged zonal components of wind at level 8 (or 926 mb) are shown for the two experiments.

The correlation coefficients for the change of the zonal component of wind from the initial time are displayed in Fig. 10 for levels 5, 8, and 9 (approximately 500, 926, and 991 mb, respectively). The correlation is low at the beginning, but increased rapidly. The restoration of the wind structure in the boundary layer occurs within about one day. It is of interest that the deterioration of the wind prediction is most rapid at 926 mb.

(d) *Boundary layer temperature* (*Experiment 53Q1*)

In this test, the modification run was made with the initial temperatures at levels 8 and 9 set to their zonal mean values for each latitude. The subsequent two-week integration is denoted by Experiment 53Q1. The measures of similarity between the two runs are shown in terms of the standard deviation (Fig. 11) and the correlation coefficients of the time changes of temperature from the initial time (Fig. 12). The 850 mb temperature has been interpolated

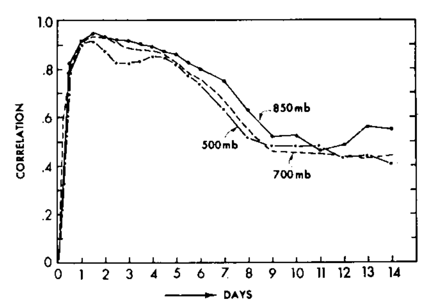

Fig. 7. Correlation coefficient of the change in relative humidity from the initial time between the control run and the modification run (Experiment 53P) for 500, 700, and 850 mb.

