Fall, 2019

Geophysical Fluid Dynamics II: Large Scale Dynamics

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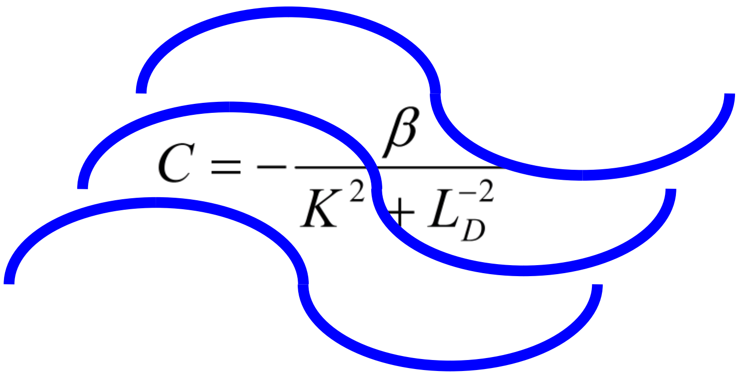

An advanced theoretical course for 1st — 2nd year graduate students majored in atmospheric and oceanic sciences

Last Revised Thursday, August 29, 2019

Syllabus

Geophysical Fluid Dynamics II: Large Scale Dynamics[1]

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Chapter 0: Introduction (9.9, 1 time)

Part I: Dynamics of Shallow Water System

Chapter 1: Basics (3 weeks, 9.11-10.7, 6 times)

Sec.1.1: Basic equations
Sec.1.2: Conservation laws*
Sec.1.3: Circulation, vorticity and Kelvin's Theorem*
Sec.1.4: Potential vorticity conservation
Sec.1.5: Shallow water waves on an f-plane
Sec.1.6: Geostrophic adjustment

Chapter 2: Shallow Water Rossby Wave Dynamics (3 weeks, 10.9-10.28, 6 times)

Sec.2.1: Quasi-geostrophic equation
Sec.2.2: Rossby waves
Sec.2.3: Group velocity and energy propagation
Sec.2.4: Reflection and normal modes
Sec.2.5: Forced waves
Sec.2.6: Non-plane waves*

Chapter 3: Forced Circulation (1.5 weeks, 10.30-11.6, 3 times)

Sec.3.1: Atmospheric circulation
Sec.3.2: Ekman dynamics
Sec.3.3: Sverdrup flow
Sec.3.4: Rossby wave and ocean circulation

Mid-term Exam (11.11)

1 Most sections require both understanding and derivation of the equations. Sections with * require understanding only. Sections with** are not required.

Part II: Dynamics of Stratified Flow

Chapter 4: Basics of Stratified Fluid (1.5 weeks, 11.13-11.20, 3 times)

Sec.4.1: Basic equations
Sec.4.2: Vorticity equation
Sec.4.3: Ertel potential vorticity*

Chapter 5: Rossby Wave Dynamics (2 weeks, 11.25-12.4, 4 times)

Sec.5.1: Quasi-geostrophic equation for stratified flow*
Sec.5.2: Rossby waves in stratified fluid
Sec.5.3: Vertical normal modes
Sec.5.4: The Elliassan-Palm theorem*

Chapter 6: Instability Theory (discussion. 2 weeks, 12.9-12.18, 4 times)

Sec.6.1: Instability problem
Sec.6.2: Baroclinic instability in a two-layer QG model
Sec.6.3: Energetics
Sec.6.4: Charney-Stearn theorem*
Sec.6.5: The Eady problem*
Sec.6.6: Barotropic instability*

Final-term Exam (12.23)

Grading:

30% homework, 10% lectures, 10% formula visualization
20% mid-term exam, 30% final term exam

References:

1: Pedlosky, J. Geophysical Fluid Dynamics (2^{nd} ed, 1987)**,** *Springer-Verlag***.**

2: Gill, A. E., Atmosphere-Ocean Dynamics, 1981, *Academic Press*.

