Introduction to Climate

~ Part I ~

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Outline of the lecture

1. Climate System (90 min. + α)
   1.1 Introduction
   1.2 Radiative Balance
   1.3 Horizontal Radiative Imbalance and Circulations
   1.4 Seasonal Change
   1.5 Role of Orography on Climate
2. Climate Variability (120 min. + α)
   2.1 Introduction
   2.2 Intraseasonal Variability: Quasi-stationary Rossby wave, MJO and equatorial waves
   2.3 Interannual Variability: ENSO, El Nino Modoki, IOD
   2.4 Decadal Variability: PDO, ENSO-Monsoon relation
3. JMA's latest one-month prediction

1. Climate System

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   - 1.2 Radiative Balance
   - 1.3 Horizontal Radiative Imbalance and Circulations
   - 1.4 Seasonal Change
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1.1 Introduction

Climate and Climate System

"Weather is what is happening to the atmosphere at any given time.

Climate in a narrow sense is the "average weather," the statistical description over a period of time."

Climate is formed in the interactions in climate system, consisting of atmosphere including composition and circulation, the ocean, hydrosphere, land surface, biosphere, snow and ice, solar and volcanic activities in its spatial and temporal variability.

Causes of Climate Variability

- **Natural origin**

  external: land-sea distribution, orography
  solar constant, orbital variations
  volcano

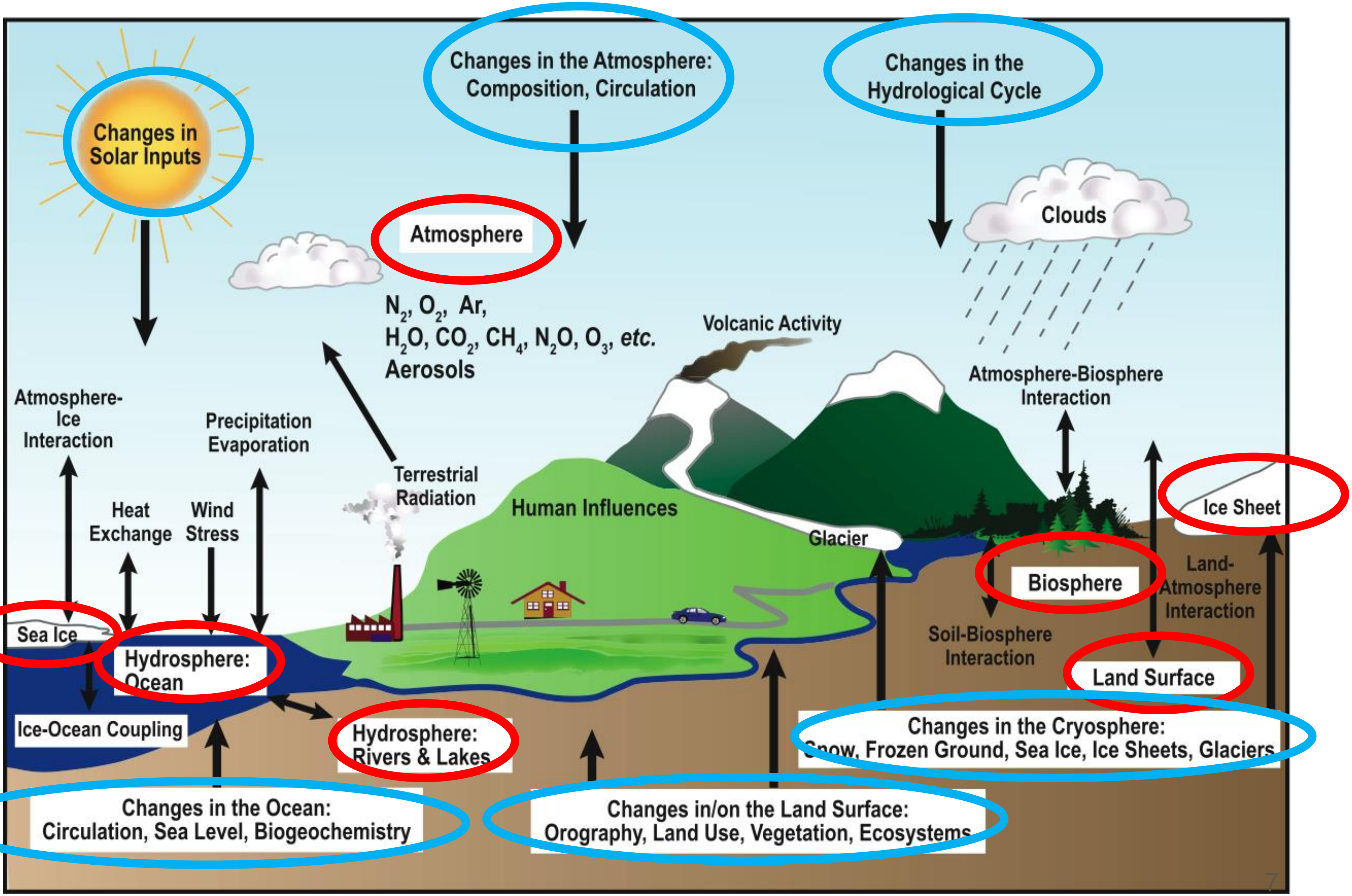
  internal variability of the climate system
  (e.g., air-sea interaction,,,)

