BOUNDARY LAYER
A N D
FREE ATMOSPHERE.

General Considerations

By

J.-F. Louis

EUROPEAN CENTRE FOR MEDIUM RANGE WEATHER FORECASTS,
BRACKNELL, U.K.

1. Introduction

The interaction between the atmosphere and the various physical (i.e. non adiabatic) processes is a problem which is very scale dependent, both in time and space. If one considers the time-averaged structure of the atmosphere on the global scale, one can, with a good approximation, explain it as a balance of various physical processes, if the motions are taken into account as some kind of diffusion process.

On the other hand, if one considers the synoptic motions of the atmosphere on a short time-scale (order of 1 day), they are very close to the adiabatic motions of an inviscid fluid.

Translating this in terms of modelling it is clear that for climate models the physics is very important and should definitely be included, whereas for short-range forecast it can be left out without too much penalty. At which point should it be included and with what sophistication is a basic problem in medium-range forecast and, in a way, the theme of this whole seminar. In this lecture we consider, in fairly general terms, whether the surface exchanges affect the general flow patterns and how accurately they should be represented.

Again if we consider the global circulation we see that it is basically driven by the excess solar heating in equatorial regions and deficit in the polar region. This is exemplified by Fig. 1 taken from Lorenz (1967), p. 54.

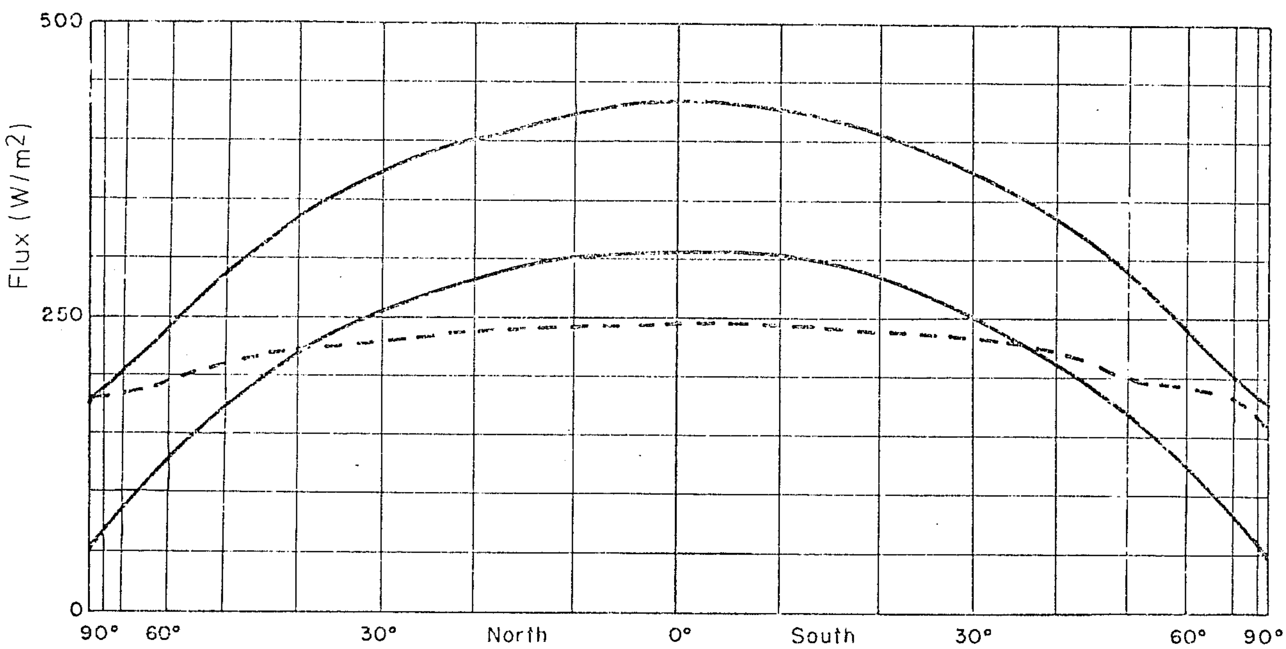

Figure 1 — Average solar energy reaching the extremity of the atmosphere (upper solid curve), average solar energy absorbed by the atmosphere-ocean-Earth system (lower solid curve), and average infra-red radiation leaving the atmosphere-ocean-Earth system (dashed curve), as given by Sellers (1968). Values are in watts m^{-2} (scale on left). (1 watt m^{-2} = 1.435×10^{-3} cal cm^{-2} min^{-1} = 0.754 kilolangleys $year^{-1}$.)
From Lorenz (1967)

Most of this solar energy is absorbed by the ground, then transmitted to the atmosphere by the surface fluxes, especially the flux of latent heat. This creates a difference in potential energy of the atmosphere between the equator and the pole, and the circulation responds to this imbalance, trying to even it out. It is evident, then, that in order to simulate the climate correctly, it is necessary to have a good representation of the surface fluxes.