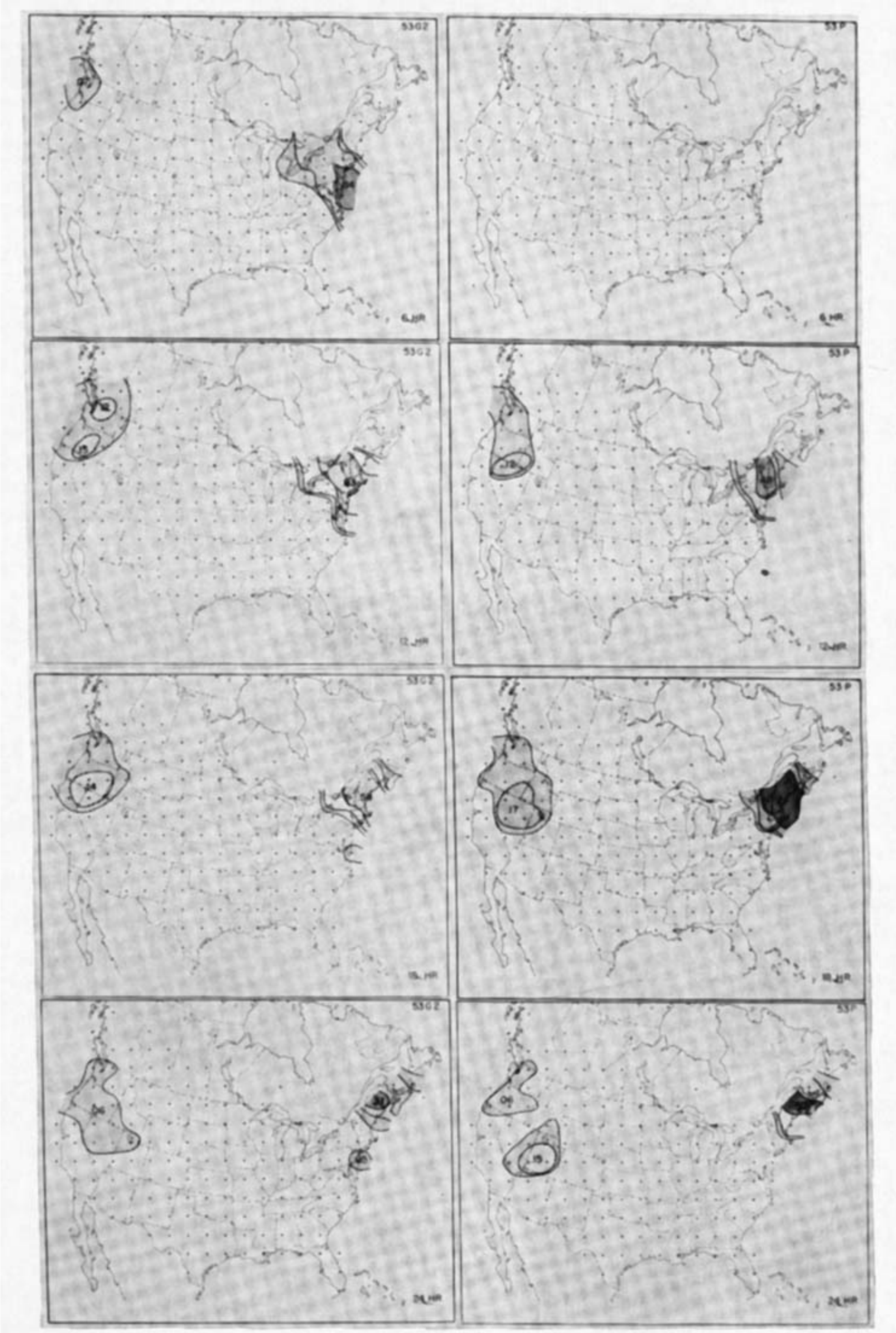

Fig. 8. The six hourly accumulation of precipitation for the control run (left) and the modification run (Experiment 53P) at six-hour intervals during the first day. The isopleths are .03, .10, .30 inches.

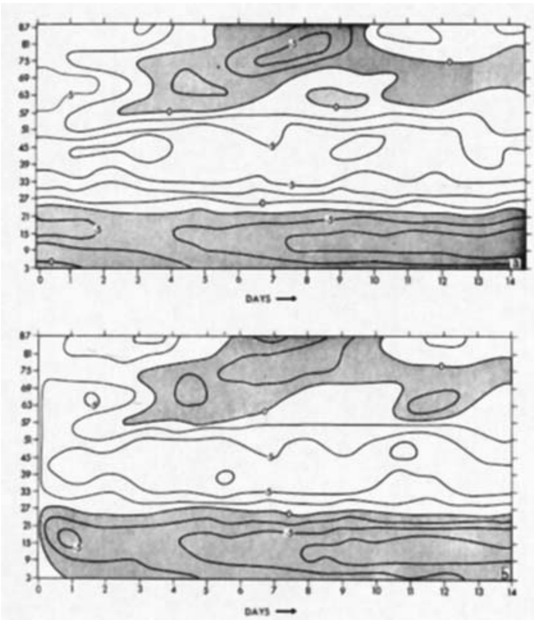

Fig. 9. Time-latitude charts of the zonally averaged zonal wind at level 8 (about 926 mb) (*A*) the control run and (*B*) the modification run (Experiment 53R). The contour interval is 2.5 m/sec.

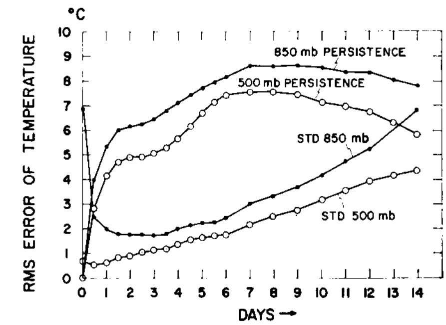

Fig. 11. Standard deviation of error in temperature between the control run and the modification run (Experiment 53Q1) at 500 and 850 mb levels.

from those of levels 7 and 8, where that of level 8 is disturbed initially.

In this case also, the correct temperature field (for example, 1,000 mb and 850 mb) in the modification run is restored, but a bit less promptly. Later, the typical error growth occurs, but the decay of the temperature prediction itself is slow. It is worth noting that the degradation of temperature prediction is fastest at the 700 mb and 850 mb levels. This result, as well as that for the test of the boundary layer winds, is consistent with the predictability study. There it was found that the temperature error developed most rapidly in the lower troposphere, where the conversion from eddy available potential to eddy kinetic energy is at a maximum.

5. Effect on the geopotential fields and precipitation patterns

Let us next examine how inaccurate initial conditions influence the geopotential height and precipitation distribution. These quantities for many purposes can be considered to be the most important elements in weather prediction.