- **Anthropogenic origin**

  emission of greenhouse gases, destruction of ozone layer, land surface modification,,, (= climate change)

Climate System

http://ipcc-wg1.ucar.edu/wg1/FAQ/wg1_faq-1.2.html

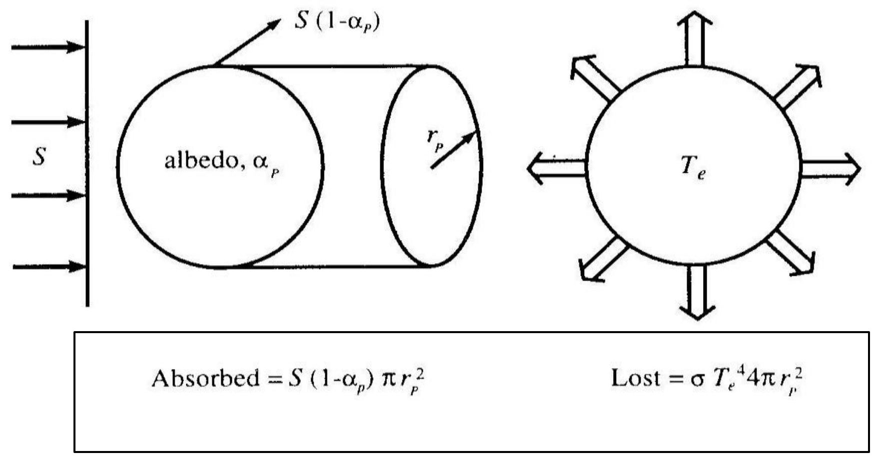


Schematic view of the components of the climate system, their processes and interactions.

1.2 Radiative Balance

Radiative Balance between Earth and Space

- **Solar Radiation**
  - Solar constant : $S \approx 1370\,Wm^{-2}$
  - insolation : $\frac{S}{4} \approx 342\,Wm^{-2}$
  - planetary albedo : $\alpha_P \approx 0.31$
  - absorption : $\frac{S}{4}(1-\alpha_P) \approx 235\,Wm^{-2}$
- **Terrestrial Radiation**
  - equilibrium radiative temperature : T_e^*
  - emission : σT_e^{*4}



- **Radiation Balance**
  - Balance between absorbed solar radiation and terresitrial emission

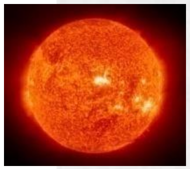
$$T_e^* = \sqrt[4]{\frac{S(1-\alpha_P)}{4\sigma}} \approx 254K \approx -19°C$$

