

Chapter 3

MORE ON DIABATIC PROCESSES

In general the total diabatic heating rate may be written as the sum of three components,

$$Q_{total} = Q_{rad} + Q_{cond} + Q_{sen}$$

the contributions on the right-hand-side being from radiative, condensational and sensible heating, respectively. Figure 3.1 shows schematically the typical vertical distribution of these contributions.

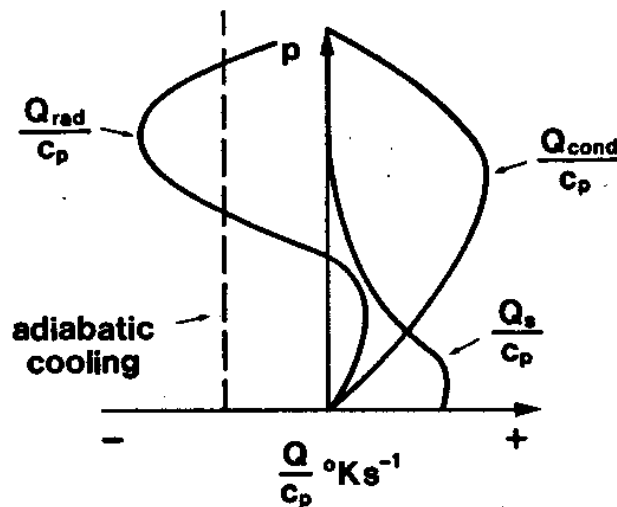


Figure 3.1: Vertical distribution of radiative, condensational and sensible heating.

While diabatic processes drive atmospheric motion, it is important to emphasize that the contributions Q_{rad} , Q_{cond} , Q_{sen} are not *pure* external functions, but are strongly coupled to the flow configuration they produce. This interdependence reflects a redistribution of the only pure external heating function, the solar energy, which impinges on the atmosphere from space (S_0). In order to understand the full relevance of diabatic process and how the drive atmospheric motions, it is necessary to understand the manner in which S_0 is redistributed.

If we assume that the atmosphere is in radiative equilibrium with outer space and note that the earth is a sphere, it is clear that considerably more energy will reach the surface near the equator than near the poles. Consequently, heat energy must be transferred poleward if there is to be an approximate steady state. The question arises, how is the energy S_0 redistributed in the vertical? We consider first the effects of radiation alone.

Figure 3.2 shows a highly simplified radiation model in which we have neglected the absorption of short wave radiation, since it is at the earth's surface that the major effect of S_0 occurs. Pure radiative balance at the surface would require that

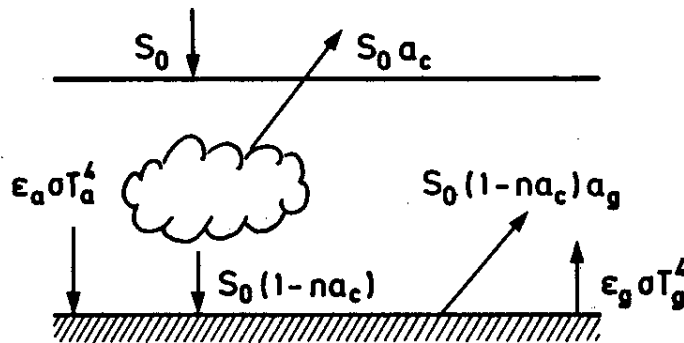


Figure 3.2: A highly simplified radiation model.

$$S_0(1 - na_c)(1 - a_g) = \varepsilon_g \sigma T_g^4 - \varepsilon_a \sigma T_a^4, \quad (3.1)$$

where n denotes the fractional area of cloud, T_s and T_a are the temperatures of the surface and the atmosphere¹, respectively, a_c and a_a are the albedos of cloud and the ground surface, ε_c and ε_s , are the corresponding emissivities, and σ is the Stefan-Boltzmann constant. This equation says that the net short wave flux into the atmosphere is balanced by the net outward long

¹For simplicity it is assumed here that the atmosphere is isothermal.

