A CASE STUDY OF AIR MASS TRANSFORMATION

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1. Introduction

This lecture deals with the comparison of several parameterisation schemes applied to an actual weather situation. A great deal of the material presented here is taken from a report written by Michael Tiedtke (1977). The case selected for this study was first described by Økland (1976). It is an example of rapid heating and moistening of arctic air as it travels southward over the Norwegian sea.

The parameterisation schemes used in the experiments are combinations of the following :

for the boundary layer processes :
- GFDL method
- "dry diffusion" method
- proposed ECMWF method

for the cloud convection:
- Manabe method
- Kuo (65) method
- ECMWF method

These various methods will be described briefly in Section 3 of this lecture.

2. The weather situation

The weather situation chosen for this study is a typical case of transformation of arctic air. Figure 1 is a composite picture of two figures from Økland's (1976) report. It shows the sea level isobars (solid lines) and the sea surface temperature (dashed lines) for 21 November, 1975 over the Norwegian sea. The surface observations (12 GMT) at the weather stations in the area are shown and the extent of the ice pack (hatched area) is indicated.

It can be seen that the cold arctic air leaving the ice covered Spitzbergen region travels perpendicularly to the sea surface isotherms for a considerable distance. Thus a considerable temperature difference between the sea and the air is maintained while the air is warming up. This unstable situation produces intense fluxes of heat and water vapour at the surface.

The weather pattern in the area of the Norwegian sea was, at that time, fairly stationary. Furthermore, there was little vertical wind shear, as indicated in Fig. 2 by the soundings for Björnöya (21 November 1975, 00 GMT) and for Bodö (twelve hours later). Hence it seems reasonable to assume that an air mass leaving the ice near Björnöya would end up some kilometres west of Bodö about 18 hrs. later. The soundings shown on Fig. 2 are assumed to be characteristic of the structure of the air mass at the beginning and the end of this trajectory. If we disregard any effect of large scale convergence and radiation, the differences between the two soundings are then entirely due to the fluxes of heat and water vapour through the surface.

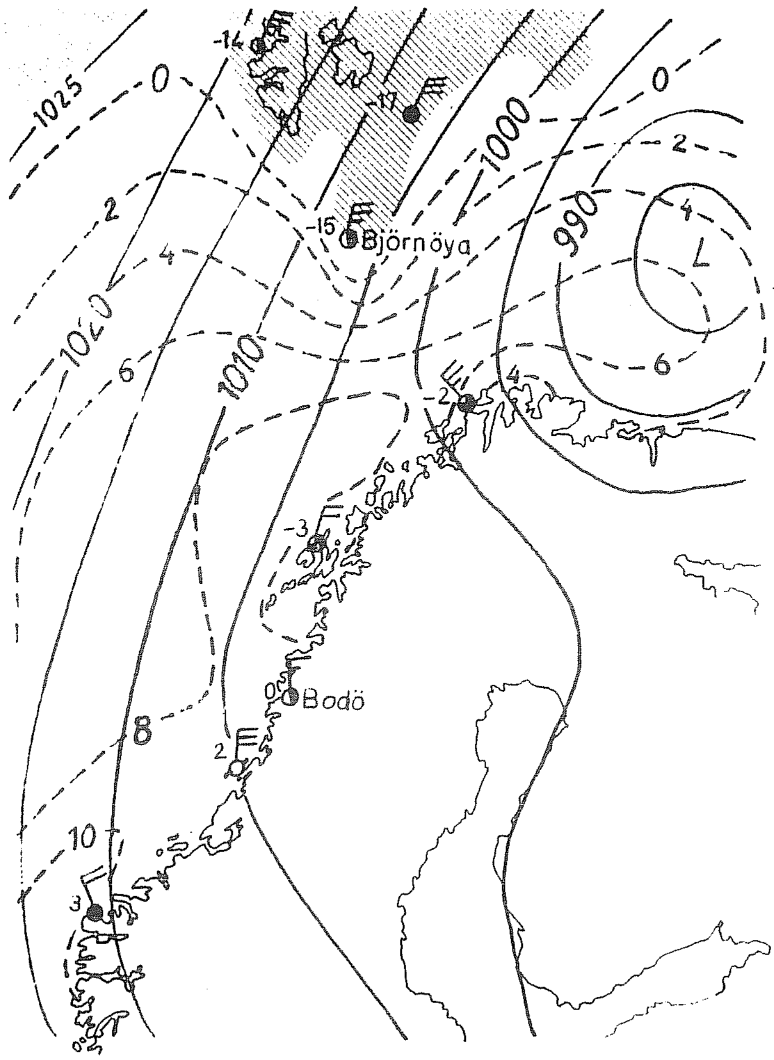

FIGURE 1: Synoptic situation in the Norwegian Sea for 21 November 1975, 12 GMT.
Solid lines : Surface isobars.
dashed lines: Sea surface temperature.

