Radiative and Radiative-Convective Equilibrium

Introduction

The energy balance of the Earth's system is quite intricate with numerous factors contributing to energy import, transfer, and exchange. We begin our discussion of this energy balance in this section by introducing the concepts of radiative equilibrium and radiative-convective equilibrium. These concepts form the basis for future lectures on the conservation of absolute angular momentum, the meridional overturning Hadley cell circulation in the tropics, and the energy balance of the Earth as a whole.

Key Concept

- What are the concepts of radiative balance and radiative-convective equilibrium and why do they matter?

Energetics of the Earth System

The sun is the primary source of energy for the Earth system, of which the tropics are a portion. Most of this energy comes in the form of radiation energy, often referred to as *insolation*, short for incoming solar radiation. To first order, insolation varies as a function of latitude and the seasonal cycle. Maximum annual insolation occurs at the equator whereas minimum annual insolation occurs at the poles. Insolation magnitudes are fairly steady (~400 W m^{-2}) near the equator throughout the year but exhibit substantial variability at higher latitudes owing to the changing tilt of the Earth as it revolves around the sun (i.e., the seasonal cycle). Also impacting insolation magnitudes are the attenuation of incoming solar energy by the atmosphere and the diffusion of solar energy due to the angle at which it intersects the surface of the Earth. These concepts are illustrated in the accompanying lecture materials.

Recall from introductory courses in physics and thermodynamics that we can express the Stefan-Boltzmann equation, relating the temperature of a blackbody emitter (emissivity = 1) to the energy emitted from that surface, as the following:

(1) $$F = \sigma T^4$$

where $\sigma = 5.67 \times 10^{-8}$ W m^{-2} K^{-4}.

The mean insolation at the top of the atmosphere, sometimes referred to as solar flux density, is given by $S_0 = 1370$ W m^{-2}. To first order, the total absorbed solar radiation can be expressed by:

(2) $$A = S_0(1 - a_P)\pi r_P^2$$

where a_P is the planetary albedo (~0.3) and r_P is the radius of the Earth. Albedo refers to the reflecting power of a surface, such that subtracting it from unity gives a measure (to first order) of the absorbing power of a surface. If we recall that the surface area of a sphere is given by $4\pi r^2$, taking the Earth to be approximately spherical, we can express (2) in terms of absorption per unit area:

(3) $$A = \frac{S_0}{4}(1 - a_P)$$

This reflects absorption (of insolation only) by the atmosphere and surface. Substituting this for *F* into (1) and solving for *T*, we obtain an estimate of 255 K (-18°C) for Earth's equivalent blackbody temperature, reflecting a balance between incoming shortwave and outgoing longwave radiation.

This leads us to the concept of *radiative equilibrium*, where the atmosphere and the Earth's surface are said to be in an equilibrium state in the absence of non-radiative energy processes such as convection and conduction. Put another way, where this equilibrium state does not exist, radiative heating drives this state toward the equilibrium state. We can represent radiative balance using a series of progressively more complex approximations (more layers, atmospheric constituents, etc.) to the Earth's atmosphere. Characteristics of emissivity, transparency, and opacity are integral to these formulations. For details on a few relatively simplistic models of radiative balance and transfer, please see the Radiative-Convective Equilibrium lecture notes in the further reading section below. The vertical profile of temperature resulting from radiative equilibrium expressed within a reasonably complex version of such a model of radiative balance is presented in the lecture notes. Several problems exist with this profile: the temperature is too warm near the surface, too cold near the tropopause, and decays too rapidly in the troposphere. The profile agrees well with observations in the stratosphere, however.

In addition to radiation, convection is an important contributor to the energetics of the atmosphere. Here, convection is most generically defined as a dry process reflecting the diffusive and advective transport of heat and mass energy by vertical motions. Convective processes play important roles in vertical energy transport and the lateral distribution of clouds and water vapor. As we introduce this concept, it is beneficial to recall concepts of parcel stability in a dry atmosphere. If the temperature lapse rate is greater than the dry adiabatic lapse rate, the atmosphere is said to be unstable to upward parcel displacements. If the temperature lapse rate equals the dry adiabatic lapse rate, the atmosphere is said to be neutral to upward parcel displacements. If the temperature lapse rate is less than the dry adiabatic lapse rate, the atmosphere is said to be stable to upward parcel displacements. Herein, the dry adiabatic lapse rate is equal to 9.8 K km^{-1} and can be derived using the first law of thermodynamics.