If we go back to the short-range forecast problem, however, especially in middle latitudes, it is usually quite sufficient to consider the surface as a boundary that the air cannot cross, and nothing more. Many short-range forecast models do just that.

A 10 day forecast is somewhere between these two extremes. After a few days the effect of the surface fluxes begins to be felt and modulates the motions of the free atmosphere. Here we will attach some numbers to this qualitative statement, and try to get a feel for the magnitude of the influence of the surface fluxes on the atmosphere and the time scale over which this influence is felt.

2. Momentum

Let us start with the flux of momentum.

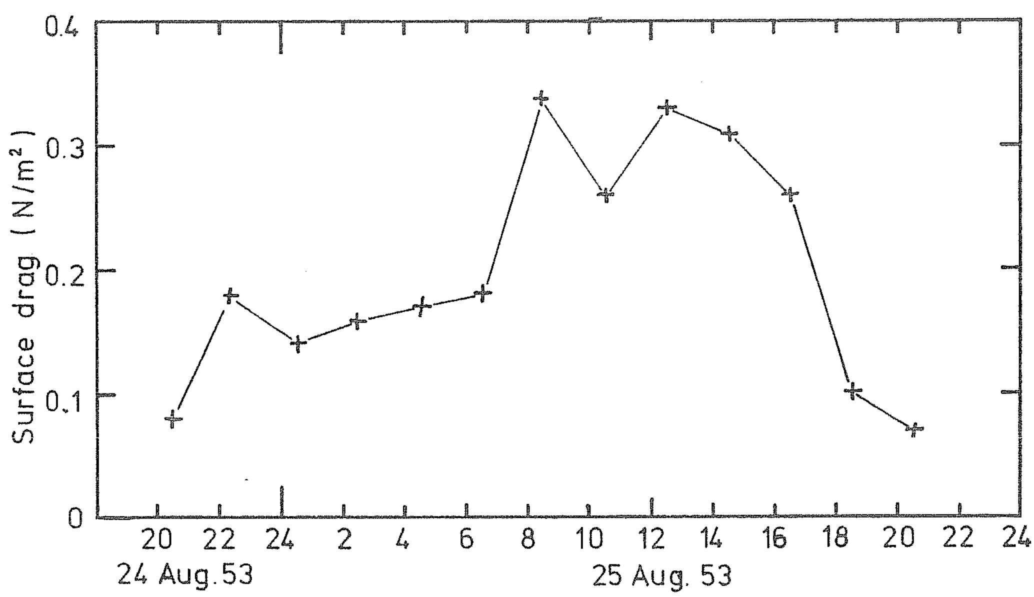

Figure 2. Measured surface drag during the 1953 O'Neill experiment. Data taken from Lettau and Davidson(1957)

Fig. 2 shows direct measurements (using drag plates) of the shear stress at the ground taken during the O'Neill, Nebraska experiment in 1953 (Lettau and Davidson, 1957). The wind was varying around 7 to 10 m/sec., and these values of about $.2 N/m^2$ for the surface stress can be considered typical for a flat terrain with fairly short vegetation. Slightly smaller values (.1 to $.2\ N/m^2$) were estimated by Kondo (1975) for the East China Sea during the AMTEX '74 experiment. In rough terrain, the stress is expected to be somewhat larger.

Let us now compute the time it would take for the whole depth of the atmosphere to be affected by a stress τ of $.2N/m^2$. Let us take an average wind speed u of 15 m/sec. and a depth of Δz 10 km. We get a time scale:

$$\Delta t = \frac{\rho u \Delta z}{\tau} = \frac{15.10^4}{.2} \simeq 7.5 \times 10^5 \text{ sec.}$$

This is roughly about 10 days. It means that, in the absence of pressure forces and internal dissipation, the atmosphere would slow down to about half its original momentum in about 10 days. It also means that if we want the 10 day forecast value of the momentum to be accurate, say, to within 10%, then the surface drag should also be accurate to within 10%, whereas for a 1 or 2 day forecast, for the same accuracy on the momentum it is only necessary to know the order of magnitude of the surface flux (assuming that everything else is correct).