(From Økland (1976))

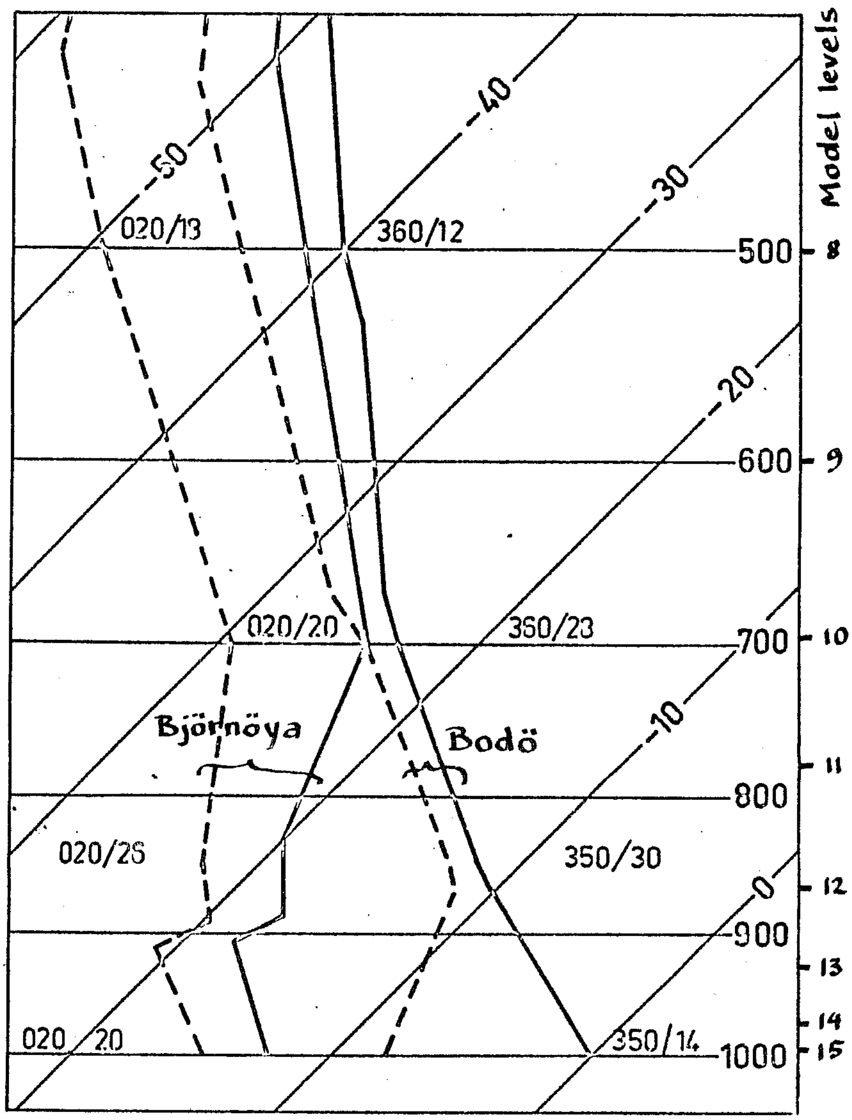

FIGURE 2: Temperature and dew-point soundings for Björnöya and Bodö.

Wind speed and direction are shown at standard levels.

(From Økland (1976))

The assumption that large scale effects do not play any role in this case is most probably not entirely valid, especially above 700 mb where the air is absolutely stable along the trajectory and cannot be reached by convection. It is difficult to see how the increase of dew-point temperature up to 500 mb and even higher can be due to surface effects only. Of course at these low temperatures around $-40^{\circ}C$ the humidity measurements are likely to contain large errors.

At any rate we will concentrate our study to the region below 700 mb and assume that the transformation of that air mass is entirely due to surface fluxes and convection. If we compare the two soundings we can compute that the increase of sensible and latent heat is 25×10^{6} Jm^{-2} and 5×10^{6} $J.m^{-2}$ respectively. This does not necessarily mean that the fluxes of sensible and latent heat have to be in this ratio since some of the latent heat is released and transformed into sensible heat in the cloud layer. A net increase of energy of 3×10^{7} J. m^{-2} in 18 hours corresponds to an average surface flux of about 460 Wm^{-2}.