wave flux. Now $\varepsilon_a \sigma T_a^4$ represents the re-radiation of the atmosphere back to the surface. As $\varepsilon_a \approx 0.7$, we cannot neglect this term. In fact such long wave absorption has a major impact on the distributions shown in Fig. 3.1. However, the solar radiation (the left hand side of Eq. (3.1) may be disposed of in other ways besides being merely radiated upwards (i.e. $\varepsilon_g \sigma T_g^4$) as long wave radiation. Adjacent to the surface there will be diffusion of heat to or from the atmosphere and possibly convective mixing upwards to the atmosphere. Over moist ground or over the ocean, evaporation may occur and there will be an evaporative flux of heat away from the surface (Fig. 3.3). In this case the total energy balance at the surface is

$$S_0(1 - na)(1 - a_g) = \varepsilon_g \sigma T_g^4 - \varepsilon_a \sigma T_a^4 + F_s + F_L, \quad (3.2)$$

where F_s , and F_L are the sensible and latent heat fluxes, respectively. The former quantity, F_s , depends on the magnitude and sign of temperature difference $\Delta T = T_g - T_a$. The behaviour is roughly as sketched in Fig. 3.4. If $\Delta T > 0$, $F_s > 0$ and there is convective heat transfer to the air, If $\Delta T < 0$, $F_s < 0$, but the heat transfer is then diffusive and relatively small (i.e. $|F_{s \text{ conv.}}| \gg |F_{s \text{ diff.}}|$).

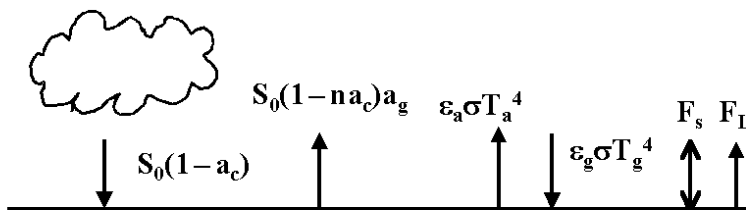


Figure 3.3: Energy balance at the earth's surface.

Generally, $F_L \geq F_s$. Thus F_s , and F_L are usually net fluxes of heat away from the surface. Note that they effectively change radiative diabatic heating into other types of heat transfer, which then contribute to redistributing the heat. Both F_s , and F_L have important effects on the total columnar heating rates, Q_g and Q_L , and both are functions of atmospheric motion and structure. The dependence of F_s on ΔT and on the temperature structure is depicted in Fig. 3.5. Note that the greater ΔT , the greater is the degree of convective instability within the boundary layer, a feature seen also in Fig. 3.5.

The functions F_s and F_L can be parameterized as

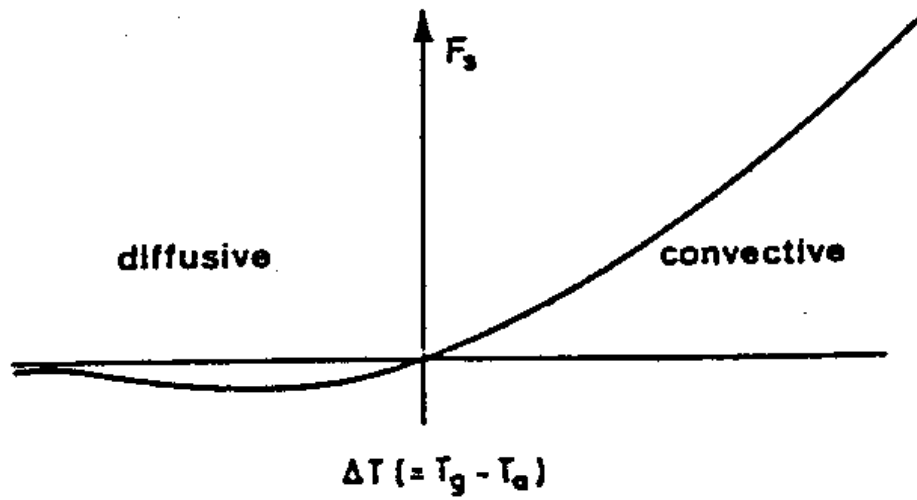


Figure 3.4: $F_s = F_s(\Delta T)$.