(a) *Geopotential fields*

Since the maps for the control and modification runs for the surface pressure were already shown, the maps displayed here are 1,000 and

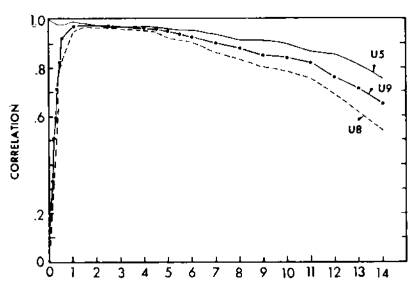

Fig. 10. Correlation coefficient of the change in intensity of zonal wind from the initial time between the control run and the modification run (Experiment 53R) for level 5, 8, and 9 (approximately 500, 926, and 991 mb, respectively).

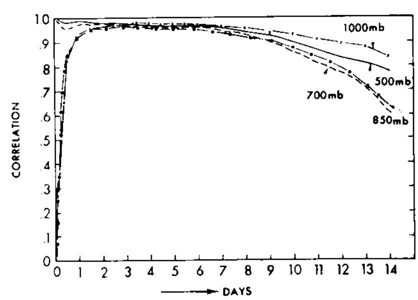

Fig. 12. Correlation coefficient of the change in temperature from the initial time between the control run and the modification run (Experiment 53Q1) at 500, 700, 850, and 1,000 mb.

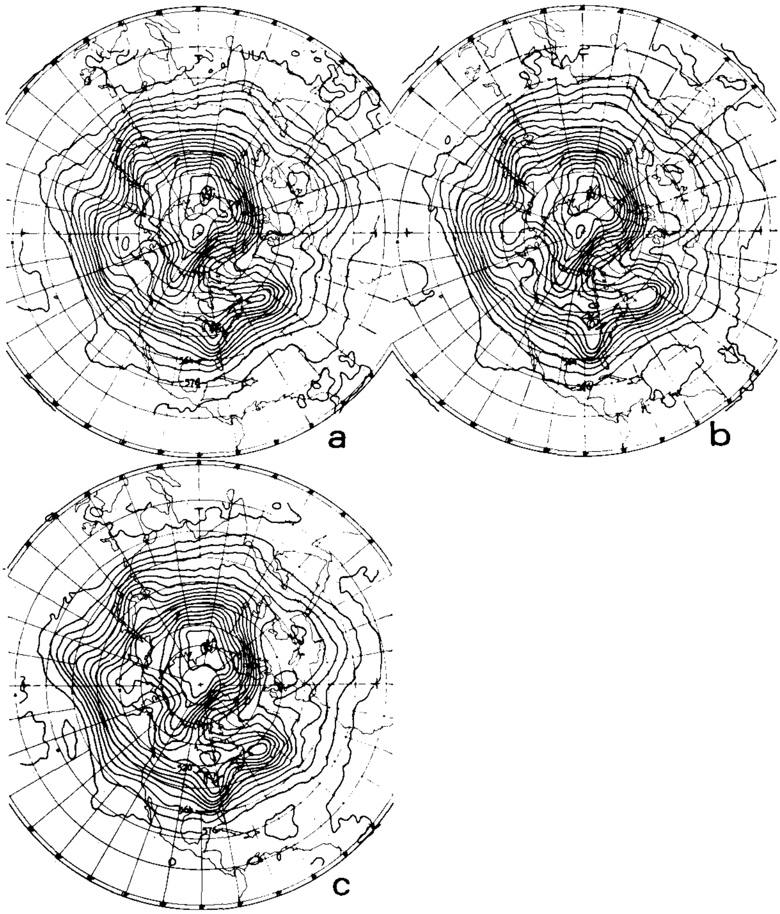

Fig. 13. The 500 mb geopotential height for the modification runs. (*a:* Experiment 53P; *b:* 53R; and *c:* 53Q1) for fourth day. The corresponding control run is in Fig. 1.

500 mb geopotential forecasts at the fourth day, in which the initial water vapor, boundary layer temperature, and boundary layer wind were disturbed.

Although not shown, the second day shows very little detectable error. The deterioration is more rapid at 1,000 mb than at 500 mb. On the eighth day, marked differences are to be found in the 1,000 mb patterns, but the 500 mb height maps are still similar. For each modification experiment, the standard deviations of the geopotential with respect to the control run (53G2) are taken at 50 mb, 500 mb, and 1,000 mb (Fig. 15).