3. The models

Using the assumption of stationarity and independence from large scale effects, as described above, we can use a simple one-column model to simulate this case of air mass transformation. In this study we use a 15 level model, in σ coordinates. The levels are distributed in the vertical according to the formula :

$$\sigma_i = \sin^2 \frac{\pi i}{30} . \qquad (1)$$

The lowest level is at about 30 m and there are 6 levels below 650 mb. The height of the levels is indicated on Fig. 2.

The model equations are the following :

$$\frac{\partial \mathbf{V}}{\partial t} = f\,\mathbf{k} \times (\mathbf{V}_g - \mathbf{V}) + \mathbf{F}_M \qquad (2)$$

$$\frac{\partial T}{\partial t} = \frac{L}{c_p} C + F_H \qquad (3)$$

$$\frac{\partial q}{\partial t} = - C + F_q \qquad (4)$$

The functions F are the eddy flux divergences or, in the schemes using instantaneous adjustments, the quantities necessary to make the adjustment. C is the condensation rate . The functions F and C are the ones to be parameterised. They are of three kinds : the surface fluxes, the boundary layer processes and the cloud processes. For each process there are several possibilities. We will now describe briefly the methods we have used.

3.1 Surface fluxes

a) GFDL scheme

The GFDL model uses the most simple form of drag formula, with a constant drag coefficient. The fluxes are written as follows :

$$\boldsymbol{\tau}_o = \rho_h \, C_D \, |\mathbf{V}_h| \, \mathbf{V}_h \tag{5}$$

$$f_{H_o} = - \rho_h \, c_p \, C_D \, |\mathbf{V}_h| \, (\theta_o - \theta_h) \tag{6}$$

$$f_{q_o} = - \rho_h \, C_D \, |\mathbf{V}_h| \, (q_s(T_o) - q_h) \tag{7}$$

The index h means the value at the lowest model level and o the value at the surface. In this case the value of the drag coefficient is

$C_D = 1.3 \times 10^{-3}$.

b) ECMWF scheme

In the method proposed for the ECMWF model the surface fluxes depend on the static stability of the atmosphere near the ground. It is based on the Monin-Obukhov similarity theory. According to this theory the profiles of wind and temperature near the ground depend only on the height, the surface fluxes of momentum and heat, and the expansion coefficient g/θ . Only one non-dimensional number can be constructed from these quantities :

$$\zeta = \frac{g \, z \, \theta_*}{\theta \, u_*^2} \tag{8}$$

The scaling velocity u_* and temperature θ_* are defined in relation to the surface fluxes :

$$u_* = \sqrt{|\boldsymbol{\tau}_o| / \rho_o} \tag{9}$$

$$\theta_* = - f_{H_o} / (\rho_o \, u_*) \tag{10}$$

We can now say that the profiles of wind and temperature (when properly non-dimensionalised) are functions of ζ only:

$$\frac{z}{u_*}\frac{\partial u}{\partial z} = \phi_m(k\zeta) \quad (11a)$$

$$\frac{z}{\theta_*}\frac{\partial \theta}{\partial z} = \phi_H(k\zeta) \quad (11b)$$

where k is von Karman's constant.

Using the formulae for ϕ_m and ϕ_H suggested by Businger, et al (1971) from their observations, we can integrate these equations between the roughness length z_o and the height h of the lowest model level. Formally, we can then eliminate ζ and obtain relations between the surface fluxes and the values of the wind and temperature at the ground and at h :

$$u_*^2 = u_h^2 \cdot f\left(\frac{h}{z_o}, Ri_B\right) \quad (12a)$$

$$u_*\theta_* = u_h \cdot (\theta_h - \theta_o) \cdot g\left(\frac{h}{z_o}, Ri_B\right) \quad (12b)$$

where Ri_B is a bulk Richardson number :

$$Ri_B = \frac{g}{\theta}\frac{h(\theta_h - \theta_o)}{u_h^2}. \quad (13)$$

This elimination of ζ , however, cannot be done analytically in the unstable case ($\zeta < 0$), but we can construct the functions f and g point by point, using an iteration method.

Since iteration methods are slow, we prefer to fit analytical formulae to the curves f and g computed once and for all. The formulae we have chosen are the following:

$$f = a_m\left(1 - \frac{b\,Ri_B}{1 + c_m\sqrt{|Ri_B|}}\right) \quad (14a)$$

$$g = a_H\left(1 - \frac{b\,Ri_B}{1 + c_H\sqrt{|Ri_B|}}\right) \quad (14b)$$

(14a) and (14b): for unstable cases ($Ri < 0$)

and

$$f = a_m \left(1 - \frac{b}{2} Ri_B\right)^2 \quad (14c)$$

$$g = a_H \left(1 - \frac{b}{2} Ri_B\right)^2 \quad (14d)$$

for stable cases ($0 < Ri < 2/b$)

$$f = g = 0 \qquad \text{for } Ri > 2/b \quad (14e)$$

b is a constant, a and c depend on h/z_o.

