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NUMERICAL MODELING OF WATER-VAPOR TRANSPORT DURING PRE-STORM AND COHMEX

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ABSTRACT

Initial conditions are designed for numerical simulation of mesoscale processes in the atmosphere using the LAMPS model. These initial conditions represent an idealized baroclinic wave in which the transport of water vapor can be simulated. The constructed atmosphere has two homogeneous air masses, polar front, polar jet stream and a stratosphere. All these simulate the basic structure of the earth's atmosphere. The hydrostatic and geostrophic balances make it possible to evaluate mutually consistent fields of wind and of the height of isobaric surfaces.

ACKNOWLEDGEMENT

This research was done in close collaboration with Dr. Michael Kalb who collaborated on the design of initial conditions and performed numerous experiments with the LAMPS model in order to test the designed initial conditions.

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INTRODUCTION

The study of results from observational programs Pre-STORM[1], COHMEX[2] and SPACE[3] will be accompanied with numerical modeling using the LAMPS[4] model. Numerical modeling will give insight in the dynamics of water-vapor transport and ensuing precipitation. Understanding of dynamics will, in turn, yield requirements for improved future satellite observations.

Modeling of atmospheric processes will be done along two different directions:

(a) prediction of development of observed states, and
(b) simulation of development under idealized conditions.

Here "prediction" means that observed data will be used in the model, even if the calculation will be done long time after the events have elapsed. "Simulation" means that the behavior of a simplified, schematized atmosphere will be studied. In such an atmosphere we may pinpoint exactly the factors that are of importance for the dynamics without having complications due to numerous other events that simultaneously occur in the atmosphere.

This report contains a study of initial conditions that can be used in simulation experiments. It is proposed that the development be studied of a synoptic-scale baroclinic wave (length of the order of 3000 km) around which mesoscale (order of 1000 km) are likely to develop. In particular the hypothesis will be tested that the water-vapor transport takes place predominantly in the low-level jet (LLJ) which is favored in stable layers of the lowest kilometer in the middle-latitude or tropical air masses.

The model used in this study is the one by Perkey (1976), as updated and described by Kalb (1985).

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1 STorm-scale Operational and Research Meteorology. An ongoing nation-wide project.

2 COoperative Huntsville Meteorological Experiment. Group of ongoing projects centered at Huntsville, Alabama.

3 Satellite Precipitation and Cloud Experiment. A research project of NASA.

4 Limited Area Mesoscale Prediction System. An analysis and prediction system developed by Drexel University.

OBJECTIVES

This work aimed to provide a hydrostatically and geostrophically balanced atmosphere that would be suitable for numerical simulation with the LAMPS model and which would show the main factors that characterize the development of the low-level jet stream in the atmosphere. The constructed idealized atmosphere was supposed to depict a baroclinic wave in the westerlies, with the polar jet stream at the elevatiion of about 9 km and an associated polar front. The schematic atmosphere was expected to be designed automatically by a Fortran program that would yield ready arrays of meteorological variables in a form in which they could be used as initial conditions in the LAMPS model.

The wave should result in development of mesoscale processes already in the first 12-24 h of development. The other approach where a zonally symmetric situation is used may yield a similar baroclinic development after about a week of development, as it has been demonstrated in the models of general atmospheric circulation. The suggested initial conditions should cut this development time to about 12-24 h of simulated time.

BAROCLINIC INITIAL CONDITIONS

The main objective of this modeling study is to design a baroclinic wave that can be used in the initial conditions in the LAMPS model. Therefore a set of atmospheric variables was designed that contain the main characteristics of an amplifying wave on the polar jet stream.

The schematic conditions in the atmosphere are prescribed by the distribution of temperature in the jet stream-polar front zone. Other variables (height of the isobaric surfaces, wind) are computed from the temperature distribution using the hydrostatic and geostrophic approximations. The basic temperature distribution is patterned after the model of the waves in the westerlies as it is known in the literature (e.g. by Palmén and Newton, 1969, pp. 176-185).

Current work concentrated on the distribution of meteorological variables in a spherical rectangle between 25 and 45 °N and 83-113 °W. The vertical extent of the model is between the earth's surface and 75 mb (about 16 km high). In this domain we use 39 grid points in x (longitude) direction, 33 points in y (latitude) and 38 in vertical. Constant increments in the x and y directions are chosen such that the latitude-longitude sectors are almost exactly square at 35 °N, with sides of 70 km. Vertical grid distance is 25 mb. Presently no mountains are used and the bottom of the model is at the mean sea level.

