THE METEREOLOGICAL EFFECT OF THE CHANGES IN SURFACE ALBEDO AND MOISTURE

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ABSTRACT

A three dimensional numerical model is applied to study the mesoscale effect of the albedo contrast between the Negev and Sinai areas. The model consists of the three dimensional hydrostatic equations of motion, the equation of continuity, the first law of thermodynamics for an ideal gas and a conservation law for water vapor. A detailed boundary layer parameterization, a surface heat budget equation and atmospheric cooling/heating due to long and short wave radiation are incorporated in the model.

It is shown that the results are very sensitive to the moisture content in the soil. Under dry soil condition the predicted surface temperatures in the Negev are significantly higher, while when a relative humidity of 5% was used for the Negev no ground temperature gradients were developed.

INTRODUCTION

The effect of the destruction of vegetation by overgrazing on the albedo has been studied by Otterman (1974). His findings show that the dark less-grazed Negev has an average albedo of .4. Otterman claims that this difference results in a decreased surface temperature in the Sinai area relative to the Negev, which in turn decreases the sensible and latent heat fluxes to the atmosphere, thereby possibly suppressing convective shower formation. On the other hand, Charney et al. (1977) suggest that an increase in plant cover causes an increase in evapotranspiration, in addition to the decrease in albedo, which implies surface temperatures may not be higher over the Negev. If, however, as Otterman et al. (1976) claim, the large contrast in albedo exists across the Negev-Sinai line, without a change in evapotranspiration (the Negev's low albedo according to Otterman is due to an appreciable amount of dead plant material), a significant horizontal temperature gradient might exist.

In this article we present results of a three-dimensional model (Mahrer and Pielke, 1976, 1977) for the following cases:

i) Control experiment (.2 albedo and no wetness for both the Negev and the Sinai areas).

ii) An albedo of .2 for the Negev and .4 for the Sinai desert (no wetness).

iii) Same as experiment (ii) but with 5% wetness for the Negev ground.

OUTLINE OF THE MODEL

The hydrostatic three-dimensional primitive equations of motion, heat, moisture and continuity are converted to a terrain following coordinate system (x,y,z^*,t) using the transformation:

$$Z^* = \bar{s}\,\frac{Z-Z_G}{s-Z_G}$$

where $\bar{s}$ is the initial height of the material surface top, s is the material surface top (s is a function of x,y and t) and Z_G is the height of the topography.

The boundary layer formulation is based on Businger (1973), while the height of the planetary boundary layer is calculated according to Deardorff's (1974) formula. For more details see Pielke and Mahrer (1975). The surface temperature is evaluated from a surface heat budget equation. Radiative heating/cooling by longwave and shortwave fluxes are incorporated into the model (including the terrain effects, such as slope and slope azimuth). A full discussion is given in Mahrer and Pielke (1977).

Although the model is capable of dealing with topography, topography was neglected in these experiments in order to investigate the other aspects of this study.

INITIAL AND BOUNDARY CONDITIONS

i) Initial conditions (t=0)

$$T_{land} = T_{sea} = 26°C, \; q_{land} = 11 \quad q_{water} = 21 \text{ g/kg} \left\{ \begin{array}{l} \text{at the ground} \\ \text{surface at the} \\ \text{initial time} \end{array} \right.$$

$$U_g = 2ms^{-1} \quad V_g = 1ms^{-1}$$

$$\frac{\partial T}{\partial Z} = -0.65°C/100m$$

$$\frac{\partial q}{\partial z} = 1.5 \text{ g kg}^{-1} \text{ km}^{-1}$$

ii) Boundary conditions

$$\left. \begin{array}{l} q_G = F_W \, q_{G}, \text{ saturated} + (1 - F_W) q(1) \\ u = v = w^* = 0 \end{array} \right\} \quad z^*=0$$

$$\left. \begin{array}{l} u = U_g, \; v = V_g, \; w^* = 0 \\ \pi = \overline{\pi} + g\,(\overline{s}-s)/(\theta\ (s) + \frac{1}{2}\gamma_{top}(s-\overline{s}) \\ q(s) = 2 \text{ g kg}^{-1} \quad \theta(s) = \text{constant} \end{array} \right\} \quad z^*=\overline{s}$$

Here F2 is a measure of surface wetness and γ_{top} is the potential temperature lapse rate at the top of the model.

At the lateral boundaries we use zero gradient boundary condition on all the variables except that w* = 0.

NUMERICAL ASPECTS

A staggered horizontal and vertical grid structure as described in Pielke (1974) was used. In the horizontal we used 30 x 30 grid points with a constant grid interval of 8 km. In the vertical the atmosphere is divided into 12 levels with heights of 0, 15, 100, 300, 600, 1200, 2000, 3000, 4000, 5000 and 6000 m in which the velocity components and pressure are defined. The potential temperature and the specific humidity are defined at intermediate levels. The time step of the integration was 90 seconds.