3: Holton, J. R., An Introduction to Dynamic Meteorology (3^{rd} edition), *Academic Press*.

Ch. 0: Introduction

Why do we study dynamics? The ultimate forcing of the atmospheric and oceanic circulation is the solar radiation. The ocean and atmosphere systems adjust themselves such that the system is in a dynamic equilibrium.

1. Radiative Equilibrium

At the first order, the equilibrium is the so called radiative equilibrium. At this state, the incoming solar radiation warms the surface of the earth. The surface is cooled by the outgoing long wave radiation into the space. Since the atmosphere absorbs mainly long wave radiation from the surface, the atmosphere is heated from below. This first picture of the atmospheric equilibrium, although simple, can explain some essential features of the observed atmosphere stratification. For example, air temperature decreases away from the surface. The surface temperature is about 200^oK to 300^oK from the high to low latitudes.

This radiative equilibrium, however, still differs substantially from the observation in several important aspects. First, the vertical change of air temperature is too large: 150 K from surface to 10 Km as opposed to about 70K in the observation; (Fig.0.1). Second, the latitudinal temperature gradient is also too large, about 100 K from pole to equator, as opposed to about 50 K in the observation (Fig.0.2). Therefore, the pure radiative equilibrium atmosphere has temperature gradients about twice that of the observation in both the vertical and latitudinal directions. Dynamic processes transport heat (Fig.0.3) to reduce the mean temperature gradients and therefore play important roles in maintaining the mean climatology.

In comparison with the atmosphere, the dynamic processes are even more important in the ocean. This is because radiation only penetrates the surface ocean water by about 10 meters or so. The stratification of the entire subsurface ocean (Fig.0.4), therefore, at the first order, is determined by oceanic dynamic transports.