Vertical and meridional distribution of temperature is constructed from a state illustrated in Fig. 1. All values are evaluated by corresponding analytical expressions. The vertical lapse rate of temperature in the troposphere is taken as constant in the amount

$$\frac{\partial T}{\partial z} = -6.0 \times 10^{-3} \ \mathrm{K\,m^{-1}} \qquad (1)$$

There is also a weak ambient meridional gradient of temperature, such that the temperature decreases uniformly toward the north in the amount of

$$\frac{\partial T}{\partial y} = -3.0 \times 10^{-6} \ \mathrm{K\,m^{-1}} \qquad (2)$$

There are two domains of temperature in the troposphere, one south of the polar front, the other north of it. There is a discontinuous transition between the air masses, amounting to 5 K at the front. The front is designed

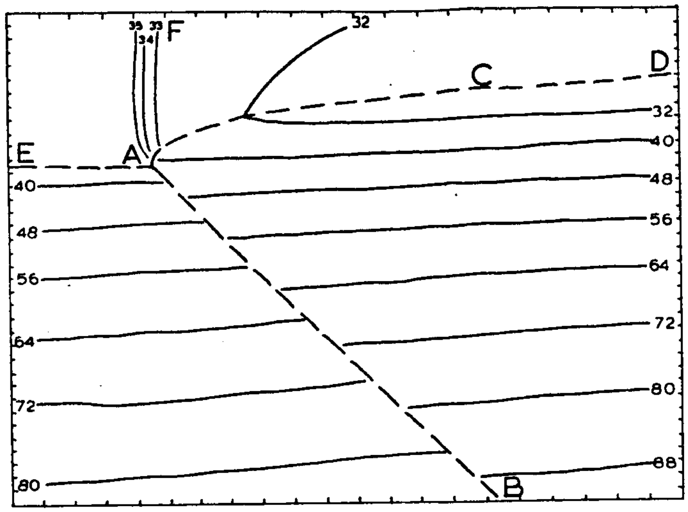

Fig. 1. Meridional and vertical section through the frontal zone in the model, showing the initial temperature distribution in kelvins, less 200 K. Isotherms are at every 8 K, with several intermediate 1-K isotherms at the discontinuity A-F. The polar front and the tropopause are shown by dashed lines. The horizontal coordinate is along the meridian, with ticks at every grid point, spaced 70 km apart. The vertical coordinate is pressure, from 75 mb on the top to 1000 mb at the bottom.

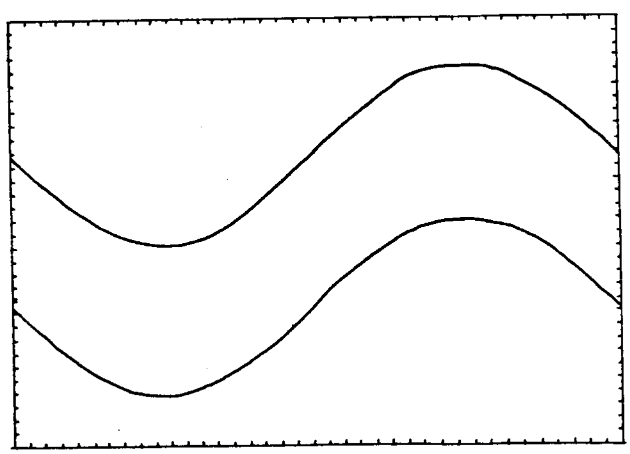

Fig. 2. The distribution of the coordinate q in the xy plane that shows the waves in the westerlies. The frontal zone is between the two sine curves. Ticks are at 70 km intervals along both (horizontal) coordinates.

slanted between the points A and B, such that its surface position (B) is 750 km south of its tropospheric position at the pressure of 355 mb (point A).

The polar tropopause is at the pressure of 355 mb, as it is typical for the earth's atmosphere. The middle-latitude troposphere is at the pressure of 235 mb between the surface position of the polar front and the southern boundary of the model (C-D in Fig. 1). Above the frontal zone (above the slanted frontal surface), the tropopause varies as a parabola, this is illustrated by the curve A-C in Fig. 1.