The numerical scheme used for the horizontal and vertical advective terms is the upstream interpolation by cubic splines developed by Purnell (1976). The details of this scheme are reported in Mahrer and Pielke (1978).

TABLE 1: INPUT PARAMETERS

$k_s = 0.03\ cm^{-2}s^{-1}$	$\delta = 20°$
$\rho_s = 1.5\ g\ cm^{-3}$	$\sigma = 1.38\ 10^{-12} cal\ cm^{-2}k^{-4}s^{-1}$
$c_s = 0.31\ cal\ g^{-1}K^{-1}$	
$z_0 = 1\ cm$	$k_0 = 0.35$
$\phi = 30°$	$T_0 = 299°K$
$U_g = 2\ ms^{-1}$	$V_g = 1\ ms^{-1}$

Experiment 1

In this experiment, which will be referred to as the control experiment, we used a constant albedo of 0.2 for both the Sinai and the Negev areas, and F_W was set to zero (i.e., a dry surface).

Figure 1 shows the predicted surface temperature for 1300 LST. Due to the cooling effect of the sea breeze, the temperature is lowest near the coastal zone and increases as we move inland as was suggested by Neumann and Mahrer (1971).

Figure 2 shows the vertical velocity at 1500 LST. The figure depicts two major cells, one with descending motion near the coast and the other one with rising velocities further inland.

Figure 3 shows the horizontal wind at 1500 LST, the strongest winds are found at a distance of 56 km inland near the maximum development of the vertical velocities.

Experiment 2

In this experiment albedo values of 0.4 and 0.2 were used as representative for the Sinai and the Negev areas, respectively. This experiment was run with $F_W = 0$. Figure 4 shows the surface temperature at 1300 LST. Due to the lower albedo of the Negev, a sharp temperature gradient is developed along the boundary line. The temperature differences are larger inland and reach a maximum value of 3-4 degrees. This temperature contrast develops its own circulation and produces a wind direction change across the albedo line (Figure 5). As a result, an increased cell of upward velocities is developed over the Negev zone, with downward motion over part of the Sinai desert (Figure 6). Another feature of this experiment are the lighter horizontal sea breeze winds over the Sinai desert, which are a result of the lower temperatures there.

Experiment 3

This experiment is identical to experiment 2, except that a wetness parameter of 0.05 (5%) was prescribed for the Negev area while the Sinai desert was held dry. The resultant surface temperatures are depicted in Figure 7. It is apparent from the figure that the results are very sensitive to the soil wetness, and a value as small as 5% is enough to completely mask the albedo influences (compare with Fig. 4). This result is consistent with Gannon's (1976) finding that soil moisture applied in a spatially uniform manner exerts a greater influence on sea breeze circulations than changes in surface albedo.

CONCLUSION

The results of the model experiments presented here suggest that the effect of differential albedo alone is to create lower temperatures where the reflection of the short wave radiation is greatest. The absorption of short wave radiation by the lower atmosphere after it is reflected is not sufficient to compensate for the higher albedo, at least for the relatively dry air assumed here. The resultant differential temperature gradient interacts with the sea-land contrast and produces significant low-level convergence along the dividing zone between the lower and higher albedo areas.

When the area with lower albedo is assumed to have even a low level of surface moisture, however, the effect of the differential albedo is masked. In this case, the cooling through evaporation is capable of dominating the contribution to heating due to the reduced albedo. Work needs to be undertaken in the Negev-Sinai region, as well as in other desert locations, in order to determine the specific surface soil moisture characteristics.

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LIST OF SYMBOLS

C_p	specific heat at constant pressure
C_s	specific heat of soil
g	acceleration of gravity
Ko	Von Karman constant
Ks	soil thermal diffusivity
p	pressure
Poo	reference pressure
q	specific humidity
r	gas constant for dry air
s	material surface top of the model
$\bar{s}$	initial height of the material surface
t	time
T	temperature
To	sea temperature
u,v,w	east-west, north-south and vertical wind component
U_g, V_g	east-west and north-south geostrophic wind
w*	vertical (z*) component of velocity
x,y,z	Cartesian coordinates
z*	vertical terrain following coordinate
z_o	roughness parameter
s	density of soil
$\emptyset$	latitude
δ	solar declination
σ	Stefan-Boltzmann constant
π	Exner's function = $c_p \left(\frac{p}{Poo}\right)^{R/C_p}$
θ	potential temperature