It can be seen that in neutral conditions (Ri=0) f and g reduce to constants and the fluxes, then, have the same form as in the GFDL model. The formula for the flux of water vapour is similar to that for heat.

3.2 Eddy fluxes in the boundary layer

a) GFDL method

In the GFDL model there is a vertical eddy flux of heat only when the temperature distribution is statically unstable. Then the transport of heat is simulated by an adiabatic adjustment, in which the super-adiabatic gradients are replaced by dry adiabatic lapse rates, while keeping the total internal energy of the layers constant.

The fluxes of momentum and water vapour are represented by a diffusion law where the diffusion coefficient is:

$$K = l^2 \, |\Delta V / \Delta z| \quad (15)$$

with the mixing length l maximum at the lowest level and decreasing linearly with height, becoming 0 at 2500 m.

b) Dry diffusion

It can be shown that the effect of the dry adiabatic adjustment is identical to that of a diffusion with the following diffusion coefficient:

$$K = \frac{\Delta z}{\Delta t \, \overline{\left(\frac{2}{\Delta z}\right)^2}} \quad (16)$$

if unstable, and zero in stable conditions. Here t is the time step, ΔZ the level thickness and $\overline{(2/\Delta z)^2}$ represents an average over two adjacent layers.
In the method which we call dry diffusion we apply this formula to all three variables: heat, momentum and water vapour.

c) ECMWF scheme

In the ECMWF scheme we also use a diffusion formula, but we make the diffusion depend on static stability in a manner similar to that of the surface fluxes. Hence we write the following:

$$K = \ell^2 \left[\left| \frac{\Delta \mathbf{V}}{\Delta z} \right| - \frac{b \frac{g}{\theta} \frac{\Delta \theta}{\Delta z}}{\left| \frac{\Delta \mathbf{V}}{\Delta z} \right| + c \sqrt{\frac{g}{\theta} \left| \frac{\Delta \theta}{\Delta z} \right|}} \right] \qquad (17a)$$

(unstable case)

$$K = \ell^2 \left(1 - \frac{b}{2} \frac{g \frac{\Delta \theta}{\Delta z}}{\theta \left| \frac{\Delta \mathbf{V}}{\Delta z} \right|^2} \right)^2 \left| \frac{\Delta \mathbf{V}}{\Delta z} \right| \qquad (17b)$$

(stable case)

Again, b is a constant and c depends on height and level thickness. We take the mixing length as suggested by Blackadar (1962) :

$$\ell = kz / \left(1 + \frac{kz}{\lambda}\right) \qquad (18)$$

where k is von Karman's constant and λ , the asymptotic mixing length, is an adjustable parameter. In this case we have used λ = 500 m.

3.3 Cloud processes

a) GFDL (Manabe)

The treatment of convection in the GFDL model is a moist adiabatic adjustment. If, at one time step, two layers are saturated and the lapse rate between them is greater than the moist adiabatic lapse rate, then at the next time step the moist adiabatic lapse rate is reestablished while keeping constant the total energy (internal + latent) of the column. The layers are also kept saturated.

b) Kuo scheme

In the Kuo method, the moisture influx below the clouds is first computed. In this case of a one-column experiment there is no horizontal advection and the only input of moisture is the surface flux:

$$I = f_{qo}. \qquad (19)$$

Then the surplus energy in the cloud is computed:

$$E = \int_{z_{Bottom}}^{z_{Top}} [\, c_p (T_c - T) + L (q_c - q)]\, dz . \qquad (20)$$

The temperature in the cloud T_c is computed following the moist adiabat from the lifting condensation level z bottom. The mixing ratio in the cloud q_c is the saturation mixing ratio for T_c .