The temperature in the stratosphere is designed constant with height above the polar tropopause (above the line A-E). In the rest of the stratosphere, the temperature increases with height with the vertical derivative $\partial T/\partial z$ varying linearly from 0 above the point A to a value of 0.03 K m^{-1} above point D. The temperature on the bottom of the stratosphere is everywhere equal to the tropospheric temperature at this level. Since the tropopause varies in height between A and C, this introduces a cooling in the stratosphere with distance going "south" between A and C. A higher tropopause has a lower temperature since the tropospheric lapse rate of temperature is constant.

The tropospheric isotherms in Fig. 1 are not uniformly spaced in the vertical since the drawing is in the pressure coordinates. A constant lapse rate of temperature with height results in the temperature (T) variation with pressure p as

$$T = T_0 \left(\frac{p}{p_0}\right)^{\frac{Rh}{g}}$$

where T_0 and p_0 are the temperature and pressure at the surface. R is the gas constant, h is the (constant) lapse rate of temperature $-\partial T/\partial z$, and g is the acceleration of gravity.

The zonal variation of temperature simulates waves in the westerlies with wave length L=2450 km, what is the length of the domain of computation. The amplitude of the waves C is 4.6E5 m. Both these measurements correspond to a developing baroclinic wave on the polar jet stream. The sinusoidal shape of the baroclinic zone is achieved by introducing a new coordinate

$$q = A - By - C \sin\frac{2\pi}{L}x$$

where the constants A, B and C are selected such that the plot of q in the xy plane shows the pattern as in Fig. 2. The selected form of q gives the sinusoidal variation of the baroclinic zone with longitude. The reason behind this choice is that the development of mesoscale phenomena will start sooner than if the computation started from a zonally uniform state with small random perturbations. In this rather limited domain there will be much influence of the boundary conditions before mesoscale structures can be developed.

HEIGHT OF THE ISOBARIC SURFACES AND WIND

The pressure at the sea level is assumed everywhere equal to 1000 mb. It is expected (and confirmed with several preliminary experiments) that a more usual state with significant thermal advection will develop in the model after a short simulated time. The establishment of a developing low in the region downstream from the upper-level trough occurs in already after 1 h, whereas the wave pattern of Fig. 2 typically takes several weeks of simulated time to develop in models of general atmospheric circulation.

The first step in the evaluation of upper-level wind is the computation of the height of the isobaric surfaces. This is done with a trapezoidal integration of the hydrostatic equation. Trapezoidal integration is particularly suitable since the variables are arranged without staggering. All variables are defined in all grid points. Temperature from Fig. 1 is used, with some horizontal smoothing. Hydrostatic evaluation of height results in upper-level highs above warm air masses and upper-level lows above cold air masses. The lower boundary condition, at the earth's surface, was a constant pressure p=1000 mb in all grid points on the bottom boundary.

The geostrophic wind is by definition related to the height of isobaric surfaces. As planned, a jet stream appears above the frontal zone, a little above the level of the polar tropopause. This is illustrated in Fig. 3. There is also a westerly flow away from the frontal zone in the troposphere, as a consequence of the weak meridional temperature gradient described by the equation (2). The wind speed increases with height in the frontal zone since the integrated difference in temperature between the two air masses increases with height. Above the level of the polar tropopause (level at A in Fig. 1), the temperature does not decrease with height in the polar stratosphere, but it decreases further in the middle-latitude troposphere. Within one grid point above the level of A the temperature is equal on both sides of the line A-F, and above that point the temperature is higher on the polar side. In this region the wind speed decreases with height. Maximum wind speed appears at the level where the vertically integrated temperature is equal on both sides of the jet stream.

Unfortunately, the above described method for evaluation of wind yields undesirable results in the stratosphere. Large positive and negative values in both wind components appear near the top of the model, as it can be seen in Fig. 3. Upon inspection of these results, it could be established that small inaccuracies in the distribution of temperature result in large deviations of wind speed. This can be shown by the thermal wind relation