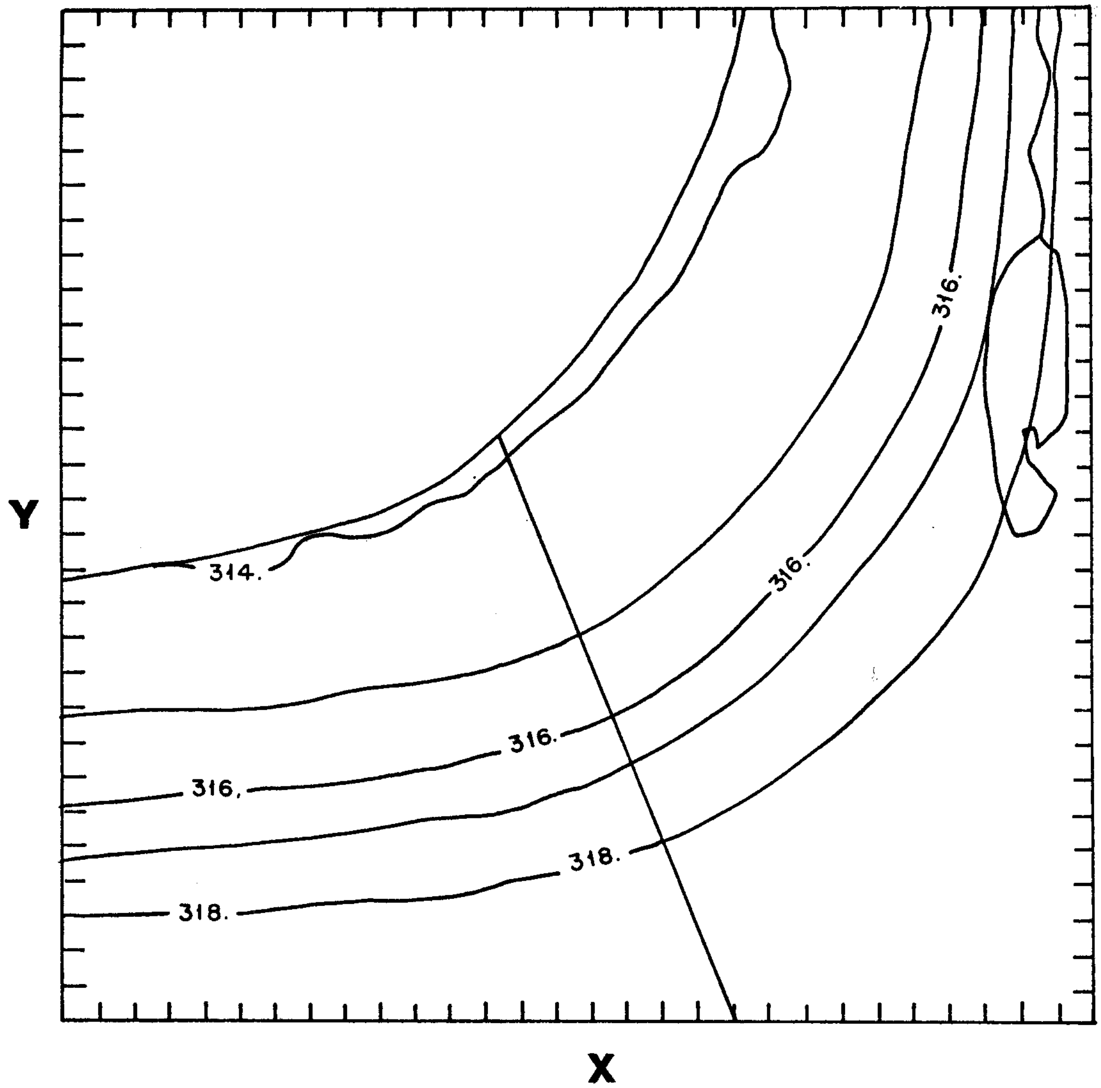

Figure 1. Predicted surface temperature at 1300 LST (control experiment).

