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# Construction of dynamic-stochastic heat and water exchange climate models using the approach of J. Adem

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**RESUMEN**
Se discuten algunas aplicaciones del modelo de J. Adem a los procesos de intercambio de calor y de humedad. Se presenta el método de construcción del modelo dinámico-estocástico. Se propone usar este enfoque en las diversas investigaciones climáticas.

**PALABRAS CLAVE:** Modelos climáticos, termodinámica, procesos estocásticos, modelo de J. Adem

**ABSTRACT**
Applications of J. Adem's model to water and heat exchange processes are discussed and the method of dynamic-stochastic model construction is presented. This approach may be useful in different climate studies.

**KEY WORDS:** Climate modelling, thermodynamics, stochastic processes, model of J. Adem.

## INTRODUCTION

Considerable progress has been achieved in recent years in thermodynamic climate modelling. The basic principles of thermodynamic climate models were provided by J. Adem (1962, 1964). A decisive advance was the relative simplicity of his low-resolution grid thermodynamic models in comparison with models based on a purely dynamical approach. The computational task becomes rather easy and the computer time is not large. The specific algorithms may differ greatly but the final results should be similar for different versions.

Thermodynamic models have many applications. Earlier versions had been designed by J. Adem for seasonal and monthly mid-tropospheric temperature forecasts (Adem, 1964, 1965). Later, parameterization of heat advection in ocean and troposphere by normal and abnormal winds and currents was introduced (Adem, 1970a, 1970b). Incorporation of humidity was discussed (Adem, 1967, 1968). Successful precipitation forecasts have also been carried out (Adem and Donn, 1981). The advanced versions of the model have demonstrated their usefulness in paleoclimatic studies, annual cycle presentation, greenhouse effect, and climate sensitivity to orbital parameter perturbations (Adem, 1981, 1982; Adem and Garduño, 1982). The optimal estimates for some model parameters were discussed by Adem and Mendoza (1987).

We present some extensions of the model, applications to water and heat exchange studies and an incorporation into a dynamic-stochastic model.

## MODELLING OF SEA SURFACE TEMPERATURES (SST) AND EVAPORATION ANOMALIES

K. Hasselmann (1976) proposed an hypothesis explaining long-period climatic variability. He considered short-period random "weather" disturbances to be responsible for slow changes of climate. The whole climate system was divided into a rapidly varying "weather" system which excited a slowly responding "climate" system. The essential part of the "weather" is the atmosphere and that of the "climate" is the ocean.

The idea of applying a stochastic approach to energy balance climate model has been used by Lemke (1977) and others. For a deterministic basis a Budyko-Sellers zonally averaged model was used. Theoretical temperature variability spectra obtained by Lemke were in good agreement with observed spectra in the range of some dozen to hundreds of years.

We propose to incorporate the more advanced numerical thermodynamical features of Adem into a dynamic-stochastic climate model based mainly on Hasselmann's theory. Following this theory, the ocean mixed layer response to random weather fluctuations was studied.

The deterministic part of the model is represented by the heat energy conservation equation for the ocean mixed layer (Adem, 1970a). Weather forcing is considered as a joint influence of drift current anomalies and evaporation and sensible heat flux anomalies, both being functions of the mid-tropospheric temperature anomalies (T'). These anomalies are assumed to be a sequence of random uncorrelated fields following one another at time intervals of one month. T' is assumed to be normally distributed at every grid point. Correlation between neighbouring points decreases from 1 to 0 as their distance varies from 0° to 20°. Weather forcing is included by generating uncorrelated random normal variables by the Monte-Carlo method; these were smoothed by specially selected filters. Dispersion of the weather forcing estimation was taken from 500 mb temperature monthly dispersion charts.

The model has been integrated over 120 months. The main features of SST anomalies could be reproduced quite realistically. The spectra of model-generated SST series proved to be quite similar to the observed SST anomalies. Spectra estimates were obtained using a maximum entropy method (Ulrich and Bishop, 1975). The principal feature of both synthetic and observed SST anomalies is that they are satisfactorily described by stationary first-order autoregressive models (Fig. 1). This result matches the results from 155 observed time series for the Atlantic Ocean (Fig. 2).