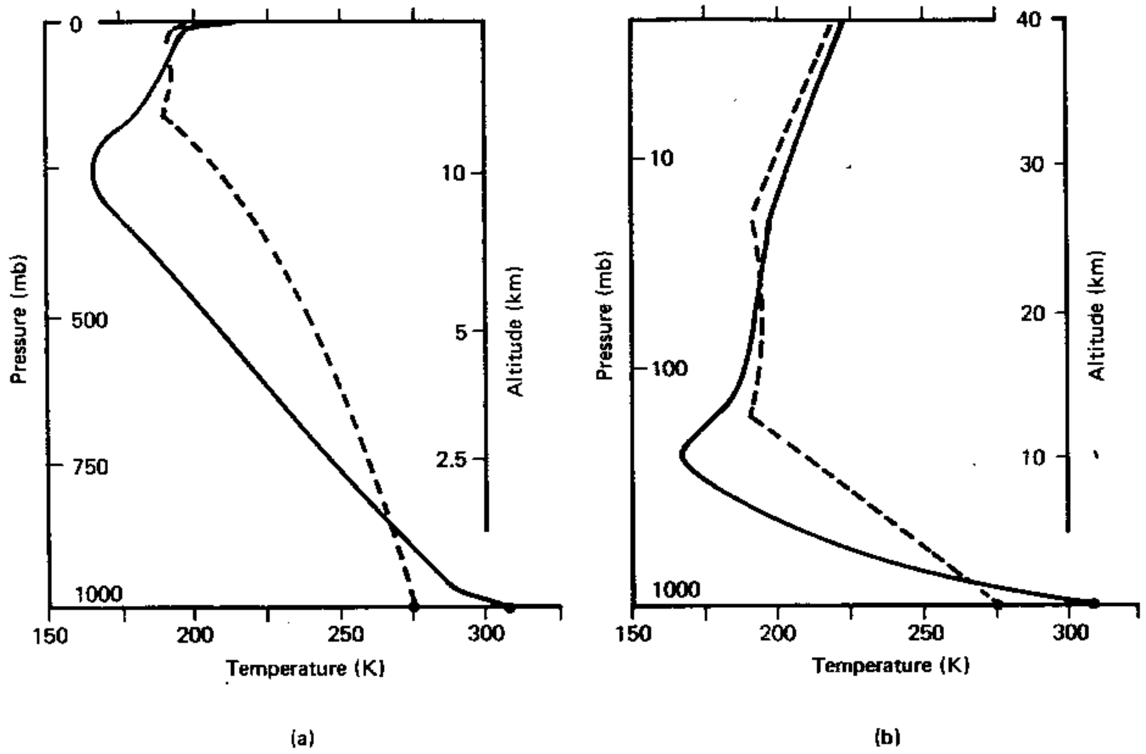


Fig. 1.4. The radiative equilibrium solution (solid line) corresponding to the observed distribution of atmospheric absorbers at 35°N in April, the observed annual average insolation for the whole atmosphere, and no clouds. The dashed line shows the effect of convective adjustment to a constant lapse rate of 6.5 K km^{-1}. In (a) the curves are drawn with a scale linear in pressure, i.e., equal intervals correspond to equal masses of atmosphere. In (b) the scale is linear in altitude. [From Manabe and Strickler (1964, Fig. 4).]