Equating the influx of latent heat and the surplus energy gives the rate of production of cloud α :

$$LT = \alpha E. \qquad (21)$$

Finally the new temperature and humidity are computed :

$$T^{N+1} = T^N + \alpha (T_c - T^N) \qquad (22a)$$

$$q^{N+1} = q^N + \alpha (q_c - q^N). \qquad (22b)$$

c) ECMWF scheme

This method, which is still in the early stages of development, simulates the vertical eddy transport of momentum, heat and water vapour in the clouds by a diffusion formula similar to that in the boundary layer. Let us write the equation for temperature and total (vapour + liquid) moisture, neglecting advection and radiation:

$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left(K \frac{T}{\theta} \frac{\partial \theta}{\partial z} \right) + \frac{L}{c_p} Q \qquad (23a)$$

$$\frac{\partial q_t}{\partial t} = \frac{\partial}{\partial z} \left(K \frac{\partial q_t}{\partial z} \right) + \frac{\partial P}{\partial z} \qquad (23b)$$

The first term of the r.h.s. is the diffusion term, Q is the condensation rate and P is the precipitation. The diffusion coefficient K is now slightly different from the one given above (3.2.c) because now the lapse rate corresponding to neutral stability is somewhere between the dry and moist adiabatic lapse rates depending on the cloudiness. Hence in the formulae given in (3.2.c) for the diffusion coefficients we replace $\frac{1}{\theta} \Delta\theta/\Delta z$ by

$$\frac{1}{\theta} \frac{\Delta\theta}{\Delta z} + \alpha \frac{\Delta \left(\frac{L q_s}{c_p T} \right)}{\Delta z} \qquad (24)$$

We must define the cloudiness α. From the work of Pham and Rousseau (1976) we relate it to the relative humidity and the height :

$$\alpha = \left(\frac{q_w/q_s - \sigma}{1 - \sigma} \right)^2 \tag{25}$$

or $$\alpha = 0 \quad \text{if} \quad \frac{q_w}{q_s} < \sigma$$

q_w is the water vapour mixing ratio and σ is the ratio of the pressure at the level to the ground pressure. The reason to use σ in the formula is that it seems to fit the data better, but it could well be considered as an adjustable parameter. To distinguish between water vapour and liquid water we do not have a separate prognostic equation for each, but we assume that any change in total relative humidity is distributed between liquid water and water vapour relative humidity according to cloudiness. This can be written :

$$\delta\left(\frac{q_t}{q_s}\right) = \frac{\delta\left(\frac{q_l}{q_s}\right)}{\alpha} = \frac{\delta\left(\frac{q_v}{q_s}\right)}{1-\alpha} \tag{26}$$

Eliminating α between this equation and the definition of α we obtain a differential equation for q_v/q_s in terms of q_t/q_s This equation can be integrated from σ to q_t/q_s to give:

$$\frac{q_v}{q_s} = \frac{1 + (2\sigma - 1)\exp\left(-2\,\frac{q_t/q_s - \sigma}{1-\sigma}\right)}{1 + \exp\left(-2\,\frac{q_t/q_s - \sigma}{1-\sigma}\right)} \tag{27a}$$

or $$\frac{q_v}{q_s} = \frac{q_t}{q_s} \quad \text{if} \quad \frac{q_t}{q_s} \leq \sigma \tag{27b}$$

The liquid water q_l is the difference between q_t and q_v.

Finally we need relations for the condensation and the precipitation. First we make the assumption that the rain is stationary, that is :

$$Q = -\frac{\partial P}{\partial z} \tag{28}$$

then we compute the precipitation with a cloud model similar to Kessler's (1969):

$$\frac{\partial (P+k_1)^{k_2}}{\partial z} = k_3 \frac{q_l}{\alpha} \tag{29a}$$ in clouds

$$\frac{\partial (P)^{k_4}}{\partial z} = k_5 \left(\frac{q_w - \alpha q_s}{1-\alpha} - q_s \right) \tag{29b}$$ in clear air.

This method seems rather complicated compared to Manabe's and Kuo's, but one must remember that the diffusion calculation (or an equivalent one) has to be done anyhow to take care of dry convection, and that this method is an attempt to take into account the transport of momentum in clouds and the evaporation of rain below the cloud, and it gives a value of the cloudiness which can be used by the radiation program. The two other methods cannot do that.

4. The experiments

The initial conditions for the experiments were the sounding at Björnöya. The observed ocean surface temperature was taken as bottom boundary condition (no flux is assumed at the top) and the geostrophic wind was kept constant. We present here the results of six experiments, using various combinations of the physical parameterisations described above. The following table shows the various combinations used.

EXPERIMENT	SURFACE	EDDY FLUXES	CLOUD PROCESSES
1	GFDL	GFDL	GFDL
2	GFDL	GFDL	Kuo
3	GFDL	Dry diffusion	GFDL
4	GFDL	Dry diffusion	Kuo
5	ECMWF	ECMWF	Kuo
6	ECMWF	ECMWF	ECMWF

The results of the experiments are shows on figures 3 to 8.