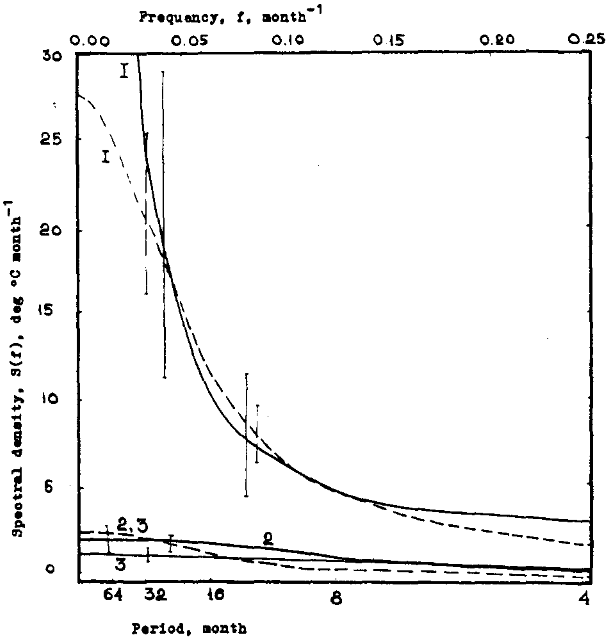



Fig. 1. Sea surface temperature monthly anomalies spectra : 1 - 42°30' N, 57°30' W; 2 - 42°30' N, 17°30' W ; 3 - 22°30' N, 37°30' W; (dashed line - model results; continuous line - observed ones; vertical lines - 95% confidence interval). (Dobrovolski, 1983).

Abnormal drift currents contribute significantly to SST anomalies variability in contrast with stationary drift currents. Mean-square latent heat anomalies computed for the Northern Hemisphere winter do not match exactly but are in fair agreement with the observations (Fig. 3).

The anomalous monthly latent heat flux acts as a negative feedback and not as a generating mechanism in the SST forcing processes.

As a first approximation, numerical experiments with this version of the model successfully illustrate the possibility of expressing simplified weather forcing through the ocean mixed layer excitation by abnormal drift currents. Thus some aspects of SST dynamic anomalies may be studied with no atmospheric detailed description for time scales of several months to several years.

## WATER EXCHANGE PROCESSES IN THE ATMOSPHERE

Some remarkable correlations between water vapour transport and total content in the layer of 0-9 km (from the surface to 300 mb), and the conditions on the 850 mb isobaric surface have been reported, to be included in advanced parameterization expressions.

Jarosh (1986) showed that total zonal and meridional water vapour transport monthly anomalies ($F'_x$ and $F'_y$) and the same anomalies on the 850 mb isobaric surface ($F'_{x850}$ and $F'_{y850}$) were essentially correlated. Correlation coefficients were higher than 0.85 for zonal transport and 0.88 for meridional transport. The mean estimate for both is 0.94. Data of 41 meteorological stations covering the entire U.S.S.R. were used. Differences of estimates for the whole set of stations were found to be negligible. It appeared to be possible to express $F'_x$ and $F'_y$ as a linear regression on $F'_{x850}$ and $F'_{y850}$.

$$F'_x = k_x F'_{x850} \tag{1}$$

$$F'_y = k_y F'_{y850} \tag{2}$$

where $k_x = 3.7$ kg m g$^{-1}$ for zonal transport;
$k_y = 3.6$ kg m g$^{-1}$ for meridional transport.

Standard deviation for $k_x$ and $k_y$ is 0.31 kg m g$^{-1}$.