The effect of the initial surface pressure abberation (Experiment 53M) is appreciable at the early stage, but does not amplify much later. The influence of the water vapor change (Experiment 53P) is smallest for the first six days and does not grow rapidly thereafter. On the fourteenth day, the decay in the accuracy of the 500 mb and 1,000 mb prediction is, in ascending order, as follows: water vapor (53P), surface pressure (53M), boundary layer wind (53R), and boundary layer temperature (53Q1). On the other hand, for the 50 mb height, the effect of the surface pressure (53M) is the smallest.

(b) *Precipitation*

The precipitation patterns for 24-hour accumulations over the United States and a part of Canada are shown for the control run and each of the modification runs in Figs. 16, 17, and 18 for the second, fourth, and eighth days, re-

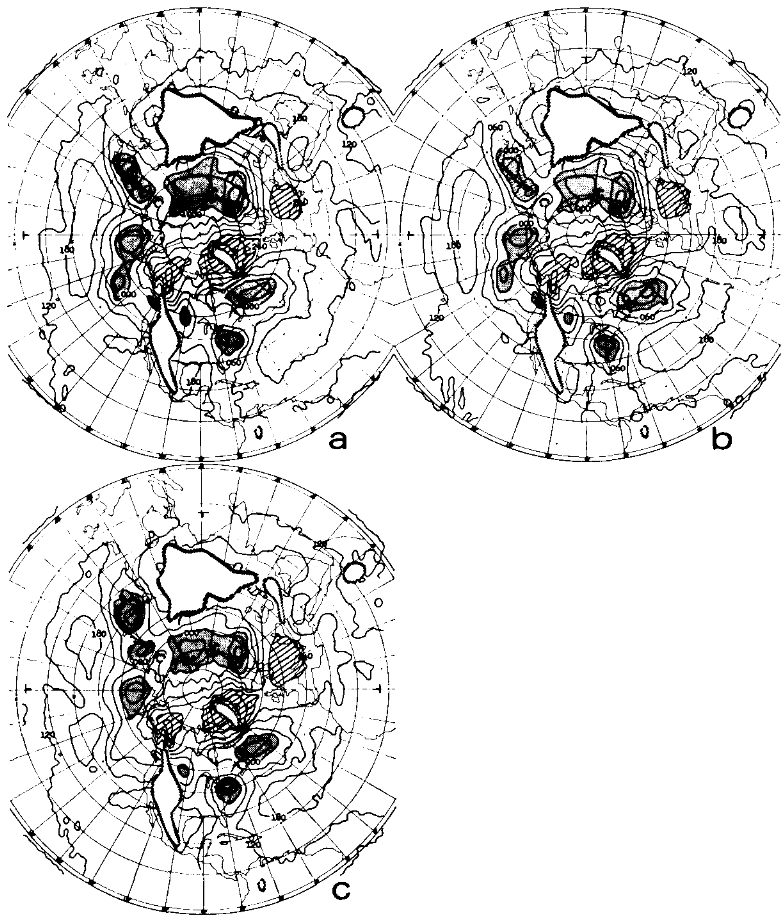

Fig. 14. The same as in Fig. 13 but for the 1,000 mb height. The corresponding control run is in Fig. 2.

spectively. The contour lines are for .1, .3, and 1.0 inches per day.

The agreement in rainfall distribution among each of the runs is good for the second and fourth days. Deviations are apparent at the eighth day, and by the twelfth day (not shown), the disagreements are quite large.

6. Concluding remarks

The tests suggest that some variables such as surface pressure, tropospheric relative humidity, boundary layer wind, and boundary layer temperature are not really essential as initial data for prediction of extra-tropical cyclone-scale weather for a few days ahead. As soon as the time integration of the prognostic equations starts, an adjustment process takes place, and the major error in the solutions due to inadequate initial data tends to disappear within 6 to 24 hours. The mechanisms of adjustment for each variable differ.

In the first 6 to 7 days (except the first 6 hours), the geopotential fields (or temperature fields) as well as the distribution of precipitation in the extra-tropics show little sensitivity to ill-defined initial data for any of the test variables.