One should note that the surface drag affects the free atmosphere only indirectly. First of all, if the surface drag increases, it does not slow down the flow but increases the cross-geostrophic angle of the wind, increasing the transfer from high pressure to low pressure areas. It is this decrease in the pressure gradient which slows down the flow. Furthermore this whole process happens mostly in the boundary layer, and it is usually only through the slow vertical advection of momentum that the free atmosphere is affected by the surface drag.

In some cases, however, the ground effect may be much more pronounced. Fig. 3 shows a case studied by Lilly and Kennedy (1973), where the Rocky Mountains produce waves of large amplitude in their lee. These waves extract momentum from a region in the upper troposphere and lower stratosphere and transport it downwards to the ground. In this case the wind at 200 mb slowed down by 10 m/sec in 100 km. Notice that these waves have wave lengths shorter than the grid mesh of most forecast models, which would pose a serious parameterisation problem.

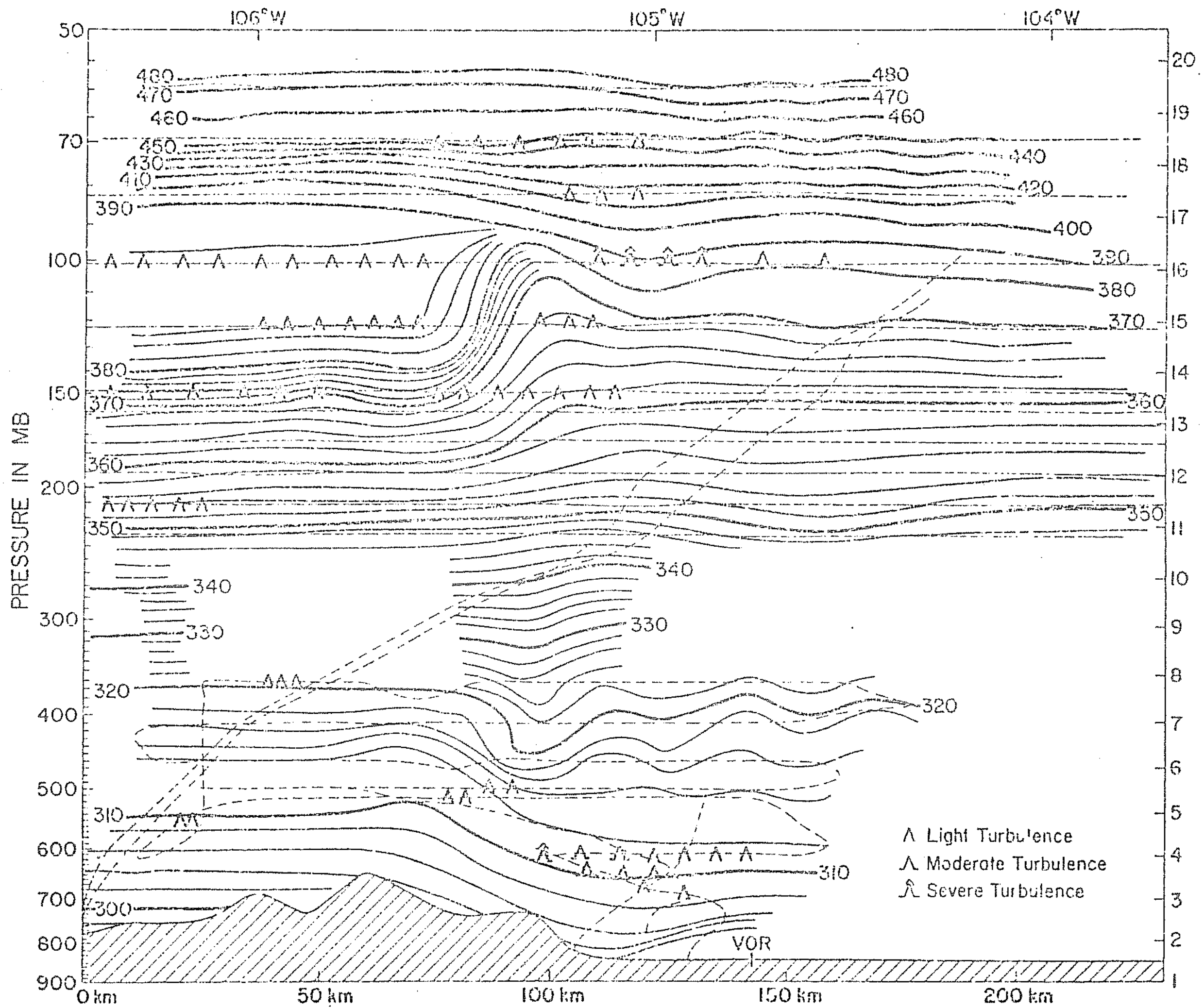

FIG. 3. Potential temperature cross section for 17 February 1970. Solid lines are isentropes (°K), dashed lines aircraft or balloon flight trajectories. The cross section is along a 275°-095° true azimuth line, crossing the Kremmling, Colo., and Denver VOR aircraft navigation stations. From Lilly and Kennedy (1973)

3. Sensible heat