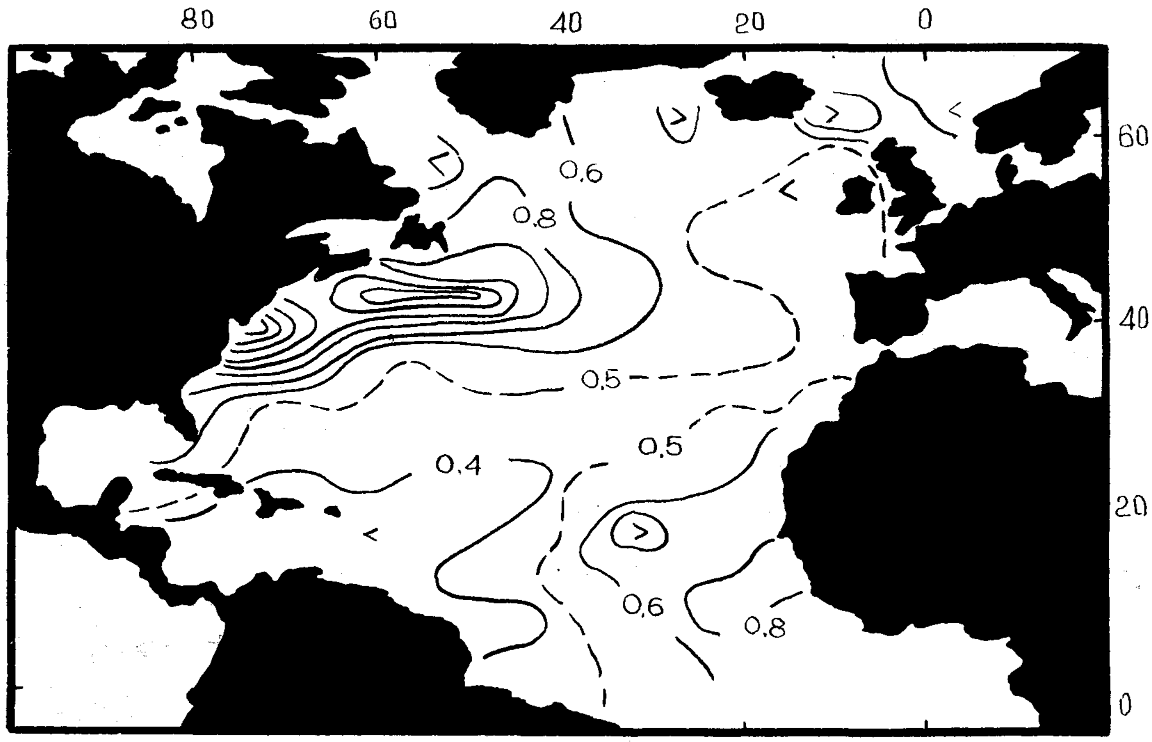
Another useful result is the rather high correlation between total water vapour content anomalies (q') and the air absolute humidity anomalies (a') on the 850 mb surface. The approximate expression is:

$$q' = k_q a' \tag{3}$$

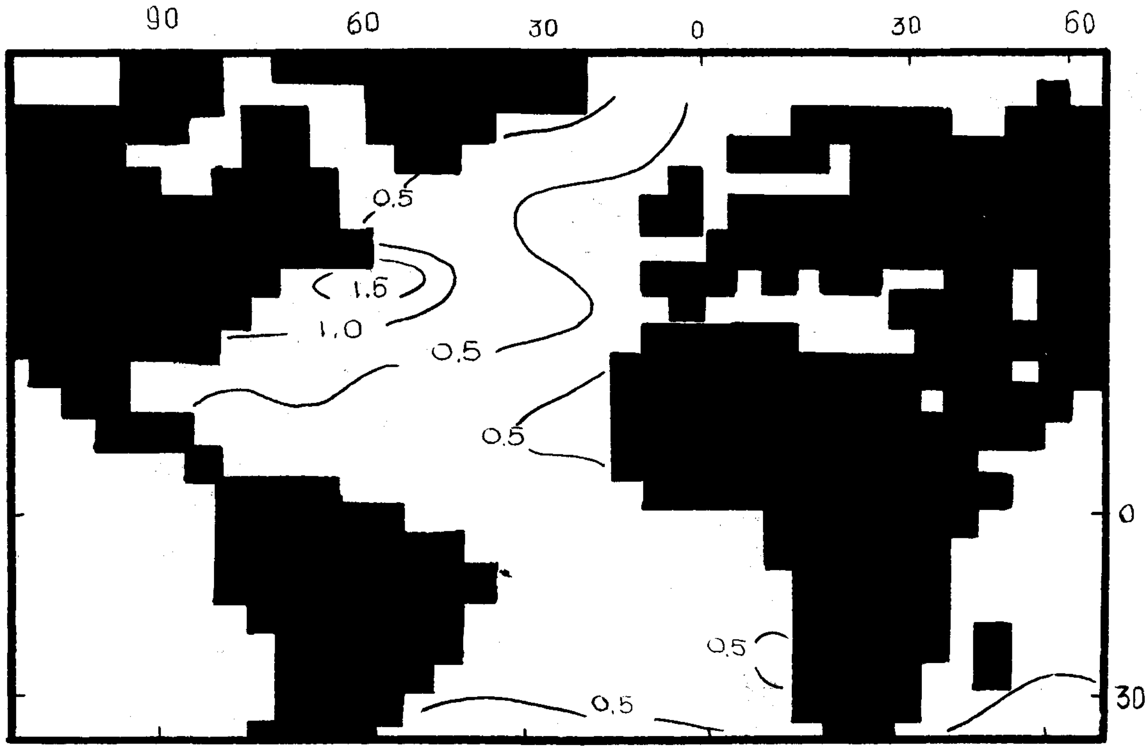
where $k_q = 3.4$ mm m$^3$ g$^{-1}$. The mean correlation coefficient between q' and a' was found to be 0.92.

