Figure 3.1. Zonal-mean temperature lapse rate $-\partial_z T$ (K km$^{-1}$) for the DJF and JJA seasons according to reanalysis data for the years 1980–2001 provided by the European Centre for Medium-Range Weather Forecasts (ERA-40 data; see Uppala et al. 2005). Negative contours are dashed. The thick line marks the tropopause, determined as a 2 K km$^{-1}$ isoline of the lapse rate. The vertical coordinate is pressure normalized by surface pressure, $\sigma = p/p_s$. 

What distinguishes the troposphere and stratosphere kinematically is that the bulk of the entropy the atmosphere receives by the heating at the surface is redistributed within the troposphere, whereas only a small fraction of it reaches the stratosphere. In fluid-dynamical parlance, the troposphere is the caloric boundary layer of the atmosphere; the tropopause is the top of this boundary layer. The question of what determines the thermal stratification is the question of what determines the dynamical equilibrium between radiative processes and dynamical entropy transport. If one accepts as an observational fact that the redistribution of the entropy received at the surface is largely confined to a well-defined boundary layer, the troposphere, the height of the tropopause can be determined as in classical boundary-layer theories: as the minimum height up to which the entropy redistribution must extend for the flow to satisfy large-scale constraints such as energy and momentum balance.

Section 3.2 discusses the general form of large-scale constraints on the thermal stratification and tropopause height, arising from radiative and dynamical considerations. Sections 3.3–3.5 discuss dynamical constraints that respectively take slantwise moist convection, moist convection coupled to baroclinic eddies, and baroclinic eddies as central for determining the thermal stratification and tropopause height. Section 3.6 presents simulations with an idealized general circulation model (GCM) that show that an atmosphere can have different dynamical regimes distinguishable according to whether convection or baroclinic eddies dominate the entropy redistribution between surface and tropopause. And section 3.7 concludes this chapter with a discussion of the results presented and of open questions.
Figure 3.2. Temperature in radiative equilibrium (solid line) and in dynamical equilibrium with fixed tropospheric lapse rate $\Gamma = 6.5 \text{ K km}^{-1}$ (dashed line). The arrow marks the ground temperature in radiative equilibrium, which is greater than the surface air temperature ($T_g \approx 297 \text{ K}, T_s \approx 285 \text{ K}$). The ground temperature in dynamical equilibrium is taken to be equal to the surface air temperature ($T_g = T_s = 280 \text{ K}$). (Calculations courtesy Paul O’Gorman.)
Figure 3.3. Tropopause height (km) determined according to the radiative constraint as a function of tropospheric lapse rate and surface temperature, with fixed relative humidity. Parameters and concentrations of absorbers other than water vapor as in Fig. 3.2. (Calculations courtesy Paul O’Gorman.)
First radiative-convective equilibrium calculations

Fig. 4. The dashed, dotted, and solid lines show the thermal equilibrium with a critical lapse rate of 6.5 deg km$^{-1}$, a dry-adiabatic critical lapse rate (10 deg km$^{-1}$), and pure radiative equilibrium.
Fig. 6c. Thermal equilibrium of various atmospheres which have a critical lapse rate of 6.5 deg km⁻¹. Vertical distributions of gaseous absorbers at 35N, April, were used. $S_e = 2$ ly min⁻¹, $\cos f = 0.5$, $r = 0.5$, no clouds.
Greenhouse effect (again)

\[ \Delta H = \Delta T / \Gamma \]

\[ \Gamma = - \frac{\partial T}{\partial z} \]