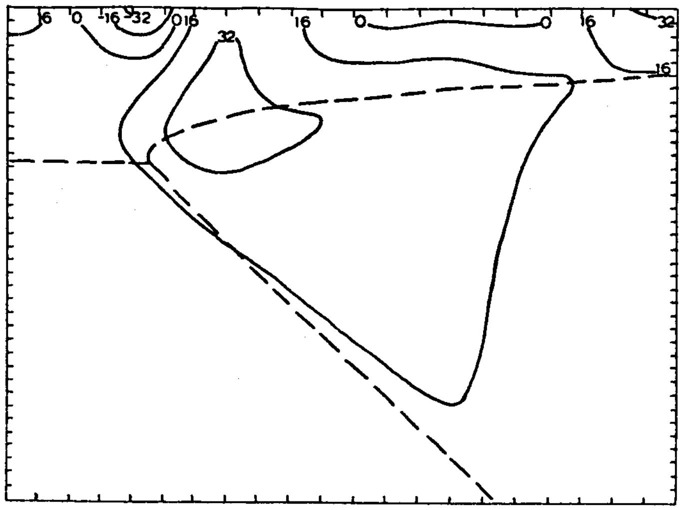

Fig. 3. Vertical meridional section through the frontal zone showing the x-component of wind based on the smoothed initial temperature distribution. Isotachs are for every 16 m s^{-1}. The coordinates are as in Fig. 1. The discontinuities from Fig. 1 (front and tropopause) are shown by dashed lines.

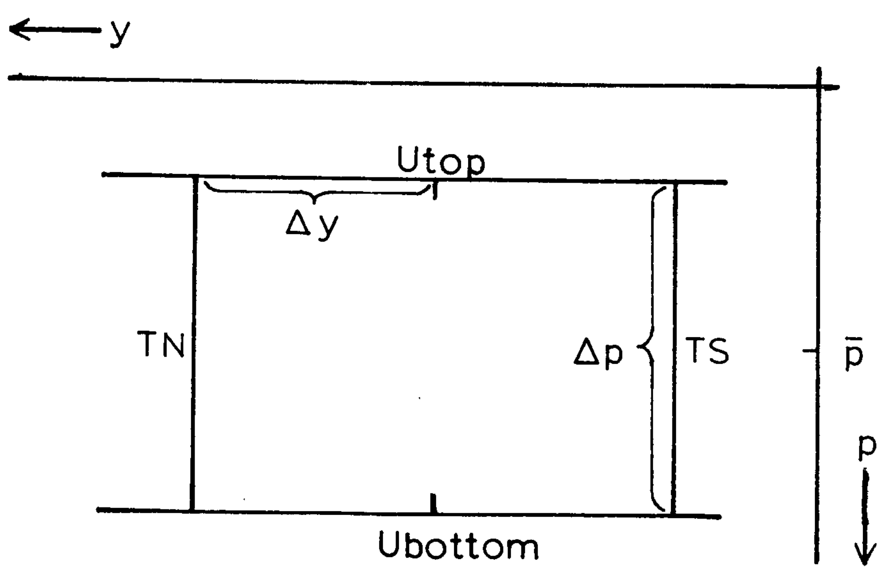

Fig. 4. Schematic representation of an area segment in the vertical plane. The quantities indicate the values used in the evaluation of thermal wind in (4) and (5).

$$\frac{\partial u}{\partial p} = \frac{R}{fp}\frac{\partial T}{\partial y} \tag{3}$$

where f is the Coriolis parameter. In finite differences this equation can be rewritten as

$$T_S = T_N + c\,(U_{top} - U_{bottom}) \,. \tag{4}$$

The subscripts S and N indicate the southern and northern values of vertically averaged temperature at the ends of the interval where $\partial T/\partial y$ is evaluated. The arrangement of variables in (4) is illustrated in Fig. 4. The local constant c is given by

$$c = 2\Delta y\, f p/(R\Delta p) \,. \tag{5}$$

For values near the 100-mb level, where both p and Δp can be estimated to be about 100 mb, and for the used grid distance Δy=70 km, c is about 0.1 K m^{-1} s^{-2}. This means that for a misestimated temperature difference of 1 K, there follows a difference of about 10 m s^{-1} in wind between top and bottom of the layer. If the thermal wind is evaluated over distances wider than $2\Delta y$, still larger discrepancies may occur in the computed geostrophic wind for every 1 K in temperature. Larger discrepancies in wind may occur if the temperature is missed by more than 1 K. The design of temperature from the previous section (Fig. 1) is quite satisfactory within the troposphere, however the choice of the form of the tropopause and the used vertical lapse rate in the stratosphere did not result in a satisfactory wind distribution. The discrepancy is particularly large around the vertical line A-F in Fig. 1 where the difference is large between the polar and middle stratospheres.