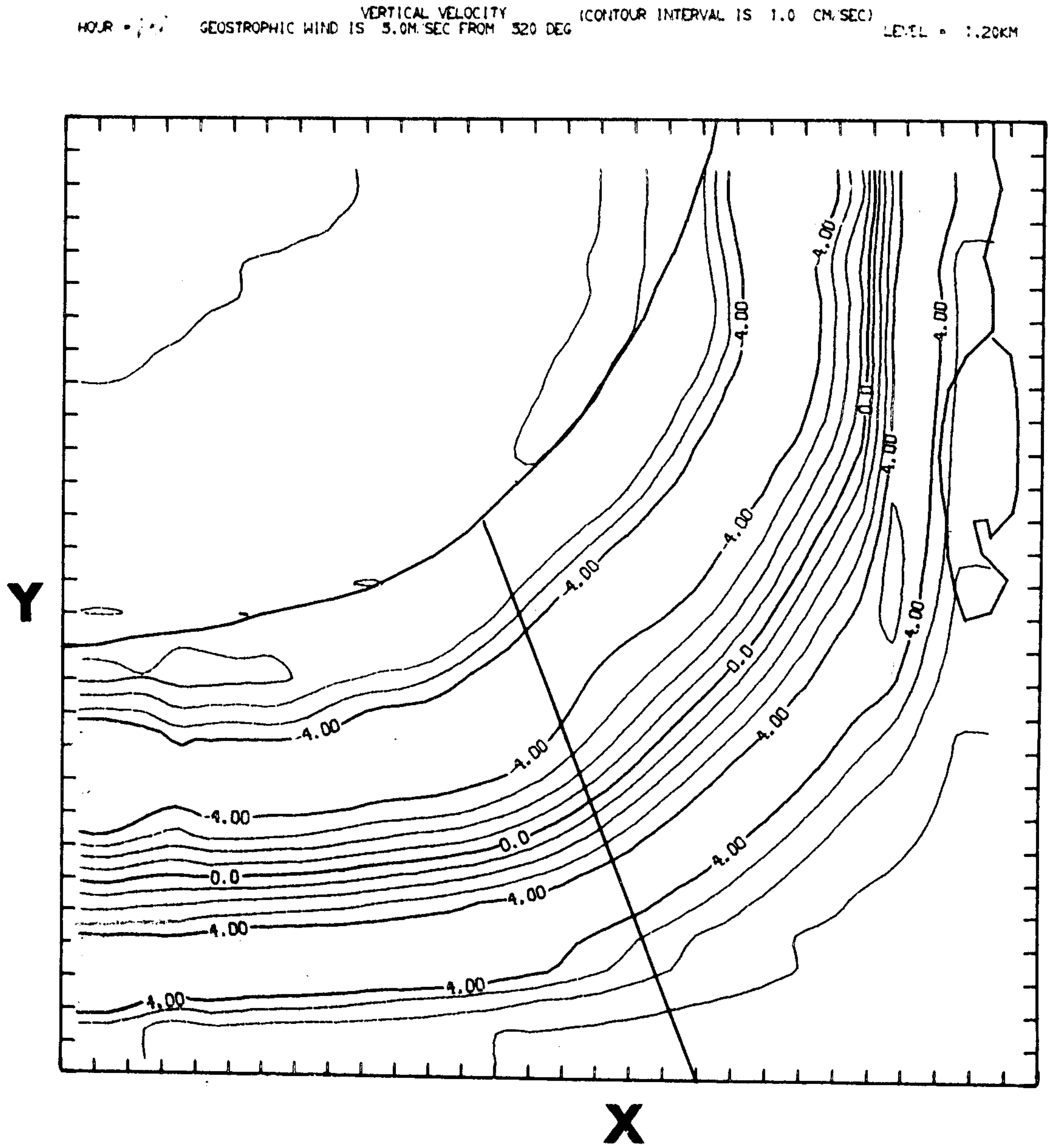

Figure 2. Vertical velocity field at a height of 1.2 km, for 1500 LST. (control experiment).