As the tropospheric lapse rate from purely radiative equilibrium is greater than the dry adiabatic lapse rate, radiative equilibrium is said to be an *unstable* situation. Thus, we introduce the concept of *radiative-dry convective equilibrium*, wherein such instability is removed by dry convection. In this, radiative and convection are said to counterbalance such that the troposphere is rendered neutral to vertical parcel displacements. Where such neutrality does not exist, radiation and convection act in concert to restore neutrality. The vertical temperature profile resulting from this equilibrium state more closely resembles observations but remains too warm near the surface and too cold at the tropopause.

Of course, the troposphere is not characterized by dry processes alone. Weather would be quite boring if it were! Rather, most atmospheric convection involves the phase changes (condensation, evaporation, sublimation, etc.) of water, the process of which is associated with latent heat release. Also accompanying such phase changes are cloud formation and dissolution, each of which can modify radiative balance through its impacts on both incoming longwave and outgoing shortwave radiation.

As before, principles of thermodynamics can be used to obtain a moist adiabatic lapse rate, one in which the latent heating due to phase changes of water is inherently accounted for in its formulation. This lapse rate varies with altitude and has an approximate value between 6.5-7 K km^{-1}. We again return to our concepts of parcel stability, this time in a moist atmosphere. If the atmospheric lapse rate is greater than the dry adiabatic lapse rate, the atmosphere is unstable to upward parcel displacements. If the lapse rate is between the moist and dry adiabatic lapse rates, the atmosphere is conditionally unstable to upward parcel displacements. If the lapse rate is equal to the moist adiabatic lapse rate, the atmosphere is neutral to upward parcel displacements. Finally, if the lapse rate is less than the moist adiabatic lapse rate, the atmosphere is stable to upward parcel displacements.

The tropospheric temperature lapse rate from radiative-convective equilibrium under dry adiabatic conditions is greater than the moist adiabatic lapse rate. This enables us to introduce the concept of *radiative-moist convective equilibrium*, or hereafter referred to simply as radiative-convective equilibrium, wherein radiative and moist convective heating processes interact to result in an equilibrium or neutral thermodynamic state. This leads to an adjustment of the vertical temperature profile resulting from radiative-dry convective equilibrium to the moist adiabatic lapse rate, resulting in cooler (warmer) near-surface (tropopause) temperatures and much better agreement with observations. The construct of radiative-convective equilibrium is used in many theoretical tropical meteorology studies.

However, this is still not perfect. We know that the atmosphere is rarely, if ever, characterized by neutral stability and a moist adiabatic temperature lapse rate. Meteorological phenomena exist because this is not the case, yet they exist to try to return the atmosphere to such an equilibrium state. More to the point, as we will see shortly, certain physical constraints are violated in the tropics if we extend the concept of radiative-convective equilibrium to the general circulation of the Earth. For instance, note that the latitudinal and seasonal variability in insolation described above results in the tropics (poles) having a net surplus (deficit) of heat energy. This alone results in an excessively large meridional temperature gradient between the equator and poles. Convection acts in part to counteract this imbalance such that latent heat release associated with convection not only locally acts to restore radiative-convective equilibrium but also drives the circulation of the tropics (as demonstrated in previous lectures). These concepts are explored further in the next set of lecture materials.

For Further Reading

- Chapter 1, *An Introduction to Tropical Meteorology, 2nd Edition*, A. Laing and J.-L. Evans, 2011.
- Lecture Notes, Radiative-Convective Equilibrium, K. Emanuel, 2011. (*Full citation: Emanuel, K., 2005:* 12.811 Tropical Meteorology. *(Massachusetts Institute of Technology: MIT OpenCourseWare), http://ocw.mit.edu. License: Creative Commons BY-NC-SA.*)