There is no method equivalent to the drag plate to measure directly the flux of sensible heat through the ground surface. One must rely on measurements of the correlation between vertical wind and temperature near the ground, using fast-response instruments. Fig. 4 shows measurements taken during the Kansas experiment (Wyngaard, 1973). It is a composite picture including several days of observations. It shows a large variation of the sensible heat flux during a 24 hour period : roughly sinusoidal during the day and almost constant, negative (i.e. from the air into the ground) at night. The maximum value during the day is, in this case, about 350 W/m^2. The average is somewhat under 100 W/m^2. These are rather extreme values, characteristic of hot summer days. The figures would be quite a bit smaller in winter or in cloudy conditions.

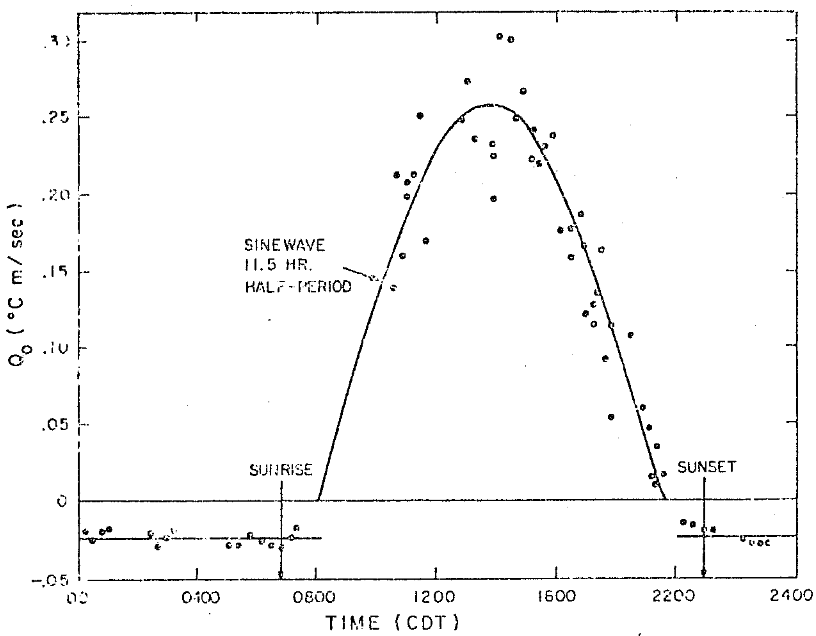

FIG. 4 The time variation of surface heat flux (determined from measurements at 6 m) during the Kansas experiments. The scatter is due to the day-to-day variations during the three weeks of runs. From Wyngaard (1973).

There are few measurements of the sensible heat flux over the ocean. An estimate by Bunker (1976) of the annual average sensible heat flux for the North Atlantic is shown on Fig. 5. A fairly low value around 10 W/m^2 prevails over most of the area, but a few places have much larger values, especially in the Gulf Stream regions, with maxima greater than 50 W/m^2.