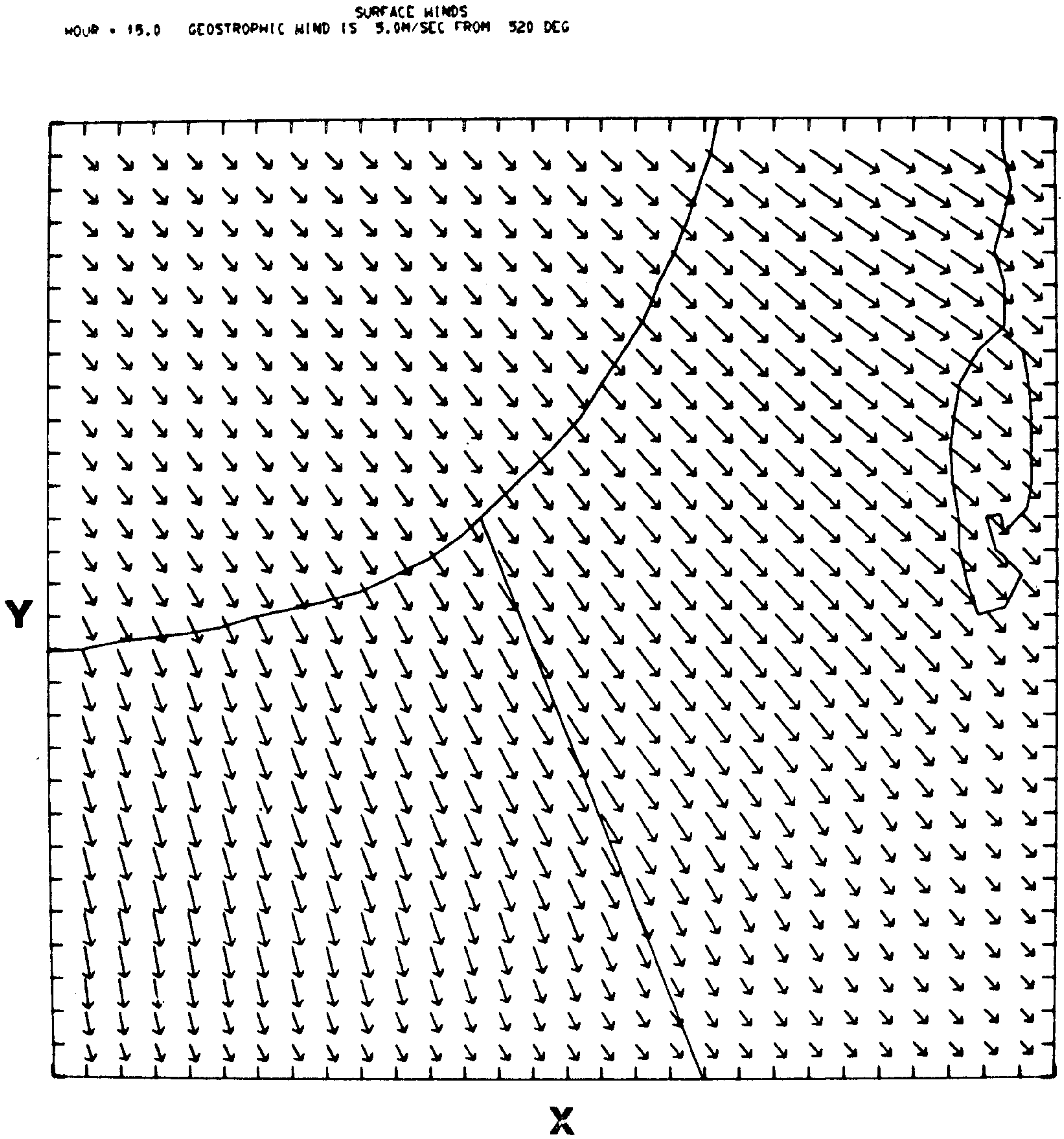

Figure 3. Horizontal wind field at a height of 5 m for 1500 LST (control experiment).