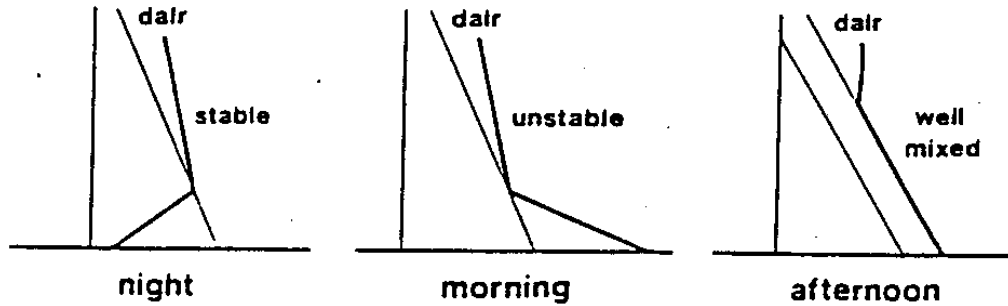


Figure 3.5: Atmospheric conditions influencing F_s .

$$F_s = \rho C_D |V| (T_g - T_a), \quad (3.3)$$

$$F_L = \rho C_g |V| (q_s - q_a), \quad (3.4)$$

where V is the wind speed near the surface, T_a , and q_a are the temperature

and specific humidity of the air near the surface, T_g , and q_g , are the sea surface temperature and saturated specific humidity at the sea surface temperature, respectively, and C_D , C_E are empirical coefficients that depend on the surface characteristics (and over the sea on wind speed). The coefficient C_D is called the *drag coefficient* and C_E is called the *heat transfer coefficient*. Generally, the fluxes F_s and F_L depend on

- (a) the degree of surface roughness,
- (b) the wind speed,
- (c) and in the case of moisture there is a dependency on the degree of saturation in the vertical.

The *sensible heating*, Q_{sen} , tends to be confined to the lowest 1-2 km, except over dry continental surfaces where it may be as high as 4 km. Further it represents an immediate acquisition of heat by the column.

The *latent heating*, Q_{cond} , is not immediate, but requires saturation in order to accomplish the heat release. This is a process that is highly dependent on the dynamics. There are three main ways of producing condensation: two are associated with advection, the other (non-dynamic) by radiative cooling (Fig. 3.6).

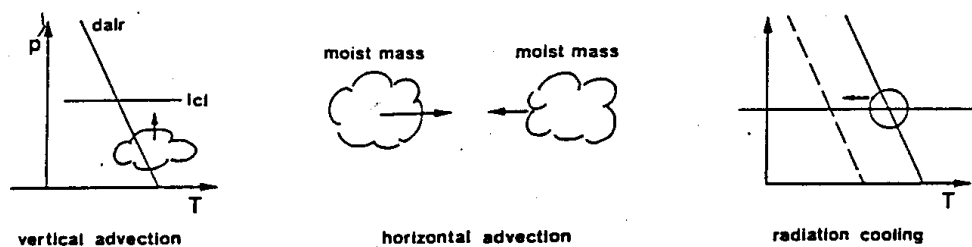


Figure 3.6: Schematic illustration of process leading to condensation.

Most of the latent heat release in the tropical atmosphere is associated with moist convection, vertical advective process. We will restrict our attention to this form of Q_L . The actual region of moist ascent is rather small and rather vigorous. Most of the tropics is involved in the flux of latent heat (F_L), but only a small part is involved in its release. The situation is depicted in the moisture cycle, shown schematically in Fig. 3.7.