After about 8 days, however, all predictions suffer inherited error growth typical of predictability experiments. However, the error development in the present experiments seems to be somewhat larger than is usually ascribed to predictability decay. In practice, it is feasible to employ more reasonable and more realistic

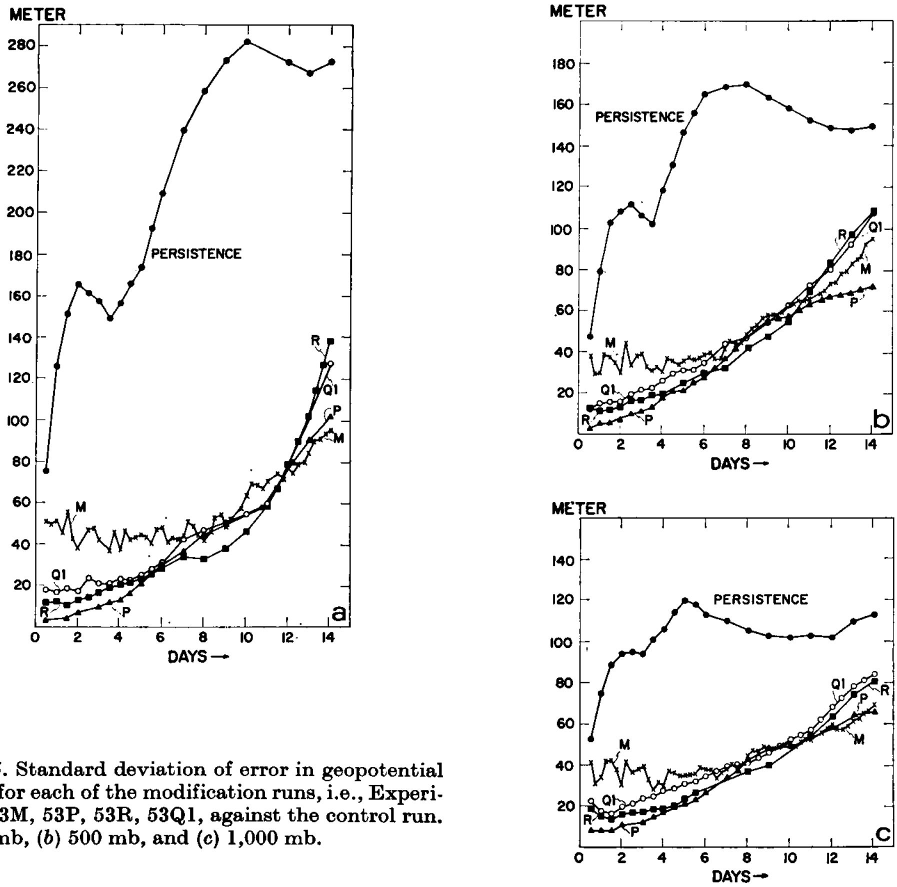

Fig. 15. Standard deviation of error in geopotential height for each of the modification runs, i.e., Experiment 53M, 53P, 53R, 53Q1, against the control run. (*a*) 50 mb, (*b*) 500 mb, and (*c*) 1,000 mb.

initial data for these variables than those used in these exaggerated tests, so that the error growth can be considerably tempered. It cannot be denied, however, that any improvement in the accuracy for these variables might raise the accuracy of the prediction after 8 days, but a greatly improved prognosis should not be expected.

Conclusions on the redundancy of water vapor information in the tropics should be reserved until a more specific test is made. The tests in this paper were performed separately for each of the variables. As a next step, it may be desirable to make a test of simultaneous modifications of all of the variables treated here. The conclusions drawn in this paper are just from the simple results as displayed. To gain a deeper insight and clearer comprehension of the problem, a theoretical analysis will be necessary. We are continuing these tests to determine the ordering of wind and temperature structure at various levels in the free troposphere and in the stratosphere.

As indicated earlier, one aim of this kind of research is to provide a guide for the design of the atmospheric observational systems by determining a priority for quantities that must be measured and the accuracy required for each. It bears also on devising successful schemes for the four-dimensional assimilation of observed data.