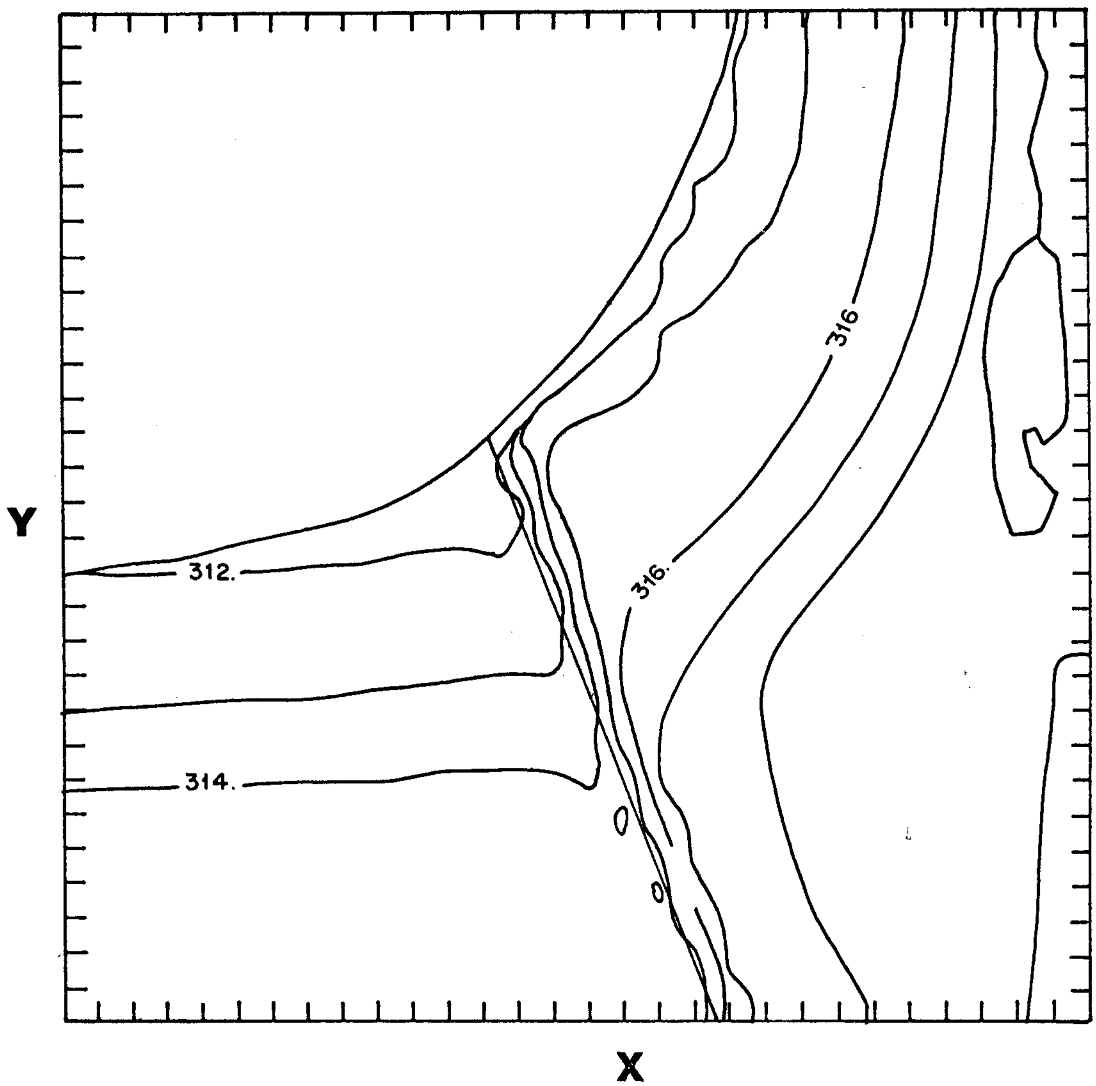

Figure 4. Predicted surface temperature at 1300 LST (experiment 2).