In summary, there are two major forms of Q . These are Q_{rad} , principally the longwave component thereof, and Q_L , which is an indirect manifestation

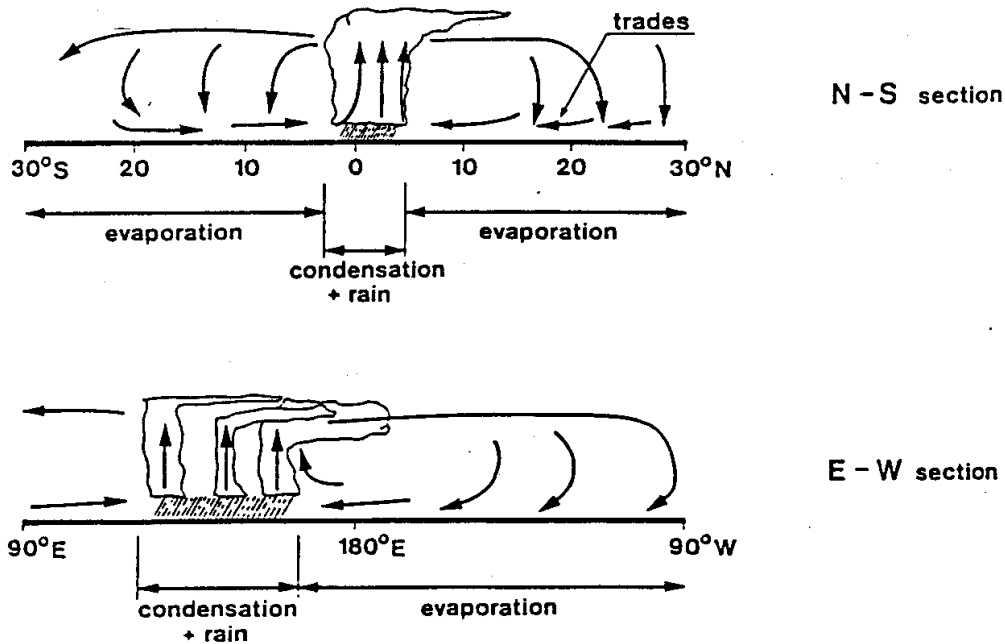


Figure 3.7: The moisture cycle in the tropics.

of S_0 via F_L . Figures 3.8 and 3.9 indicate something of their disposition in the east-west and north-south cross-sections. Of particular importance in Fig. 3.8 is that the *longitudinal gradient* of net flux is of the same magnitude as the *latitudinal gradient*. Most of the variation in the latitudinal profile may be accounted for in the latitudinal gradient of the solar input. However, as the solar input is constant along 25°N , the longitudinal variation in net flux must be due to other effects such as ground albedo and cloud cover; the latter being closely associated with the dynamical system. Note that the desert regions (10°E to 50°E) appear as net radiative sinks with the outgoing longwave radiation (OLR) greater than the net incoming solar flux. On the other hand, the convective monsoon regions (80°E to 18°E) act as net radiative heat sources. In the case shown the radiative heating distribution is probably indicative of the total heating field. The condensational heating will be a maximum in the monsoon regions as Q_{cond} will be strongly tied to the precipitation patterns. In the desert regions the sensible heating will act in a sense opposite to that of radiative cooling, but this component will be smaller and restricted to at most the lowest few kilometres of the atmosphere. Thus atmospheric columns above the deserts should be continually cooling and the columns in the monsoon regions continually heating. A dynamic response is

necessary to rectify the imbalance. Recall that from the scaling of Chapter 2, [see Eq. (2.29)], $N^2 w \approx Q/(c_p T)$, which states that the diabatic heating ($Q_{rad} + Q_{cond}$) is nearly exactly balanced by w . As a consequence, dynamical processes are involved and, consistent with mass continuity, a circulation develops. Note, however, that Q_{cond} will depend on w so that feedback loops are extremely important. To determine the form of the dynamic response, Webster and Stephens (1979) used data from Newell *et al.* (1972) to calculate the heat convergence into the longitudinal section between the arid regions of Saudi Arabia (I), the Arabian Sea (II) and the Bay of Bengal (III). The resultant fluxes, together with the estimates of the vertical profiles of the components of the total heating are shown in Fig. 3.9.