The wind distribution in the stratosphere is not particularly important for tropospheric processes. It is not known that the stratosphere directly influences the mesoscale processes in the troposphere. Still, undesirable effects may appear in the model due to high wind speed in the stratosphere. Therefore it was deemed necessary to find a more realistic distribution of temperature than shown in Fig. 1. It is a safe assumption that a more realistic development may be expected if there are no spurious jet streams in the stratosphere.

A modification of the temperature distribution was introduced using a plausible requirement that the wind speed at the top of the model should be equal to zero. Also the assumption was used that the modification of temperature for this purpose should be introduced only above the level of 300 mb, i. e. above the polar jet-stream level. This correction of temperature is then applied to the columns of air between 275 and 75 mb, equally to all grid points in these columns. The amount of correction is evaluated from (4) where the corrected values of temperature are introduced as

$$T_S + corrS = T_N + corrN - c\,U_{bottom} \tag{6}$$

Here $corrS$ and $corrN$ are the corrections added to the southern and northern columns of grid points. Thereby U_{top} has disappeared from (6), by definition. The correct amount of correction is obtained from (4) and (6) by subtraction:

$$corrS = corrN - c\,U_{top} \tag{7}$$

Here U_{top} is the formerly (presumably spuriously) obtained value of the wind speed at the top of the column in question.

There are two unknown terms (corrections) in (7). Therefore this correction must be introduced starting from one boundary where the correction must be prescribed. In the results presented below the correction on the northern boundary was used as $corrN=0$. In order to justify this boundary condition, the variation of temperature described by Fig. 1 was modified so that the temperture was exactly equal in the northernmost 4 grid points in every meridional row. The marching process described by (7) was done over the intervals of $2\Delta y$, with the wind component U at the top exactly between two columns of temperature. This was a leap-frog type of scheme, as it is appropriate for (3), which is a the first-order equation for T.

The results of the thermal wind correction (7) did not yield a good distribution of wind. There was a residual wind on the top of the model. This residual was nearly exactly proportional to the original spurious wind at the top of the model. Therefore the residual was removed satisfactorily by an empirical correction. Instead of (7), the correction was used with an empirical multiplier 0.84:

$$corrS = 0.84(\,corrN - c\,U_{top}\,) \tag{8}$$

So far an explanation for this multiplier of 0.84 cannot be offered; it may have to do with two-dimensionality of the geostrophic wind, whereas the correction is computed only along y.

With correction (8), the wind at the top became equal to zero, as forcefully adjusted, whereas there were comparatively small changes in temperature. The wind that resulted is illustrated in Fig. 5.

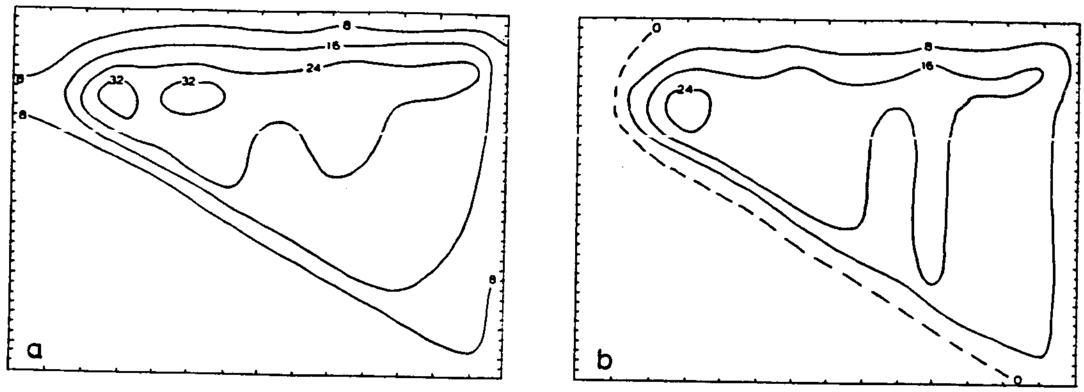

Fig. 5. Vertical section illustrating the wind components u (a) and v (b) with the correction in the stratosphere that caused the wind to vanish at the top of the model. The coordinates are as in Fig. 1, except that the grid distance Δy is drawn wider.