The spatial variance of $k_q$ is greater than that of $k_x$ and $k_y$. Thus the error in q' computed from (3) is greater than for $F'_x$ and $F'_y$ computed from (1) and (2).

These results allow the water vapour exchange process to be incorporated into a numerical thermodynamical and dynamic-stochastic model for the U.S.S.R. and even tentatively for other extratropical regions.




A




B

Fig. 2. Sea surface temperature monthly anomalies standard deviation:
A - results of autoregressive analysis of 155 time series; B - model results.

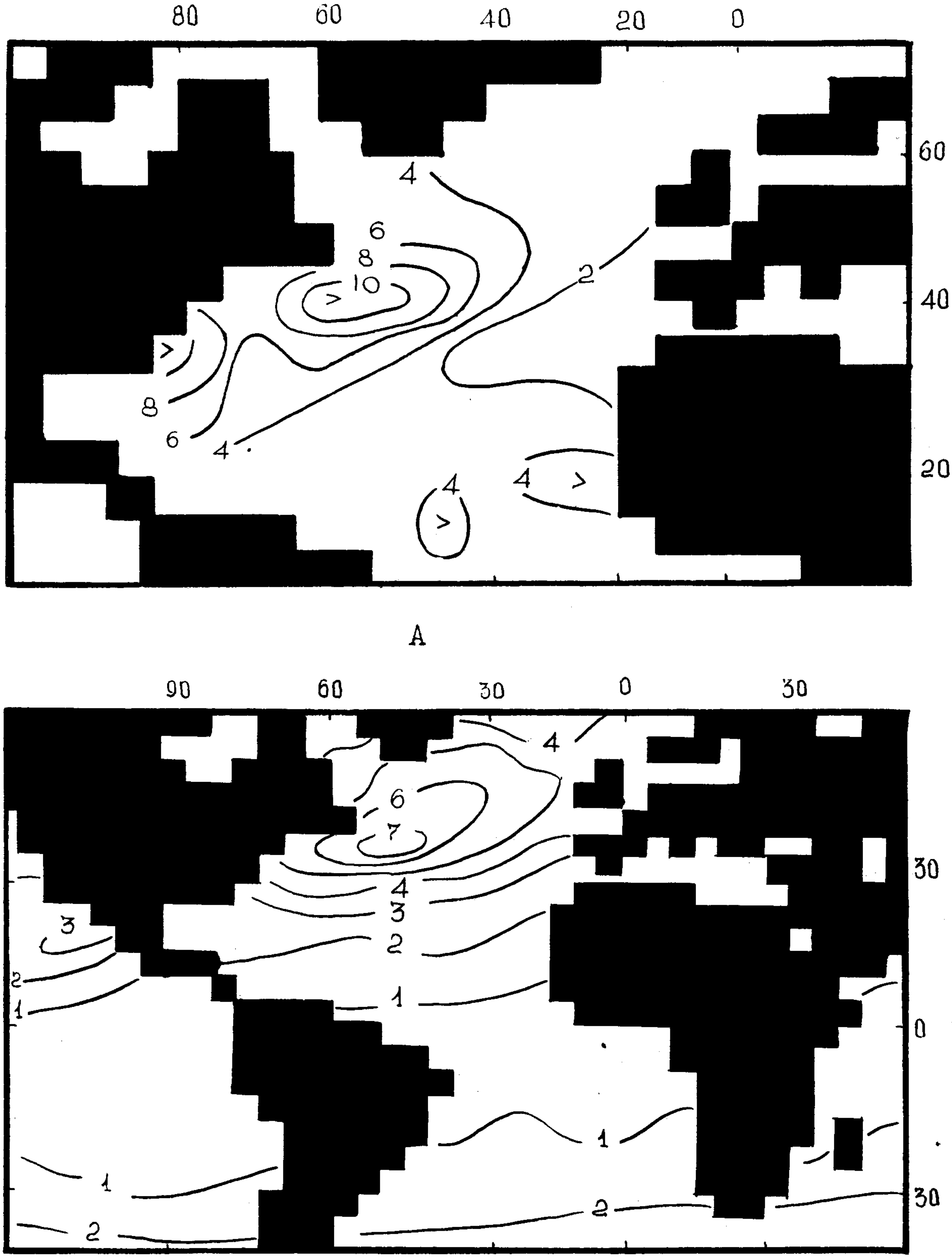



Fig. 3. Latent heat flux monthly anomalies (MJ $m^{-2}$ $day^{-1}$):
A - observed for February, Northern Atlantic: B - model results for Northern Hemisphere Winter.

Zonal and meridional water vapour transport on any isobaric surface may be written as:

$$F'_x = a\,V_x \tag{4}$$

$$F'_y = a\,V_y \tag{5}$$

where a is the absolute humidity, and $V_x$ and $V_y$ are the geostrophic wind components.

Following J. Adem's approach, all terms will be represented as a sum of "normal" and "abnormal" parts; thus we obtain

$$F'_x = \bar{a}\,V'_x + a'\bar{V}_x \tag{6}$$

$$F'_y = \bar{a}\,V'_y + a'\bar{V}_y \tag{7}$$

where the bars designate normal climatological quantities.

$V_x$, $V_y$, $V'_x$, $V'_y$ can be easily estimated as zonal and meridional components of the geostrophic wind and its anomalies, dependent on the horizontal gradients of the isobaric surface height monthly anomalies (850 mb surface in this case), which in turn may be expressed through the 850 mb temperature monthly anomalies:

$$H'_{850} = \frac{g}{\beta}\,(1 - k^{*}\exp(-R\beta/g))\,T'_{850} \tag{8}$$

where k* is a coefficient dependent on the height of the upper boundary of the model tropospheric layer. In our case we took k* = 6/17.

Now it is easy to obtain the gradients of $H'_{850}$:

$$\frac{\partial H'_{850}}{\partial y} = \frac{g}{\beta}\,(1 - k^{*}\exp(-R\beta/g))\,\frac{\partial T'_{850}}{\partial y} \tag{9}$$

$$\frac{\partial H'_{850}}{\partial x} = \frac{g}{\beta}\,(1 - k^{*}\exp(-R\beta/g))\,\frac{\partial T_{850}}{\partial x} \tag{10}$$

After transformation of (9) and (10) into finite differentials they are substituted into the following expressions for the anomalies of geostrophic wind components:

$$V'_x = -\frac{1}{f}\,\frac{\Delta H'}{\Delta y} \tag{11}$$

$$V'_y = \frac{1}{f}\,\frac{\Delta H'}{\Delta x} \tag{12}$$

In (8) - (12) we use the following symbols: g, acceleration of gravity; β, vertical lapse rate; R, dry air gas constant; f, Coriolis coefficient dependent on latitude; and the meaning of k* as explained above.

Finally, we proceed as described in the previous section to simulate the process of water vapour transport and total water vapour content anomalies. The only unknown is the 850 mb temperature monthly anomaly ($T'_{850}$). It can be obtained from the idea of atmospheric "weather" forcing as random uncorrelated fields of monthly weather temperature anomalies, but on the 600 mb surface. It is not correct to use white noise fields of the 850 mb surface directly because of the possible influence of the underlying Earth surface on a monthly average time scale. The height of 600 mb is chosen because it is close to the mid-tropospheric height (4.2 km approximately). The $T'_{600}$ field is assumed to be normally distributed at every grid point. Space correlation is assumed to be isotropic, and the space correlation coefficient used is the same as in the previous section of this paper. $T'_{850}$ can be easily obtained from $T'_{600}$. The Monte-Carlo method is used to simulate $H'_{600}$ fields. Dispersions of $H'_{600}$ are taken from charts. A resolution of 5° for latitude and 10° for longitude is used in this model version.