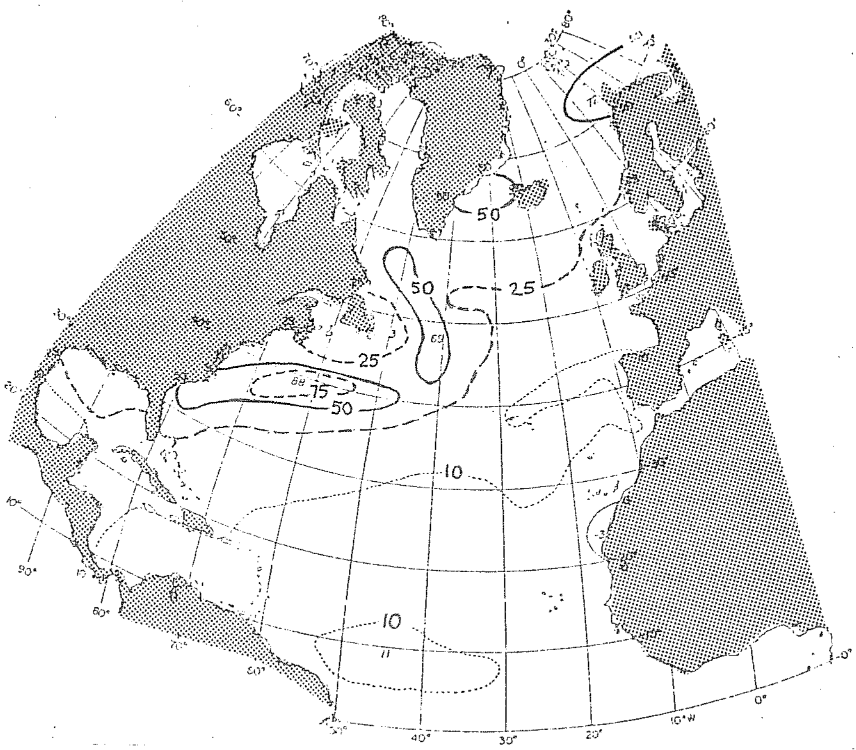

Figure 5. Annual average sensible heat flux ($Watt/m^2$) for the North Atlantic. From Bunker (1976)

Let us now translate these figures in terms of heating rate. If absorbed evenly through the depth of the atmosphere, a surface flux of 10 to 50 W/m^2 would produce a heating rate of only about .1 to .5° K/day. These figures are not very meaningful, however, because most of the sensible heat coming from the ground is absorbed in the boundary layer, in the first 1 or 2 km of the atmosphere. In our first example of the Kansas summer day, the heating rate in the boundary layer during the day is up to about 10° K/day, a very large figure. But almost all of this heating is cancelled at night, partly by the smaller downward sensible heat flux, and partly by radiative cooling. The net contribution of the sensible heat flux to the energy balance of the free atmosphere is, in this case, very small indeed.

Over the oceans, where the flux is the same for long periods, the effect may be more pronounced, but it is overshadowed by the flux of latent heat, as we shall see presently.

4. Water vapour

Measurements of the water vapour flux are difficult. Over land the evaporation depends very much on the type of ground and the vegetation. The evaporation of a free water surface can be measured directly, using open pans, but it is difficult to take into account the effect of spray and breaking waves in the open sea. However, overall, the total evaporation must be balanced by the precipitation which is rather easier to measure, at least over land.

Measurements over well irrigated turf in California are shown in Fig. 6 (taken from Gadd and Keers, 1970). They show that, in this case, the surface flux of latent heat (LF_W) is up to 500 W/m^2 in the afternoon and almost balances the net radiative flux (R_N) reaching the ground. It is several times larger than the flux of sensible heat (F_H) or the flux into the ground (F_G). It is only in desert areas that the sensible heat flux consistently exceeds the latent heat flux.