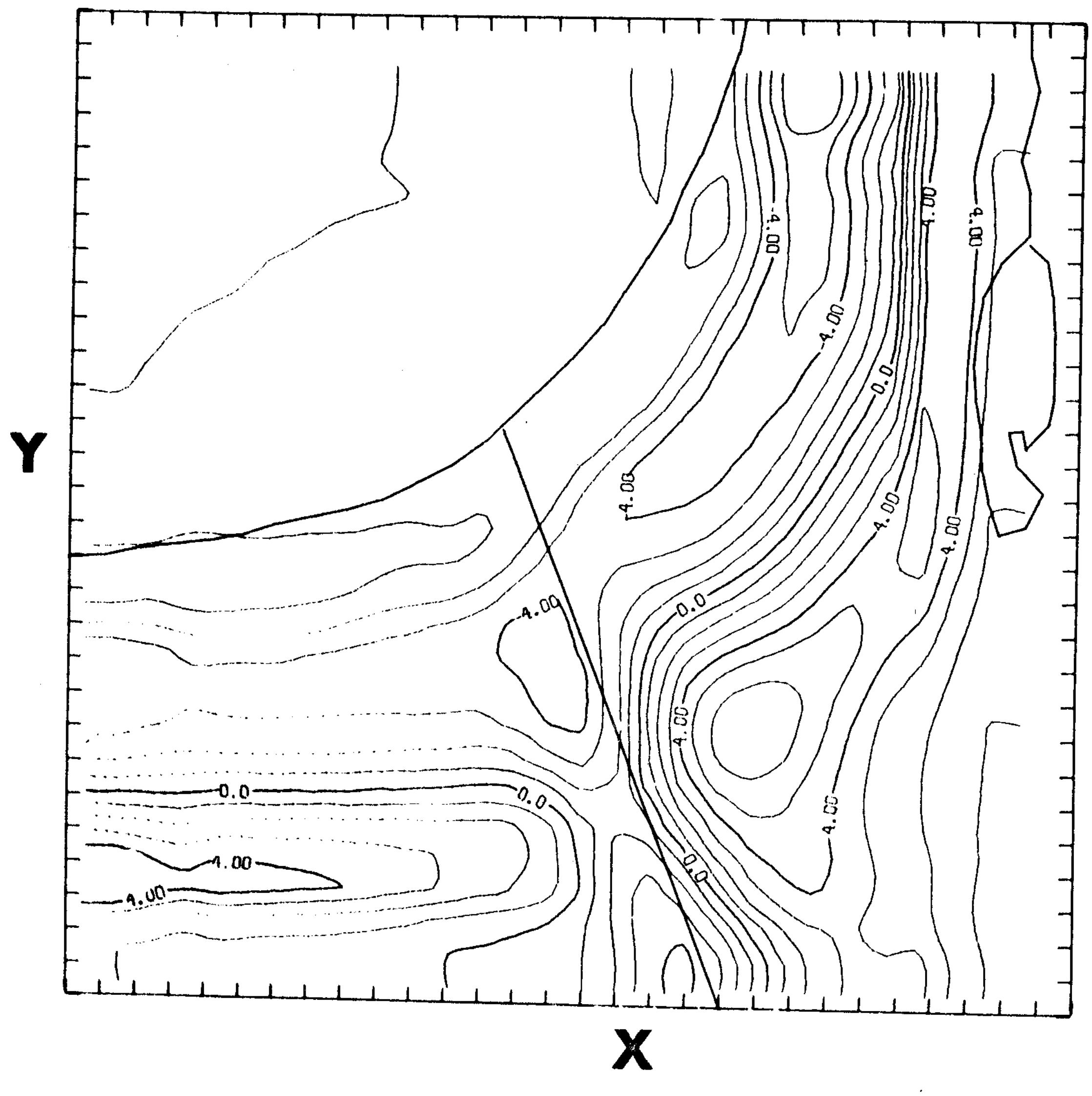

Figure 5. Vertical velocit; field at a height of 1.2 km for 1500 LST (experiment 2).