Figure 0.1 Radiative and radiative convective equilibrium in the atmosphere

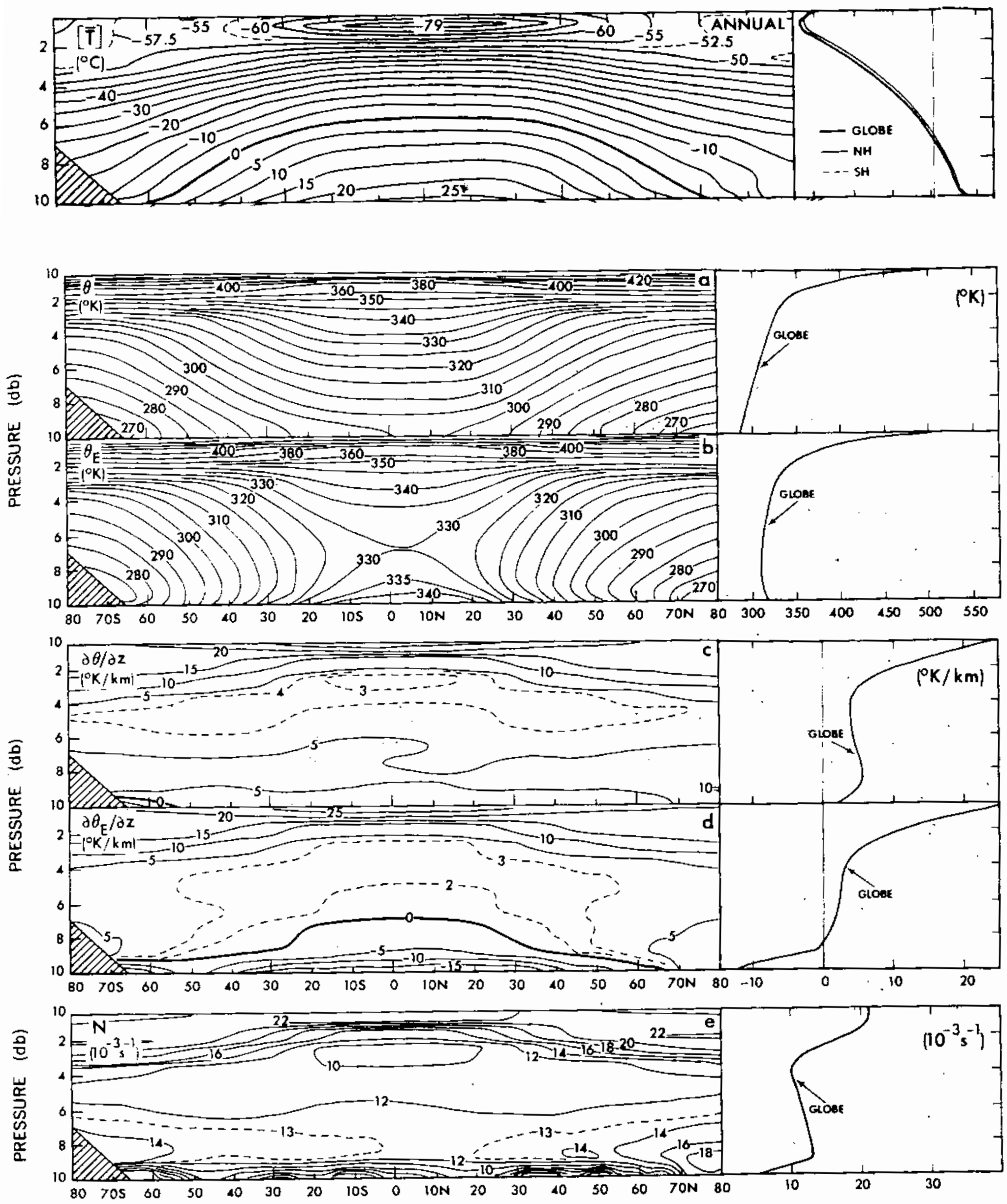


FIGURE 7.6. Zonal-mean cross sections of the potential temperature (a) in K, equivalent potential temperature (b) in K, vertical gradient of potential temperature (c) in K/km, vertical gradient of equivalent potential temperature (d) in K/km, and Brunt–Väisälä frequency in 10^{-3} rad s^{-1} for annual-mean conditions. Vertical profiles of the global mean values are shown on the right.