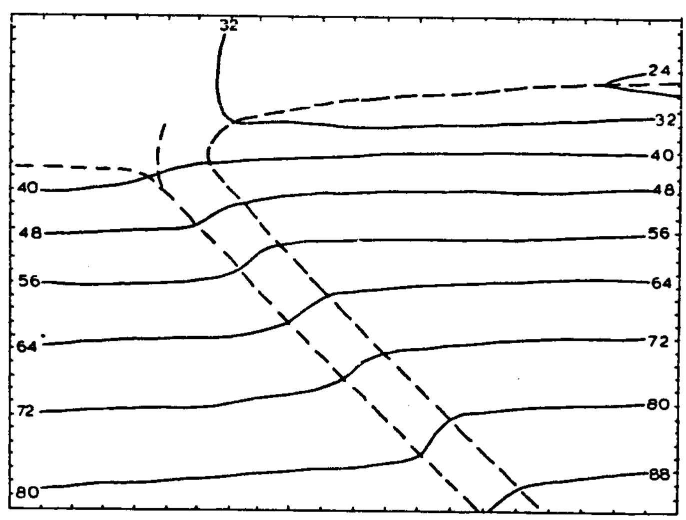

Fig. 6. Vertical and meridional section showing the smoothed and adjusted distribution of temperature. The frontal layer and the troppopause are outlined with dashed lines. The coordinates are as in Fig. 1.

Unfortunately, not all problems were resolved so far. The correction of temperature in the upper part of the model (8), where the correction reached in the upper troposphere, introduced spotwise hydrostatic instability. Correction of this static instability, in turn, had the effect to change the geostrophic wind such that values different from zero, albeit smaller than before, appeared near the top of the model.

Other changes in the model were introduced into the model too in order to diminish the contrast in temperature between the parts of the stratosphere across the line A-F of Fig. 1. The present state of the construction of initial conditions is shown in Figs. 6 and 7. The temperature in Fig. 6 shows a frontal layer that is more stable than the rest of the troposphere, but it is not an inversion as the first design in Fig. 1. The wind components (Fig. 7) again have values different from zero on the top, but much smaller than without adjustment of temperature with correction from (7).

The initial wind and height of the 500-mb surface in the model are illustrated in Fig. 8. These show the region used in LAMPS for an experimental full-scale forecast. The one-hour forecast results are shown in Fig. 9. The surface pressure (Fig. 9 a) already fell several millibars in the warm sector of the upper-level wave. There is already a noticeable high under the upper-level trough. Both these features indicate that the baroclinic wave is on the way to assume the typical atmospheric patterns, with cyclogenesis downstream from the trough aloft. There are still some formal problems with the forecasting model, as it can be seen in Fig. 9 b. Short waves near the jet stream are of numerical character, possibly due to spotwise thermal instability in the initial conditions. Larger numerical error grows on both ends of the jet stream, due to improperly handled boundary conditions. Since this run was made, the model has been improved. Cyclic boundary conditions have been introduced by Drs. Kalb and Perkey. A new experimental run has not yet been performed since a one-hour forecast takes about 10 h of computer time on the Perkin-Elmer computer. Besides, in the last week of my stay at NASA, Perkey and Kalb transferred the model to the Cray computer. New experiments will be made as soon as the model becomes operational on Cray and the above indicated adjustments in initial conditions are made.

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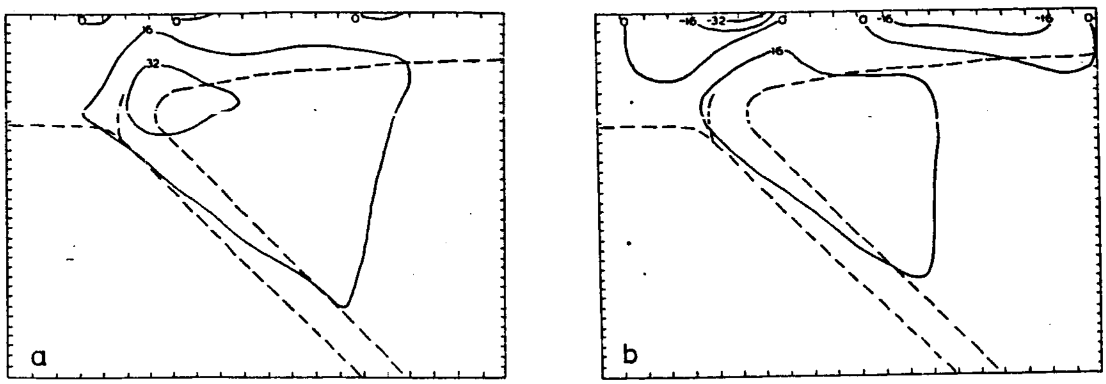

Fig. 7. Wind components u, v after adjustment for stable stratification (a and b, respectively). The coordinates are as in Fig. 1.