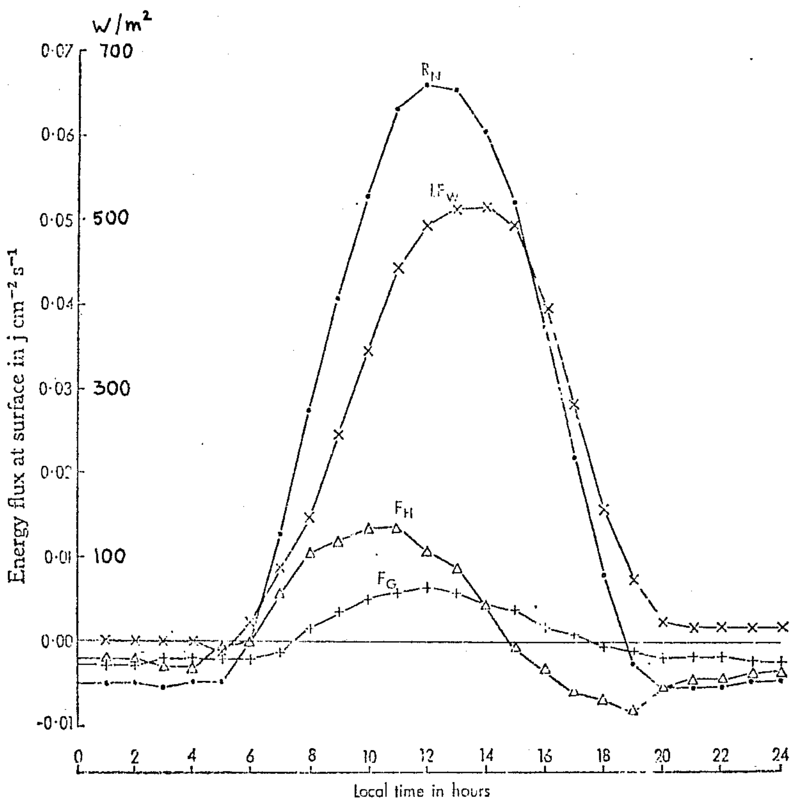

Figure 6. The diurnal variation of the energy fluxes at the surface as measured by Brooks and Goddard (1966) on 3 June 1965 over well irrigated turf at Davis, California. From Gadd and Keers (1970).

If we compare Fig. 7 to Fig. 5 we see that over the oceans the average flux of latent heat is also several times greater than the flux of sensible heat (an evaporation of 1cm/year corresponds to a latent heat flux of approximately .8W/m^2).

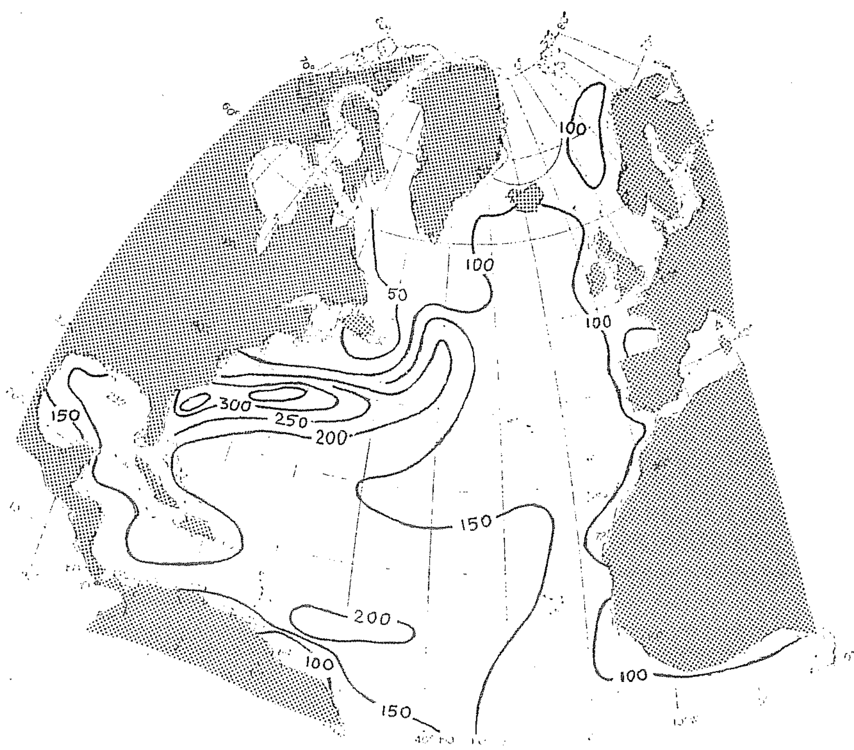

Figure 7. Annual average evaporation (cm/year) for the North Atlantic. From Bunker (1976).

Evaporation has two effects on the atmosphere : directly, it affects the water vapour distribution, and indirectly it modifies the temperature structure through the release of latent heat in clouds. How fast are these effects ? Let us take a mean value of 100 cm/year (roughly the global average precipitation); this corresponds to about 3 x 10^{-5} kg/m^2 sec. Since the average quantity of water present in an atmospheric column is around 25 kg/m^2, the time it takes to renew most of the water in the atmosphere is only about 8.10^5 sec., or about 10 days.