Figure 0.2 Zonal mean potential temperature of the atmosphere

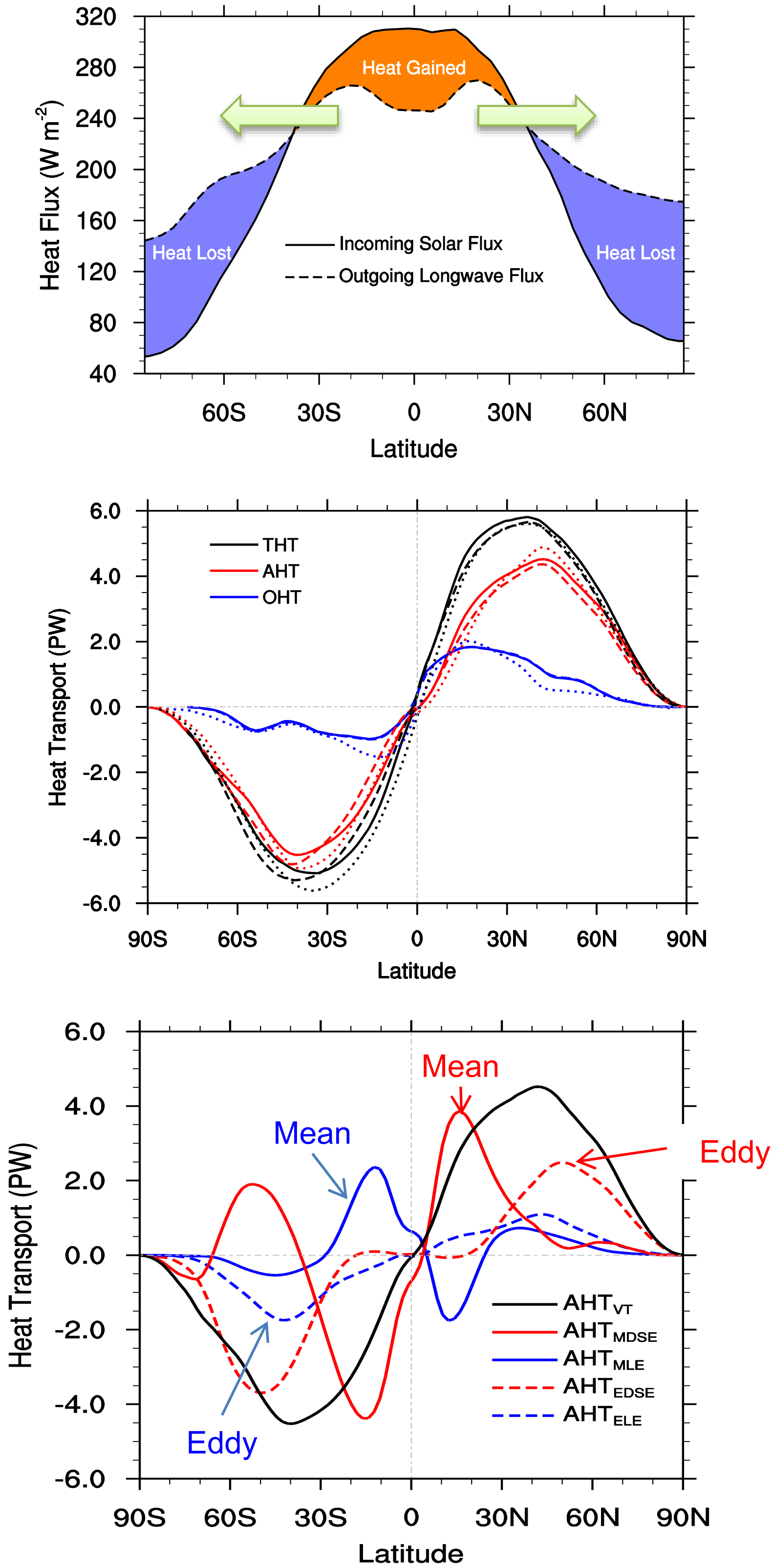


Figure 0.3 Latitudinal distribution of