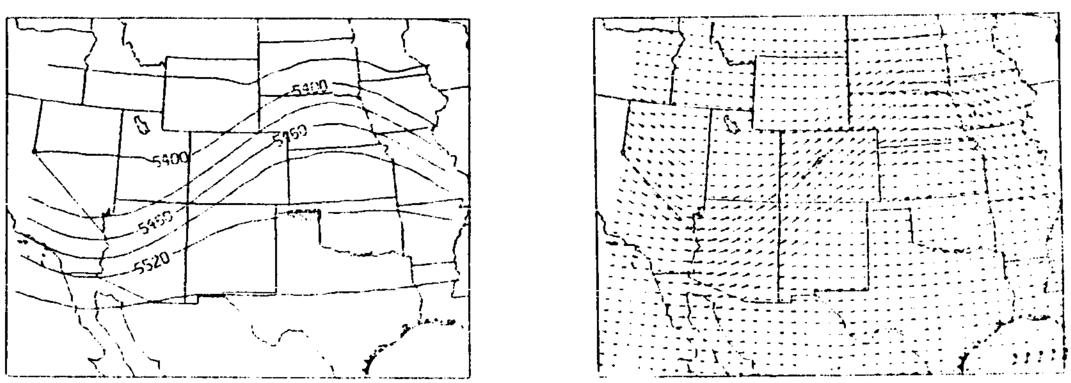

Fig. 8. Schematic initial conditions drawn over the region used in numerical simulation. Height of the 500-mb surface (a) and wind at 500 mb in relative units (b).

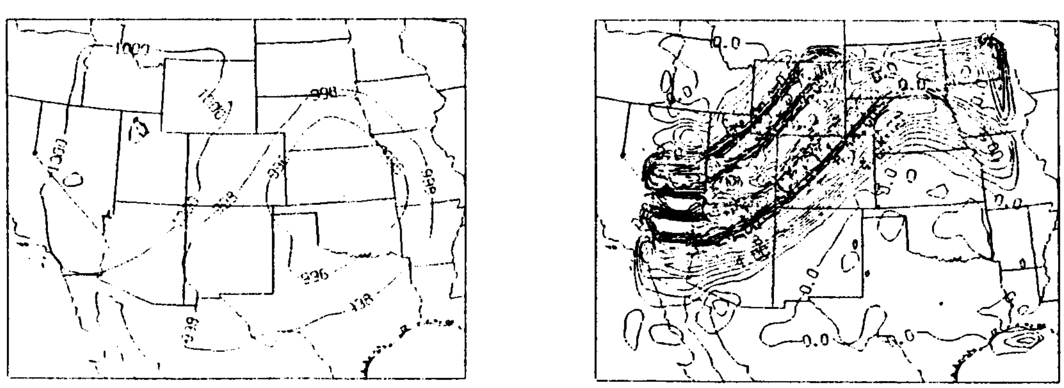

Fig. 9. One-hour forecast with the LAMPS model using the initial conditions from Fig. 8. Surface pressure in mb (a) and vertical motion in relative units (b).

CONCLUSIONS AND RECOMMENDATIONS

The experiments with analytically defined schematic initial conditions show that it is feasible to perform experiments with the LAMPS mesoscale model. The results of the one-hour forecast show development of surface high and low in expected positions in relation to the upper-level jet stream. An extension of such forecasting there fore can be expected to bring useful results.

There are still several shortcomings in the Fortran program. The correction in the stratosphere is presently done in three segments, separately for the polar part, part above the frontal zone and the part above the middle-latitude air mass. This should be changed into a uniform routine.

Further experiments should be done with different stability in the air masses and with presence of a stable layer in the lower part of the middle-latitude air mass where the low-level jet and accompanying transport of water vapor takes place.

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