The methodology of air absolute humidity parameterization is discussed in Adem (1967). We also use it in our model version, but with some reservation concerning cloudiness. The anomalies of total cloud cover are assumed to be uncorrelated with $T'_{850}$, because earlier experiments have shown that they are practically independent.

The air temperature vertical profile is assumed to be linear. Vertical lapse rate spatial and time differences have been found to contribute negligibly to the final result. Therefore, the vertical lapse rate is taken to be constant. Computed $F'_x$, $F'_y$ and q' fields are shown in Fig. 4 - 6. Notice that the main features of the computed fields are quite close to the observed ones, with the exception of the arid zones.

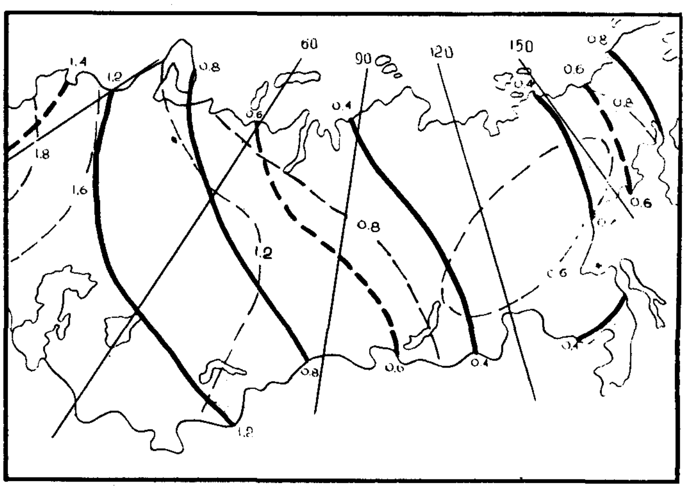



Fig. 4. Total water vapour content standard deviation, January (thick line - computed on model; thin line - observed ones), mm.

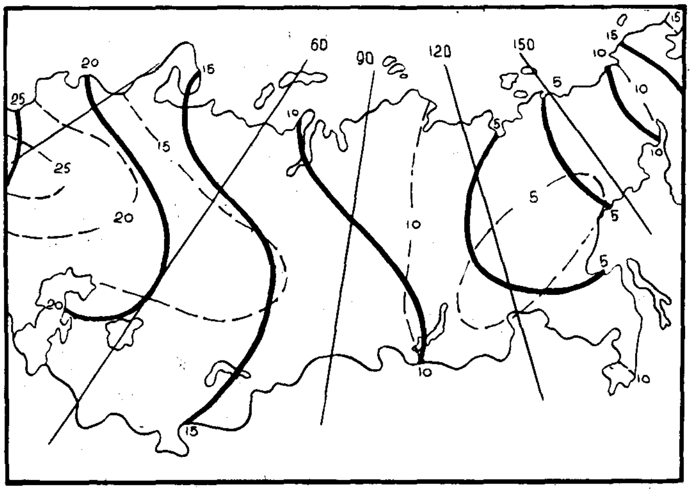



Fig. 5. Water vapour total zonal flux standard deviation, January (thick line - computed on model; thin line - observed one), kg $m^{-1}s^{-1}$.

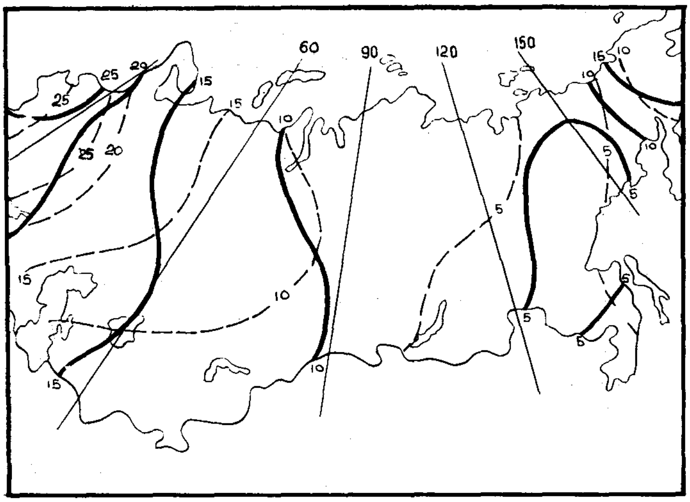



Fig. 6. Water vapour total meridional flux standard deviation, January (thick line - computed on model; thin line - observed one), $kg^{-1}s^{-1}$.