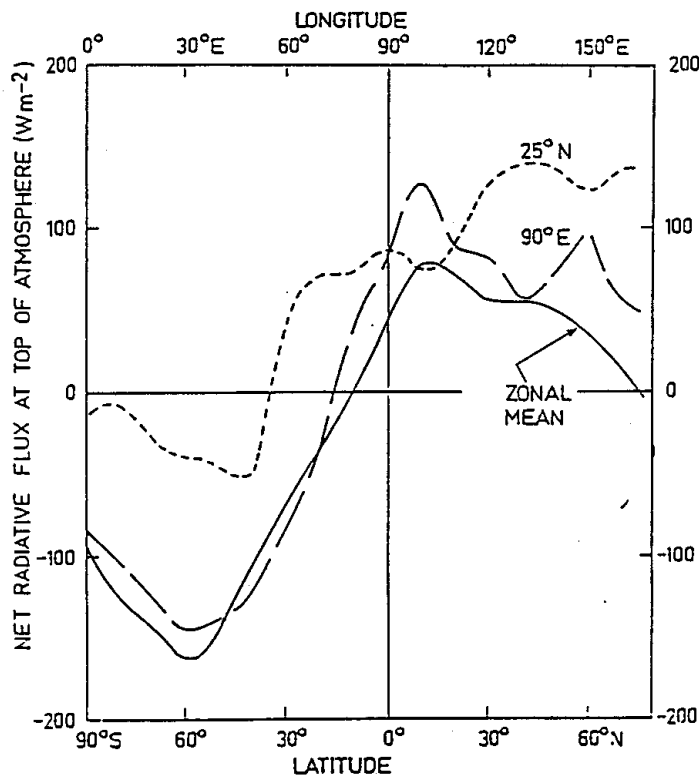


Figure 3.8: The distribution of the net radiative flux at the top of the atmosphere inferred from NIMBUS 3. Plots of the zonally averaged net flux (solid line), the flux along 90° E (largest dashed line) and along 20° N (dashed curve) for July 1969 are shown. (From Webster & Stephens, 1979).

The dynamic response to the heating imbalance is such as to converge

heat into the upper troposphere of the desert regions and out of the convective regions. The net radiative cooling is compensated by adiabatic warming over the deserts and an adiabatic cooling over the Bay of Bengal. In other words, the dynamic response to the longitudinal imbalance of total heating is the generation of a rather vigorous thermally-forced circulation.

More details about the role and representation of moist processes and, in particular, moist convection are given in a later chapter.

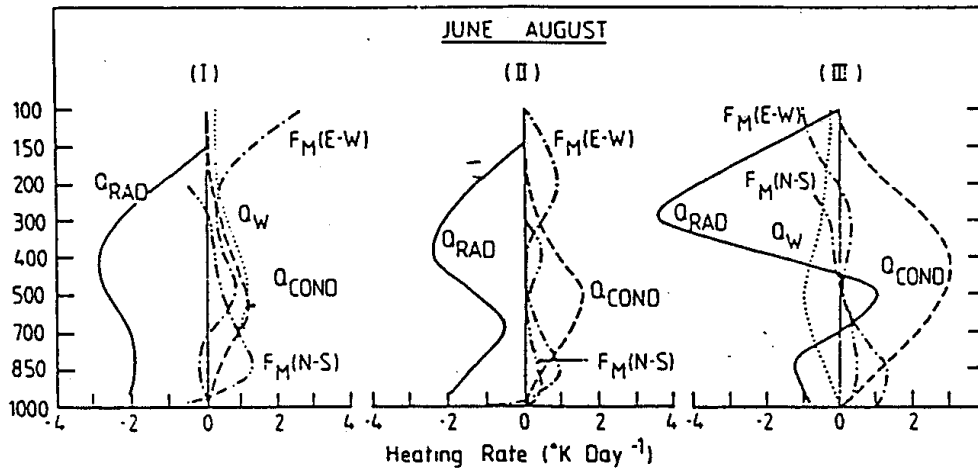


Figure 3.9: The vertical distribution of the heating components in the atmospheric column above Saudi Arabia (I), the Arabian Sea (U) and the Bay of Bengal (III). The quantities Q_{rad} , Q_{cond} , $F_M(E - W)$ and $F_M(N - S)$ refer to heating due to radiation, condensation and heat flux convergence due to mean zonal ($E - W$) and meridional ($N - S$) motions.