radiation forcing and heat transports in the atmosphere and ocean

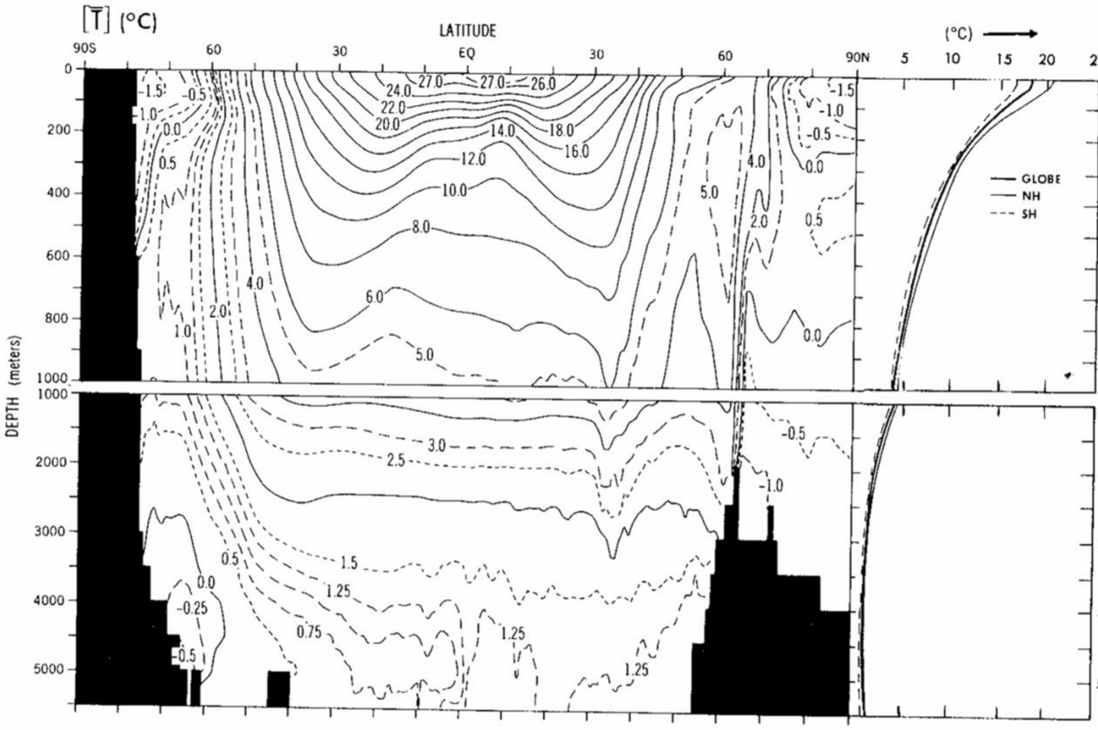
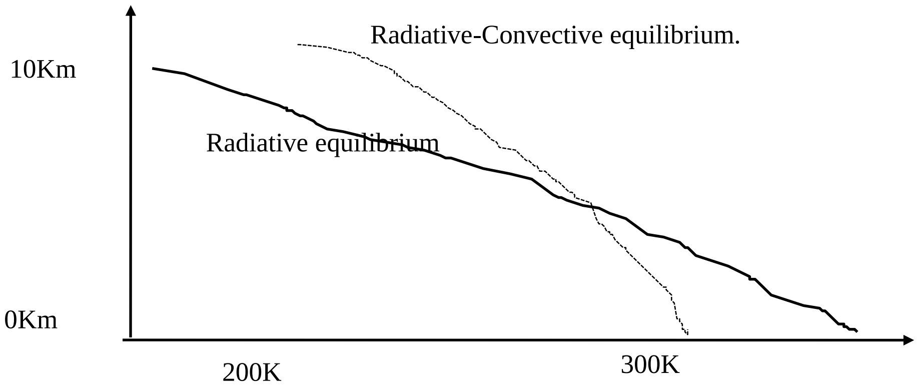