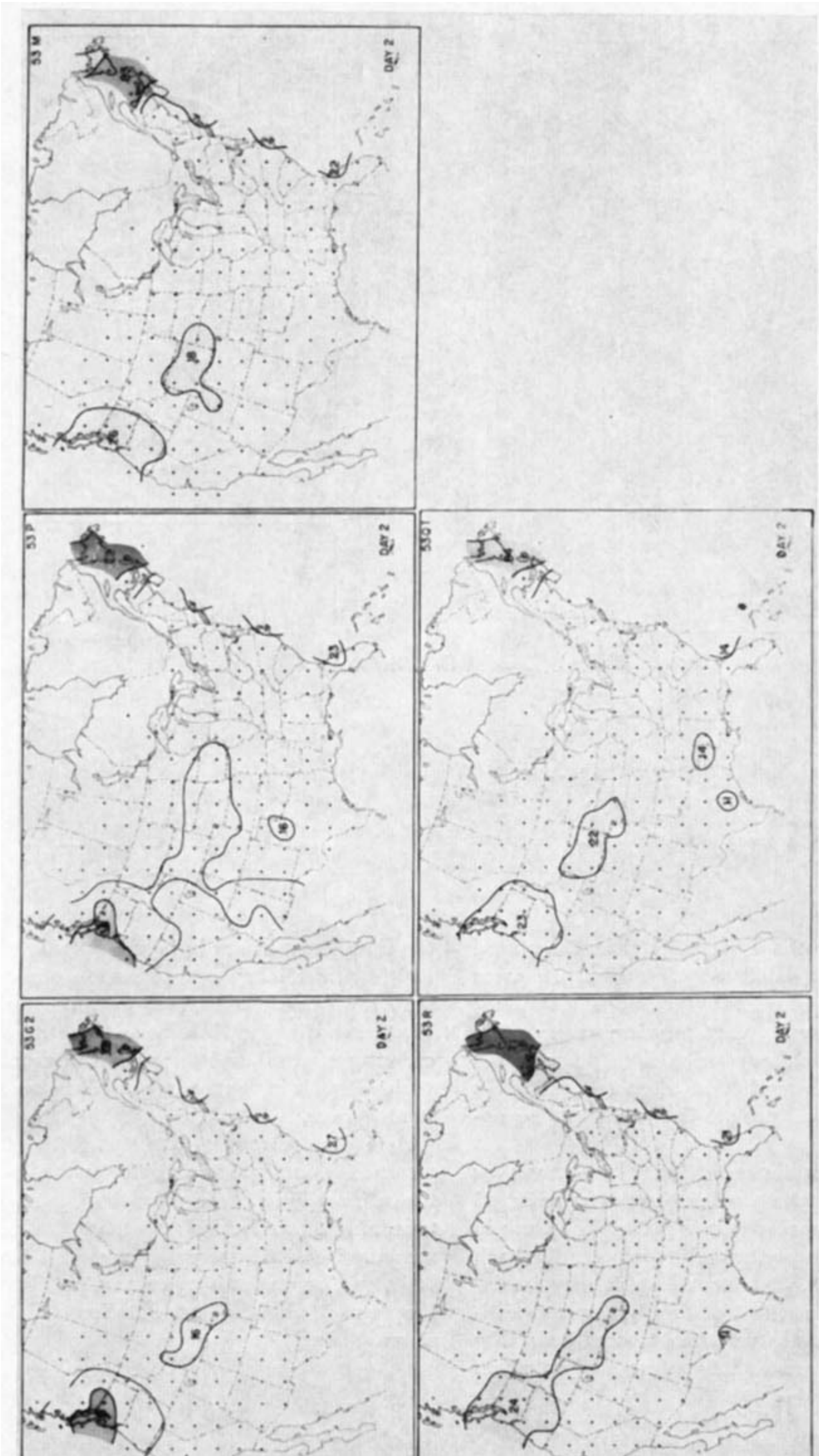

Fig. 16. Twenty-four hourly accumulation of precipitation for the control run and the modification run (53M, 53P, 53R, 53Q1) for second day. The isopleths are .1, .3, and 1.0 inches.