The thermal effect of water vapour, the release of latent heat, is also very important, not only because of its magnitude (several degrees per day), but also because the release takes place in the middle of the atmosphere, as opposed to the flux of sensible heat which affects mostly the boundary layer.

This can be seen, for example, in the analysis by Johnson (1976) of an easterly wave in the Pacific. Fig. 8 shows schematically his estimate of the budget of water vapour in the boundary layer. It can be seen that, despite a surface flux of .35 cm/day (equivalent to a latent heat flux of 110 W/m^2) and an advection of .36 cm/day the net moisture content of the boundary layer is almost in steady state. In fact, in this case, a small loss of 0.1 cm /day occurs. The pumping of water vapour in the updrafts of the clouds is so strong that it counter-balances not only the advection of moisture and the surface flux, but also the downward fluxes due to the downdrafts in the environment and the base of the clouds, as well as the evaporation of rain in the sub-cloud layer.

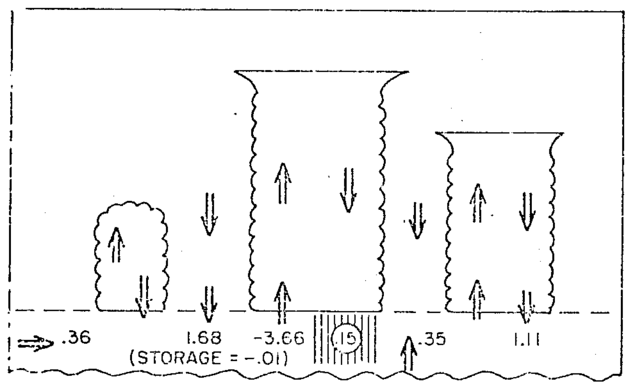

Figure 8. Example of water vapour budget for the subcloud-layer. From left to right: convergence, environmental sinking, updraft transport, rain evaporation, surface flux and downward transport. Units: cm/day. From Johnson (1976).

His computation of the fluxes in the cloud layer shows that the moisture converging in the sub-cloud layer is carried by the updrafts all the way to the 200 to 500 mb layer.

5. Conclusion

When we look at the magnitude of the surface fluxes of momentum, sensible heat and water vapour, two things become apparent. First of all we see that, on average, these fluxes have a significant effect on the structure of the atmosphere over a period of the order of 10 days. The sensible heat flux is less important because it is in general smaller than the latent heat flux and because it affects only the boundary layer. It can influence the mean flow only inasmuch as the diurnal cycle can modulate the growth of clouds which release latent heat in the free atmosphere.

The second observation that we can make is that the surface fluxes vary widely in time and space so that, if their mean effect has a time scale of about 10 days, locally or in certain conditions their influence can be felt much more quickly throughout the depth of the troposphere.

References:

Bunker, A.F. (1976) Computations of surface energy flux and annual air-sea interaction cycles of the North Atlantic Ocean. Mon. Wea. Rev., 104, 1122-1140.

Gadd, A.J., J.F. Keers (1970) Surface exchanges of sensible and latent heat in a 10-level model atmosphere. Quart. J. R. Met. Soc. 96, 297-308.

Johnson, R.H. (1976) The role of convective-scale precipitation downdrafts in cumulus and synoptic-scale interaction. J.Atmos. Sci. 33, 1890-1910.

Kondo, J. (1975) Heat balance of the East China sea for AMTEX '74. Scientific Report of the 4th AMTEX Study Conference, Tokyo, Sept. 75, pp. 12-15.

Lettau, H.H., B.Davidson (1957) Exploring the Atmosphere's First Mile. Vol. 2. New York, Pergamon Press, 578 pp.

Lilly, D.K., P.J. Kennedy (1973) Observations of a stationary mountain wave and its associated momentum flux and energy dissipation. J. Atmos. Sci., 30. 1135-1152.

Lorenz, E.N. (1967) The Nature and Theory of the General Circulation of the Atmosphere. World Meteorological Organisation No. 218. TP. 115, 161 pp.

Wyngaard, J.C. (1973) On surface-layer turbulence. in: Workshop on Micrometeorology, Ch. 3., D.A.Haugen, ed., Am. Meteor. Soc. Publication.