FIGURE 8.4. Zonal-mean cross section of temperature for the top ocean layer 0–1000 m and the layer below 1000 m in °C for annual-mean conditions. Vertical profiles of the hemispheric and global mean temperatures are shown on the right (after Levitus, 1982).

Figure 0.4 Zonal mean temperature of the ocean

2. Radiative-convective equilibrium (vertical)

In the atmosphere, at the next order is the convective. The cold air is too heavy in the upper atmosphere such that the air column becomes convectively unstable. Convection reduces the vertical temperature gradient significantly to a neutral stable stratification – radiative-convective equilibrium (about $6.5^oK/1km$ in the dry atmosphere) (Fig.0.1)

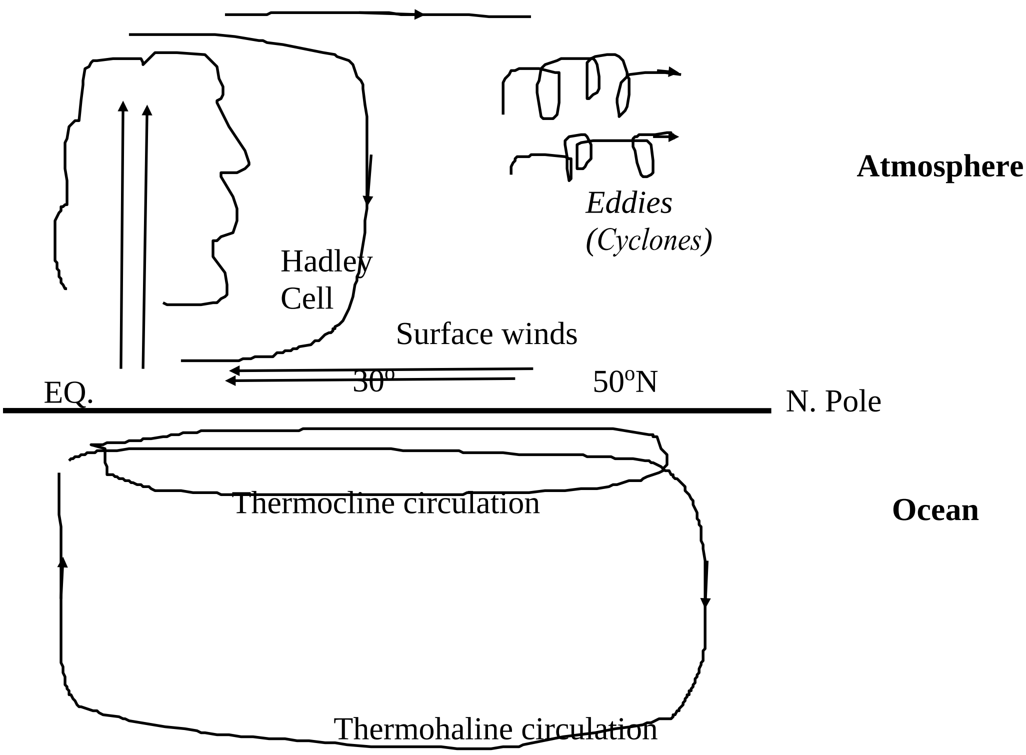


Convection also plays important role in setting the deep stratification at the high latitude ocean, and in turn the deep circulation and stratification. Different from the atmosphere, however, oceanic convection alone adjusts the water column in the subsurface ocean to convective equilibrium, rather than convective-radiative equilibrium.

3. Transport: dynamics (horizontal)

Even after vertical dynamic process, we are still left with important questions. In the atmosphere, the latitudinal gradient of temperature in the atmosphere is still too large. Even worse is the ocean. Neither the horizontal nor the vertical temperature distribution can be explained by the local equilibrium.

The global adjustment will be accomplished by the horizontal transport in the atmosphere and ocean – the so called dynamic processes. The heat transport is partitioned about equally between the atmosphere and ocean (Fig.0.3). Like the vertical convective process that is trying to bring the system to gravitational stability, the horizontal transport of heat eventually bring the climate system to some sort of equilibrium. As a result, the horizontal temperature gradient will also be reduced greatly.



In general, the atmospheric process involves the Hadley Cell (Fig.0.5) in the low latitude and storm track activity in the midlatitude. The oceanic process involves shallow thermocline processes and the thermohaline process (Fig.0.6). The former is wind-driven while the latter is buoyancy driven.