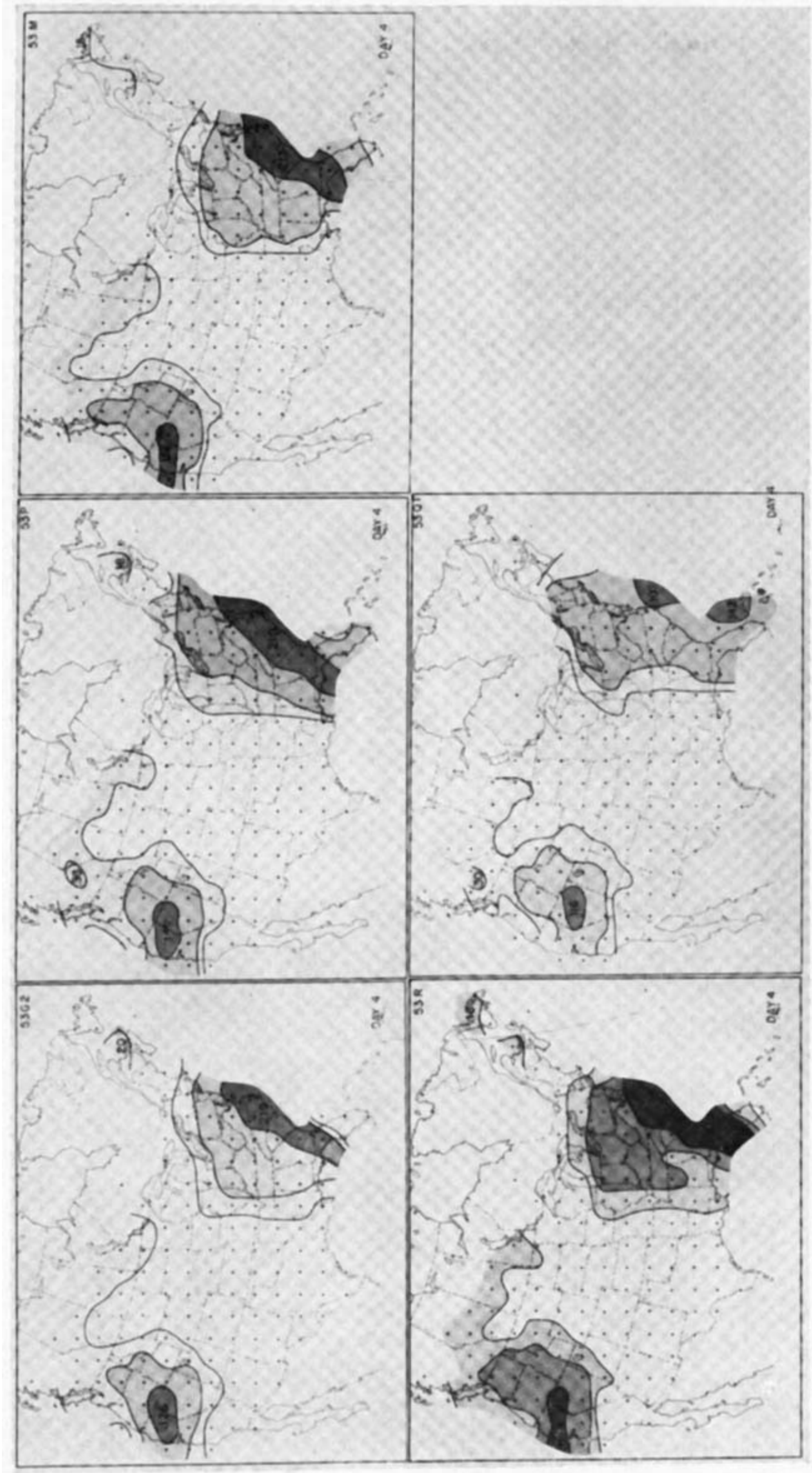

Fig. 17. The same as in Fig. 16 but for fourth day.