The surface fluxes in the two methods were roughly the same. This is due to the fact that the degree of instability observed produced, in the ECMWF method, an equivalent drag coefficient approximately the same as the one used in the GFDL model.

If the dry diffusion and the ECMWF method for the eddy fluxes give similar results, the GFDL method does not. The temperature profile obtained is fairly good, but the dew point temperature profile is quite wrong. This is due to the fact that this method does not take into account the strong instability and limits the boundary layer to an arbitrary height.

Finally the Kuo scheme transports moisture higher than the other two methods, which behave in a fairly similar fashion. It is not clear, from this set of experiments, whether Kuo's method is better, or whether the suggested ECMWF method of dealing with the cloud processes is satisfactory.

In conclusion we can say that this kind of experiment cannot necessarily point to which parameterisation method is best, but it can show important deficiencies in some schemes, and should be used to show whether a proposed scheme is flexible enough to respond to extreme situations which may be important for accurate weather forecasts.

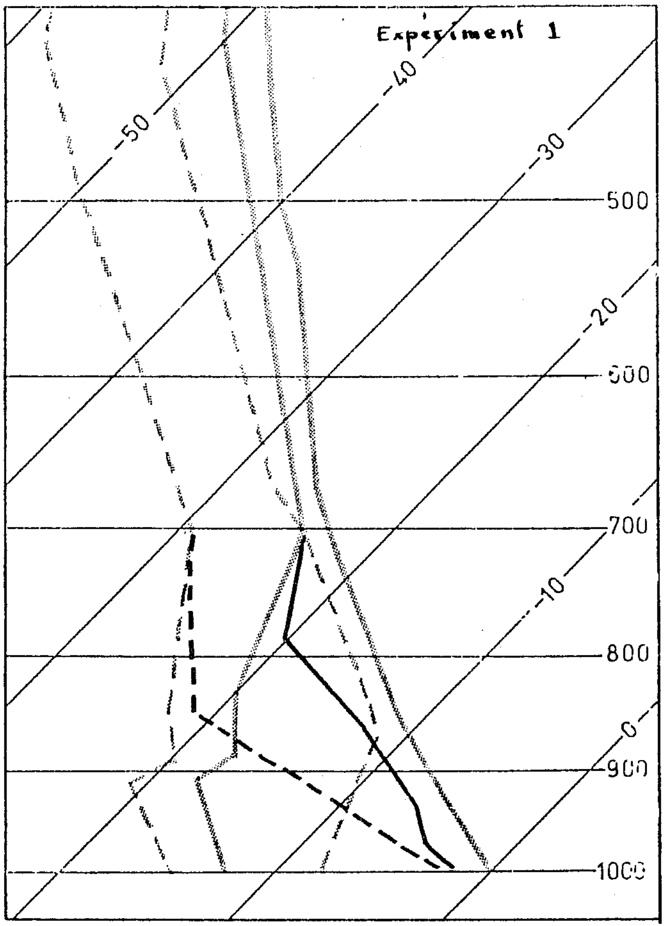

FIGURE 3:

Results with the complete GFDL parameterization scheme.

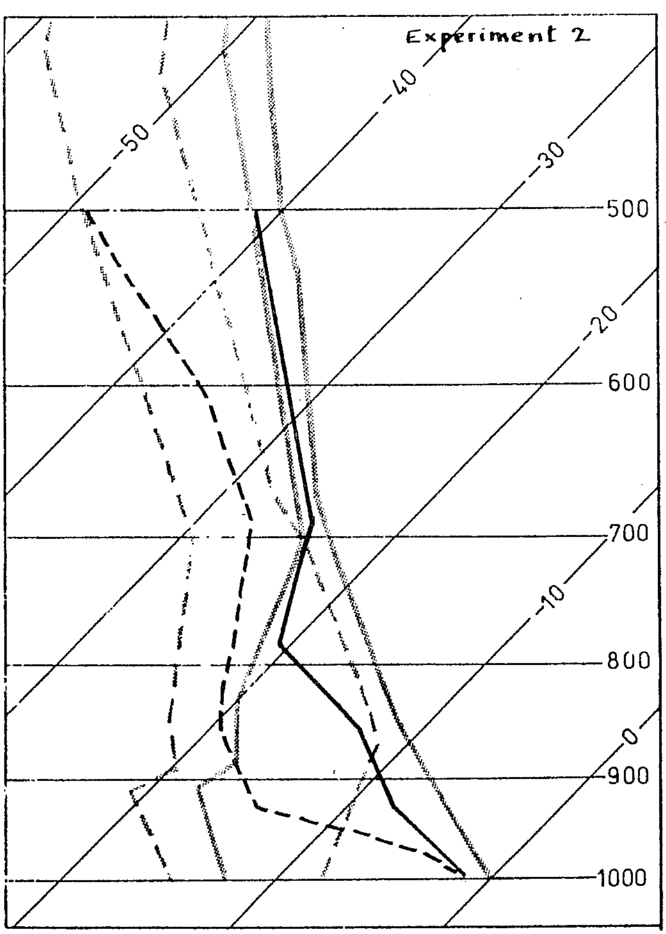

FIGURE 4:

Results with the GFDL boundary layer and the Kuo convection.

