Radiative-Convective Equilibrium and Self-Aggregation

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 by introducing the concept of radiative-convective equilibrium, and describe how instabilities in the equilibrium state can give rise to the self-aggregation of cumulus clouds in the tropics. 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, radiative-convective equilibrium, and self-aggregation, and why do they matter?

Energetics of the Earth System and Radiative Balance

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, often referred to as *insolation* (portmanteau 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, including by water vapor, and the diffusion of solar energy due to the angle at which it intersects the atmosphere. 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 $\varepsilon = 1$) to the energy emitted from that surface, as the following:

$$F = \sigma T^4 \tag{1}$$

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 S_0 = 1370 W m^{-2}. To first order, the total absorbed solar radiation can be expressed by:

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

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 lateral transports. 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. The vertical profile of temperature resulting from radiative equilibrium expressed within a reasonably complex radiative transfer model is presented in the lecture materials. 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 vertical heat and moisture transport by diffusion and advection. 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 phase changes (condensation, evaporation, freezing, melting, etc.) of water, the process of which is associated with latent heat release. Also accompanying such phase changes are cloud formation and dissipation, 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} (higher near the surface, where saturation mixing ratio – and thus moisture content – is typically higher, and thus so is latent heating when water changes phases). We 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 lapse rate from radiative-dry convective equilibrium 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 and is the basis for the "large-scale control" class of numerical model cumulus parameterizations, wherein moist convection on short time scales is said to result from – and consume – the instability generated primarily by radiative processes on longer time scales.

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. Moist 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 tropical circulations.

Self-Aggregation

At first glance, it would seem to follow that radiative-convective equilibrium in a horizontally homogeneous environment (neglecting temperature contrasts between water and land, oceanic circulations, and under the weak temperature gradient approximation) would result in a homogenous distribution of moist convection to maintain the equilibrium state. This is an overly simplistic view, however: the surface is heterogeneous; oceanic circulations exist; and yet, even neglecting these factors, the real atmosphere is characterized by finite-amplitude perturbations of various scales and magnitudes. Consequently, even in a radiative-convective equilibrium-like environment, tropical moist convection does not exist uniformly, but rather in a localized, organized manner that develops quasi-spontaneously. This organization is known as *self-aggregation*, and it is an active area of research in the field given its potential importance to driving tropical circulations and sensitivity to climate variability.

Our consideration of self-aggregation begins, and largely ends, with first principles. Both theory and observations suggest that moist convection preferentially occurs where near-surface temperature and moisture are relatively high, given their controls on CAPE and CIN, as well as where tropospheric water vapor content is relatively high. This latter state can be accomplished by surface latent heat, or moisture, fluxes; lateral transports, or horizontal advection; and by detrainment from existing moist convection – i.e., the expulsion and subsequent evaporation or sublimation of liquid or frozen water from a cloud, often in the middle-to-upper troposphere. In turn, tropospheric water vapor content and clouds exert a control on radiative transfer for both shortwave and longwave radiation; for example, longwave cooling is mitigated in regions of high water vapor content and cloudiness. Further, the local circulations generated by existing moist convection – given the relationship between diabatic heating, horizontal divergence, and cyclonic vorticity established in a previous lecture – can also enhance surface heat and moisture fluxes (given their dependence on surface wind speed), maintaining near-surface warmth and moisture even in the presence of precipitation.

This feedback loop, a manifestation of a departure from or instability in the radiative-convective equilibrium state, favors continued moist convection where it initiated, further moistening the air column. In turn, it also promotes the continued absence of moist convection elsewhere; here, moist convection is initially suppressed by subsidence outside of the adjacent moist convection, in turn affecting water vapor concentration and radiative transfer in the opposite sense to that described above for moist convection. Over time, individual convective clusters merge, in part aided by outflow boundary collisions, resulting in self-aggregation. An animation of this process in a simplified numerical model is provided in the accompanying lecture materials. It should be noted, however, that we have assumed an environment absent of, for instance, vertical wind shear (which can be viewed as the result of the weak temperature gradient approximation, at least in as much as thermal wind balance holds in the tropics), which mitigates against self-aggregation. A full treatment of the limitations of the arguments advanced herein is beyond the scope of this class, however.

For Further Reading

- Chapter 1, *An Introduction to Tropical Meteorology, 2nd Edition*, A. Laing and J.-L. Evans, 2016.
- 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.*)
- Bretherton, C. S., P. N. Blossey, and M. Khairoutdinov, 2005: An energy-balance analysis of deep convective self-aggregation above uniform SST. *J. Atmos. Sci.*, **62**, 4273–4292.
- Emanuel, K., A. A. Wing, and E. M. Vincent, 2014: Radiative-convective instability. *J. Adv. Model. Earth Sys.*, **6**, 75–90.
- Sobel, A. H., J. Nilsson, and L. M. Polvani, 2001: The weak temperature gradient approximation and balanced tropical moisture waves. *J. Atmos. Sci.*, **58**, 3650–3665.