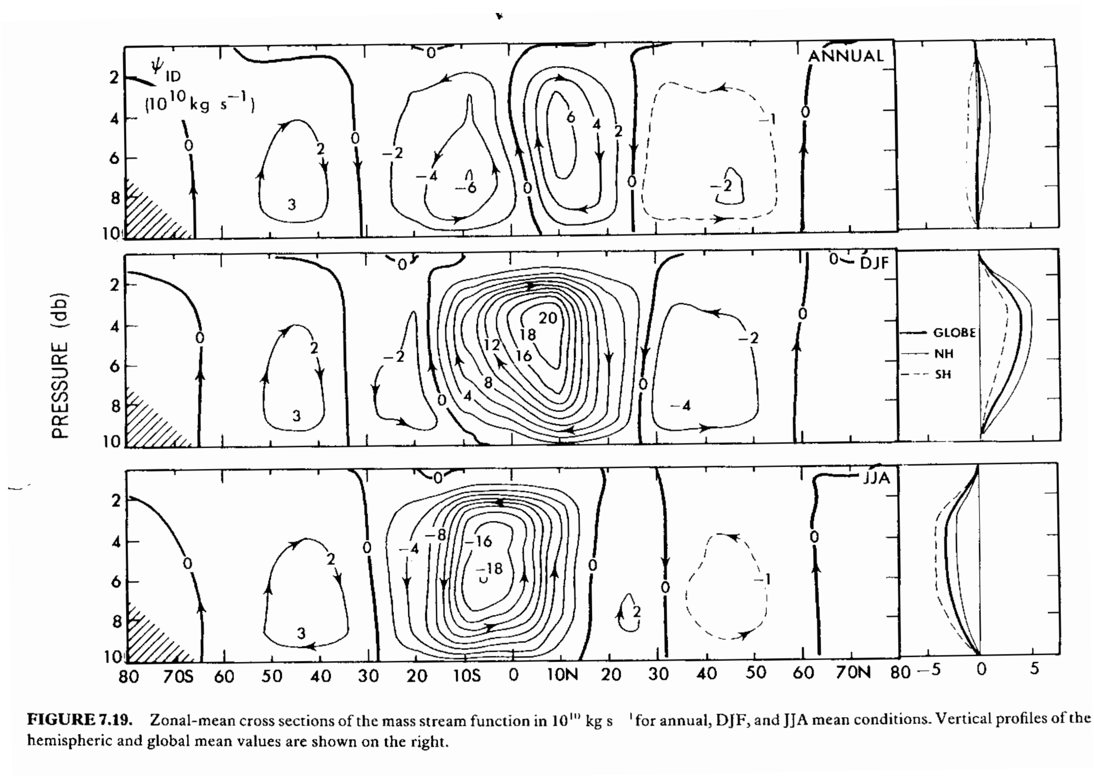


Fig.0.5 Zonal mean circulation in the atmosphere

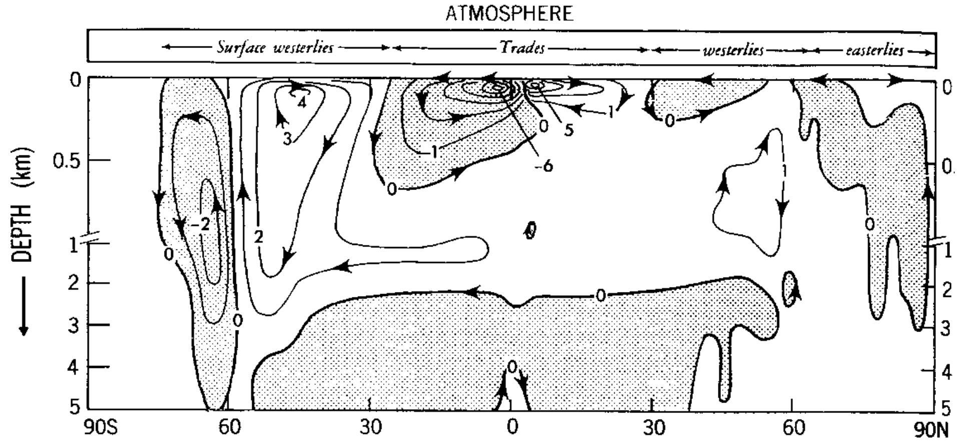


FIGURE 8.20. Zonal-mean cross section of the annual-mean flow of mass in 10^{10} kg s^{-1} for a ocean model (adapted from Bryan and Lewis, 1979).

Figure 0.6 Zonal mean circulation in the ocean

4. Comparison of atmospheric and oceanic dynamics

The dynamics in the atmosphere and ocean have many similarities and differences. We should try to keep in mind of both systems. This will give us a deeper understanding of both systems.

a) Similarities:

Both systems are affected by: i) rotation, ii) stratification, iii) global scale processes. It is these similarities that makes the GFD here in principle apply to both systems, and makes the GFD different from classical fluid dynamics.

b) Differences:

	Atmosphere	**Ocean**
Forcing	pure buoyancy driven	wind and buoyancy driven
Horizontal Domain	no horizontal boundary	with horizontal boundary
Vertical	No upper boundary	Bounded vertically
Weight, speed	light and fast	heavy and slow
In addition to temperature	water vapor	salinity
radiation	Radiative critical	Dynamics critical
stratification	strong	weak
compressibility	Compressible	incompressible
	….	….