Earth's temperature observed from the cosmic space

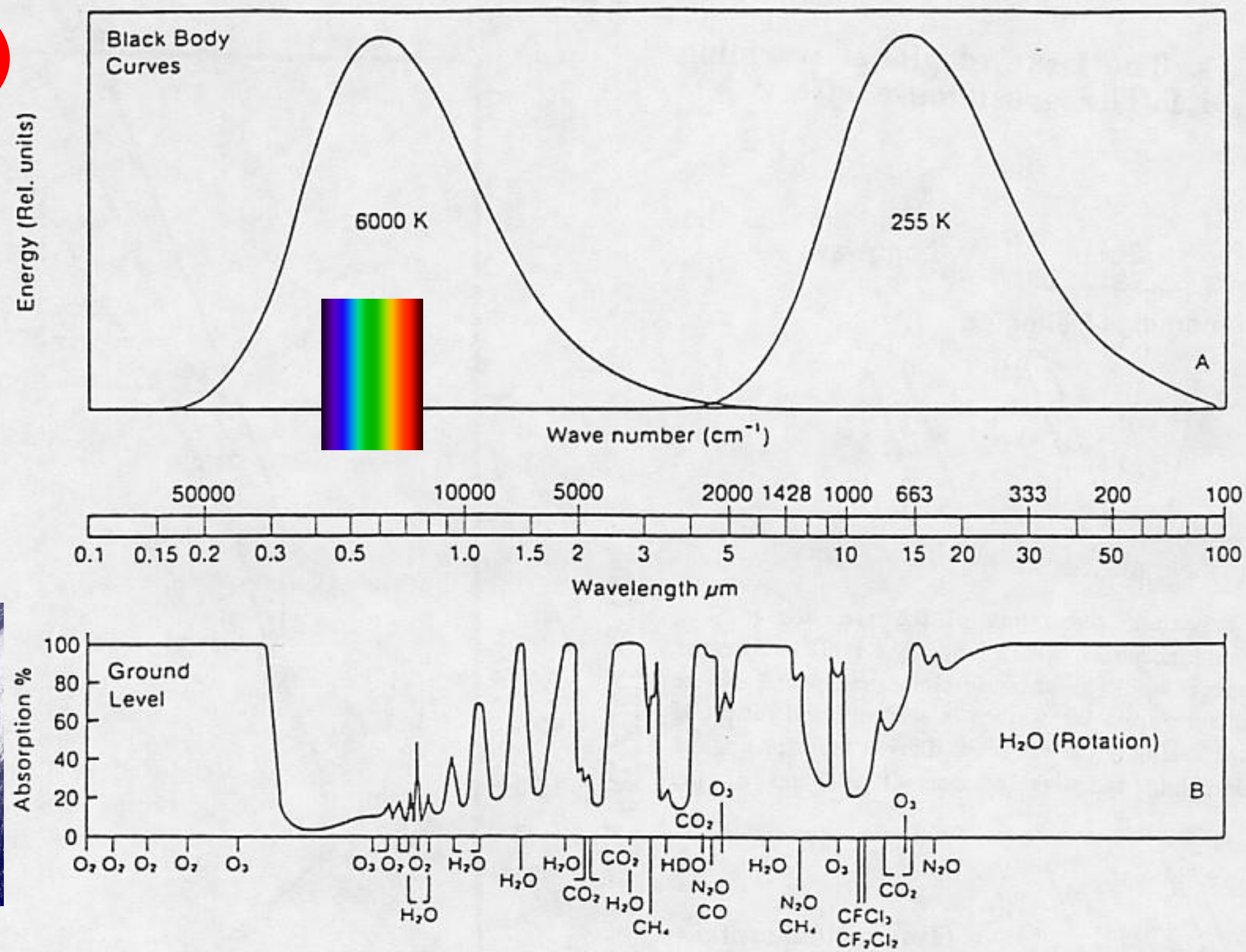
Surface Temperature = 15° C

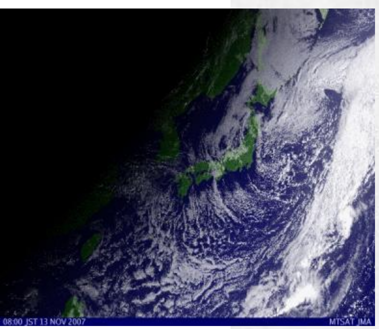
Absorption of Radiation from 6000K and 255K Blackbodies