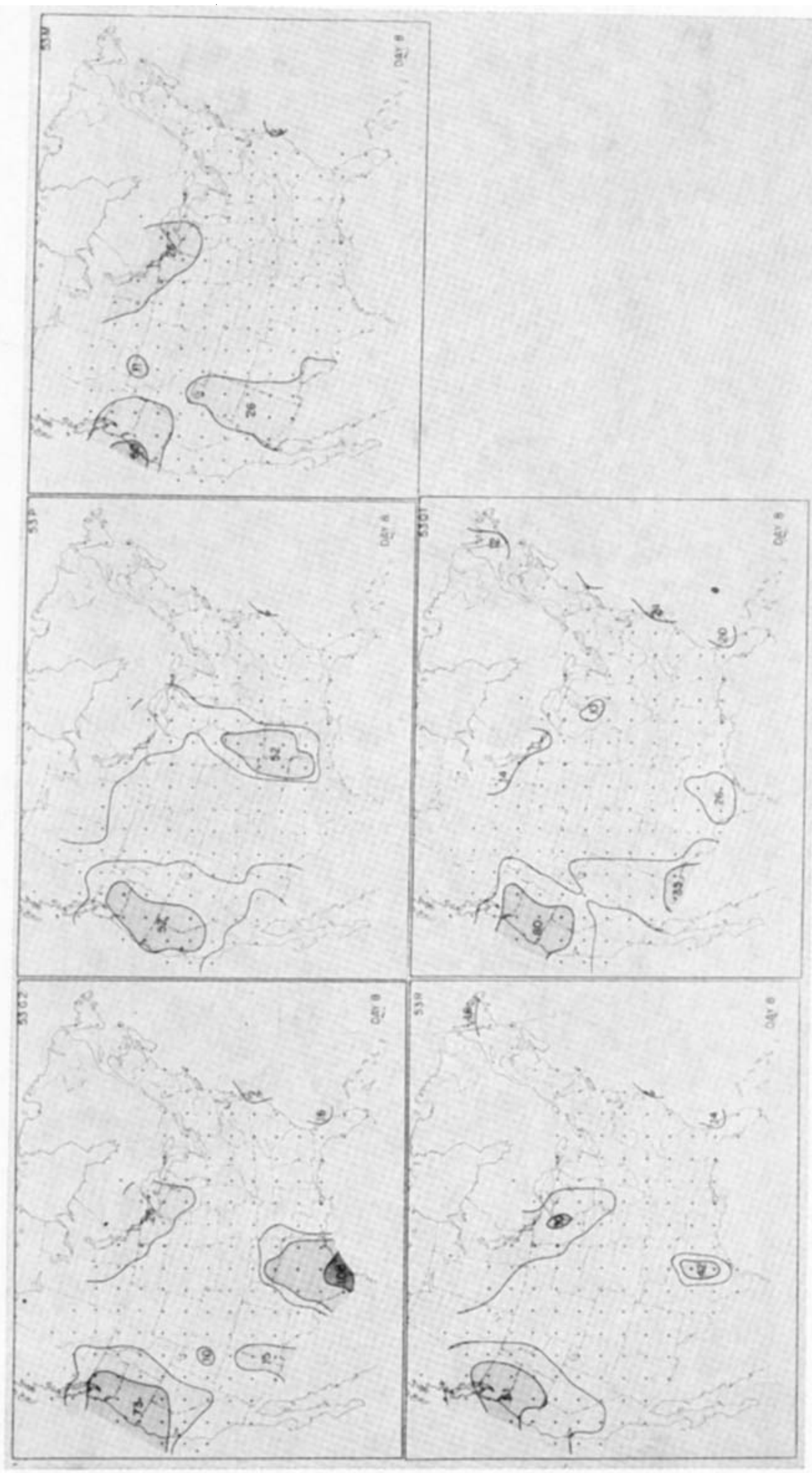

Fig. 18. The same as in Fig. 16 but for eighth day.

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ОТНОСИТЕЛЬНАЯ ВАЖНОСТЬ ПЕРЕМЕННЫХ В НАЧАЛЬНЫХ УСЛОВИЯХ ДЛЯ ДИНАМ ИЧЕСКОГО ПРЕДСКАЗАНИЯ ПОГОДЫ

Были проделаны опыты по оценке относительной важности больших ошибок в некоторых переменных, формально необходимых для начальных условий при интегрировании по времени гидротермодинамических уравнений атмосферных движений. Первоначальная цель состояла в анализе процесса адаптации, непосредственно следующего за начальным возмущением. Испытывались следующие переменные: поверхностное давление, водяной пар, температура пограничного слоя, ветер в пограничном слое. Вероятно, что эти величины ни индивидуально, ни возможно даже коллективно не существенны для определения начальных условий из-за определяющего действия динамики. Эксперименты показали, что прогнозы, по-видимому, становятся одинаковыми после, приблизительно, шести часов, независимо от начальных значений этих величин. Однако через несколько дней ошибка ведет себя так же, как в экспериментах по предсказуемости.