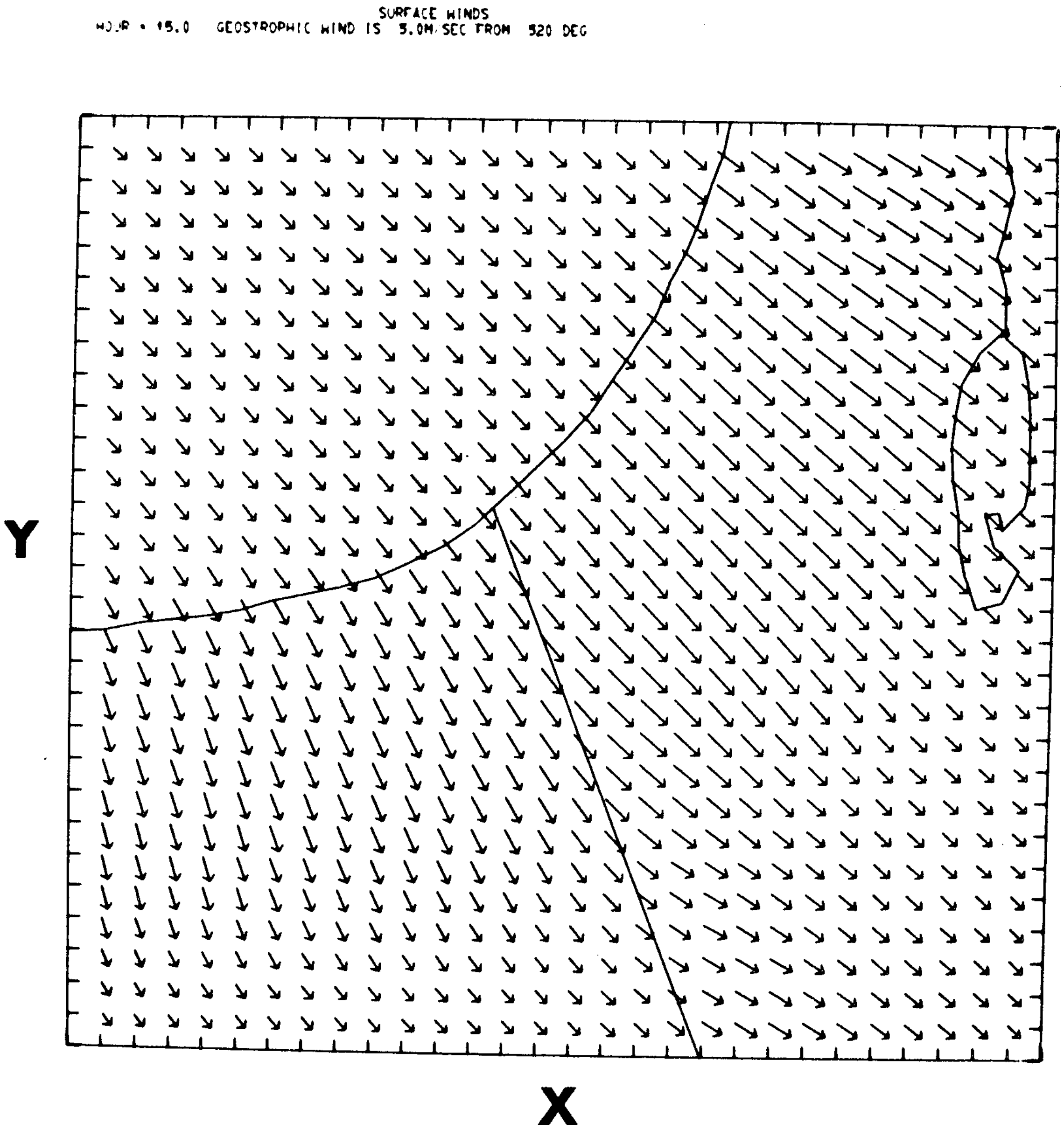

Figure 6. Horizontal wind field at a height of 5 m for 1500 LST (experiment 2).

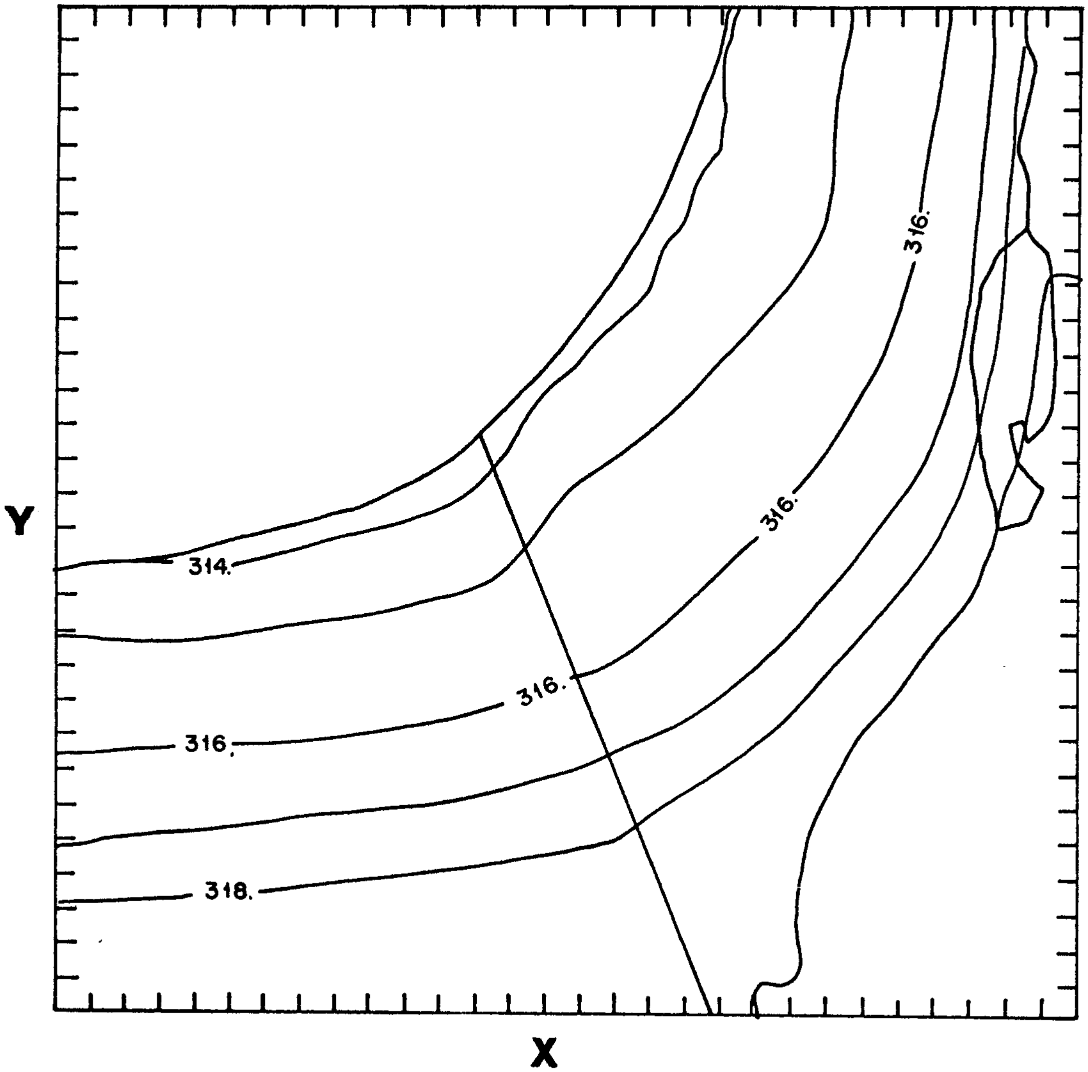

Figure 7. Predicted surface temperature at 1300 LST (experiment 3).

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