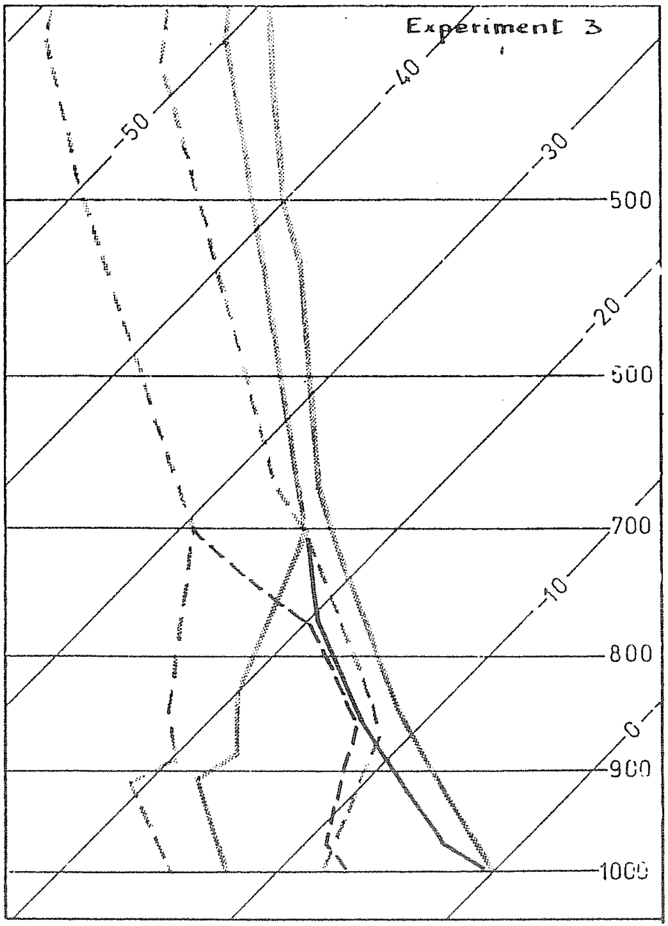

FIGURE 5:

Results with dry diffusion
and moist adiabatic adjustment

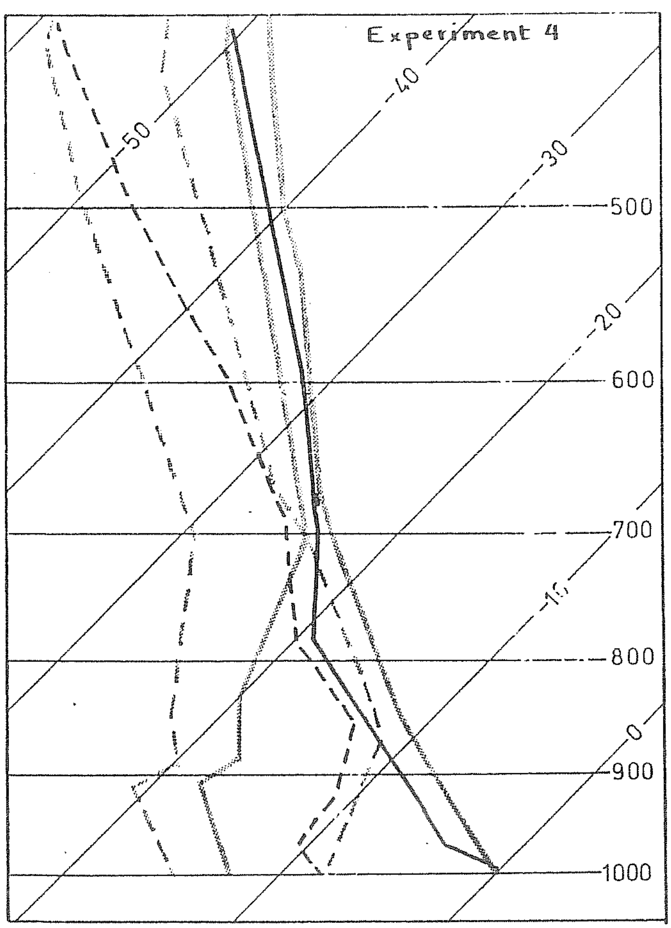

FIGURE 6:

Results with dry diffusion
and Kuo convection.