## PARAMETERIZATION OF WATER EXCHANGE PROCESSES IN THERMODYNAMIC MODELS: A NEW APPROACH

The methodology for the computation of $F'_x$, $F'_y$ and q' discussed in the previous section can be used in a numerical thermodynamical model of climate. If the condition of water vapour conservation is assumed (Adem and Garduño, 1984), then:

$$G_3 - G_5 = L E \tag{13}$$

where $G_3$ is the heat lost by evaporation at the surface of the Earth; $G_5$ is the heat gained from the water vapour condensation in the cloud layer; L is the specific heat of condensation; E is the term describing water vapour transport and total content in the considered tropospheric layer.

Taking into account the results of the previous section one can obtain an equation for E.

$$E = k_x F'_{x850} + k_y F'_{y850} + k_q q' \tag{14}$$

where $F'_{x850}$, $F'_{y850}$ and q' have been shown to be functions of $T'_{850}$, which is obtained from the mid-tropospheric temperature monthly anomaly (T') as described above. This procedure is useful as an alternative description of either $G_5$ or $G_3$. Another reason for using this method is the possibility of incorporating precipitation into the model.

It is the authors' opinion that the way of incorporating precipitation should be a statistical one. Monthly amounts of precipitation ("normals") and precipitation monthly anomalies should be expressed by parameters computed from the model itself by simple regression equations.

Table 1

Correlation of mean monthly precipitation with different pairs of atmospheric parameters (predictors) in three regions of the U.S.S.R.

| Region | Predictors | Multiple correlation coefficient |
|---|---|---|
| 1 | 850 mb Q<br>850 mb T° | 0.85 |
| | 850 mb r(%)<br>850 mb T° | 0.83 |
| | q<br>850 mb T° | 0.84 |
| 2 | 850 mb Q<br>850 mb T° | 0.69 |
| | 850 mb r(%)<br>850 mb T° | 0.80 |
| | q<br>850 mb T° | 0.77 |
| 3 | 850 mb r(%)<br>850 mb T° | 0.67 |
| | 850 mb Q<br>850 mb T° | 0.54 |

Q = absolute humidity, T = air temperature, r = relative humidity, q = total water vapour content.

Preliminary statistical tests have shown this approach to be well-founded. Two- and three-dimensional low-order autoregressive models were fitted to time series of meteorological data and to monthly means observed at some tens of meteorological stations in three different climatic regions on the U.S.S.R. The first region (Byelorussia) is located in a rather humid area, the second one (Mid-Volga Region) has a continental climate, and the third one (Turkmenia) has an arid climate. Mean monthly amounts of precipitation were taken as predictands and different pairs of atmospheric parameters as predictors. Some results of this preliminary analysis are presented in Table 1.

These tests, of course, are not definitive. A thorough and comprehensive analysis is needed to study the relation between precipitation monthly anomalies and anomalies of other atmospheric parameters. The annual cycle of precipitation can be reproduced, though fairly approximately. One can take advantage of the fact that the processes taking place on the 850 mb isobaric surface govern in some way the processes in the whole tropospheric layer. Thus one need not consider a multilayer troposphere: a single layer is sufficient.

## CONCLUSIONS

A numerical thermodynamic model can be a reliable, inexpensive and natural model for different types of climate studies.

We have shown how one can profitably incorporate a thermodynamical model into a general dynamic-stochastic model. If we consider the monthly averages to be elementary intervals of climatic variability it becomes unnecessary to describe the dynamics in detail. The best argument for such an approach is still Hasselmann´s two-scale division hypothesis, because the monthly interval is considered the climatic variability lower boundary (GARP, 1977). The minimum in the atmospheric variability spectrum corresponds to the monthly interval. But when the armosphere is viewed as separate from the rest of the climatic system, the atmospheric variability at a time scale of one month can be approximated as a white-noise excitation; then weather forcing can be described stochastically, discarding the dynamical and thermodynamical equations. As for the long-period climatic variability, it can be determined in many extratropical regions in terms of the local heat and water conservation equations.

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