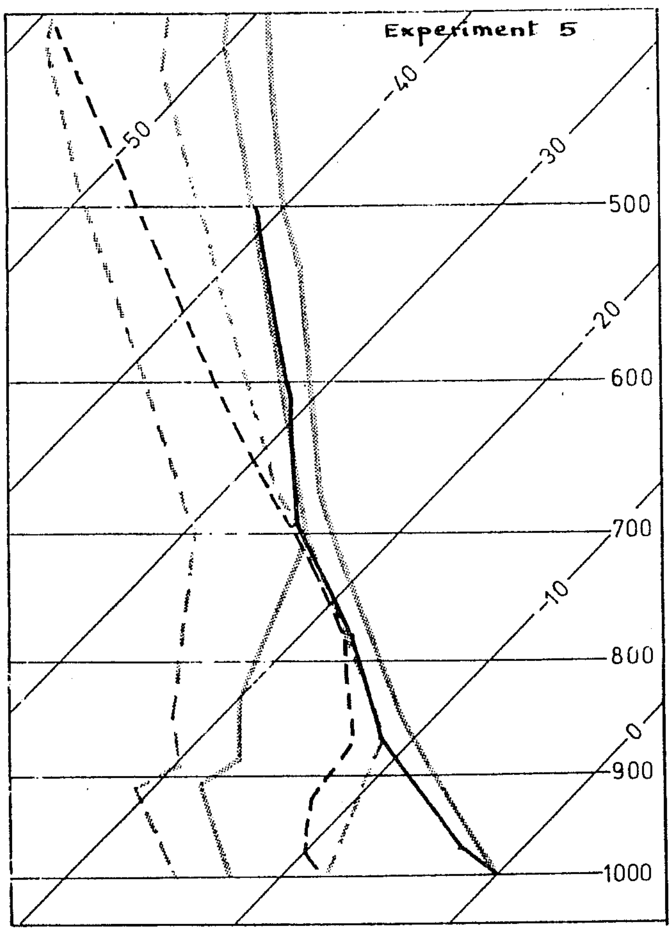

FIGURE 7:

Results with ECMWF diffusion scheme and Kuo convection.

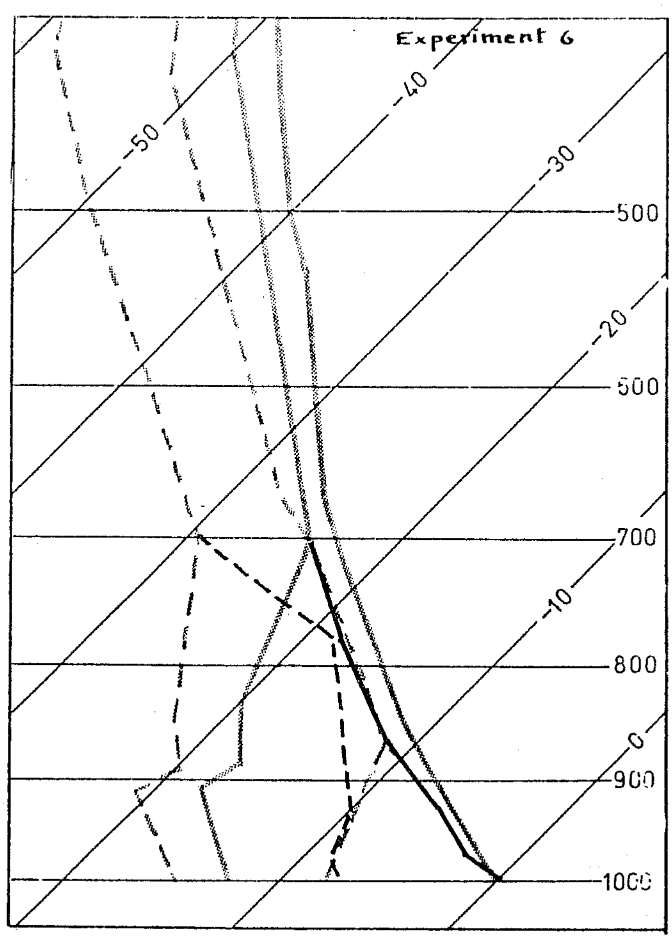

FIGURE 8:

Results with proposed ECMWF method.

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