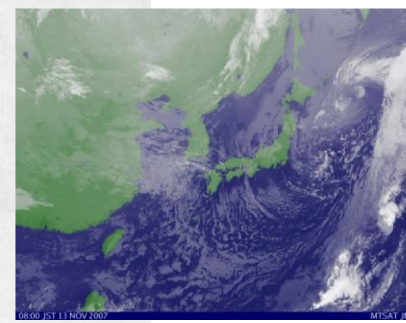
Solar radiation



Terrestrial radiation

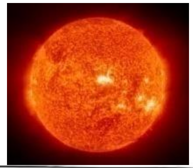






(*a*) Spectral distribution of long-wave emission from blackbodies at 6000 K and 255 K, corresponding to the mean emitting temperatures of the Sun and Earth, respectively, and (*b*) percentage of atmospheric absorption for radiation passing from the top of the atmosphere to the surface. Notice the comparatively weak absorption of the solar spectrum and the region of weak absorption from 8 to 12 μm in the long-wave spectrum [from *MacCracken and Luther*, 1985].

The simplest greenhouse model



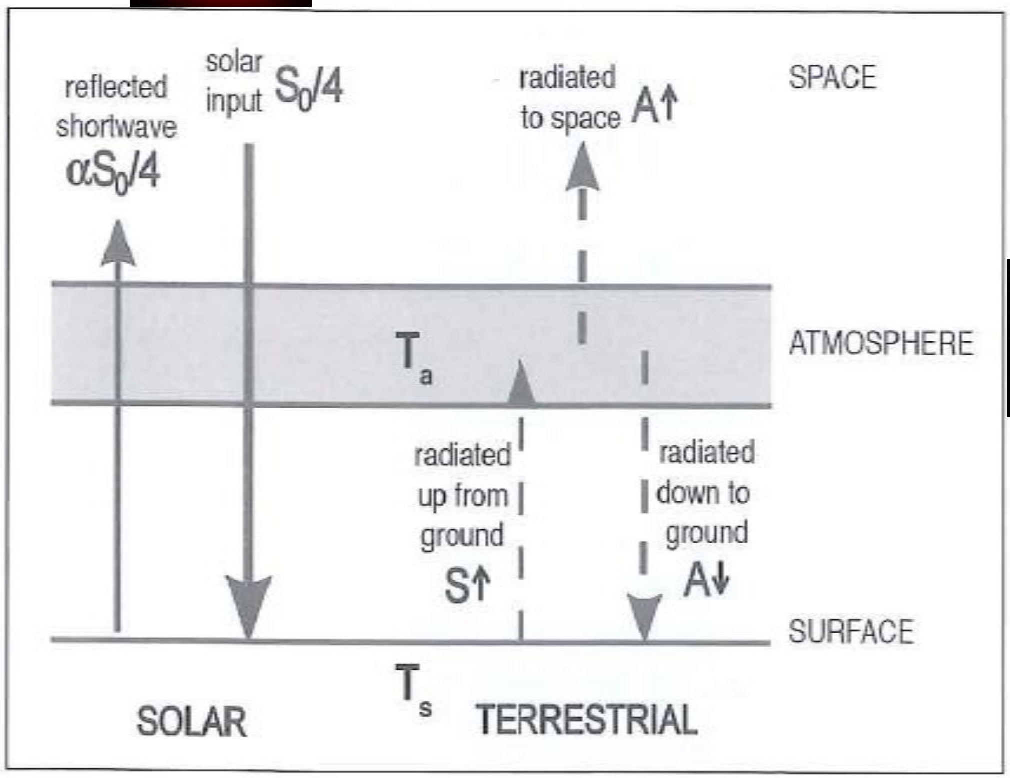


FIGURE 2.7. The simplest greenhouse model, comprising a surface at temperature T_s, and an atmospheric layer at temperature T_a, subject to incoming solar radiation $S_o/4$. The terrestrial radiation upwelling from the ground is assumed to be completely absorbed by the atmospheric layer.

Pictures are from NASA web-sites

Difference between Equilibrium radiative temperature and Ground Surface Temperature

- **Venus**
  - **Solar constant : 2600 W/m2**
  - **planetary albedo : 0.77**
  - **Equilibrium radiative temperature : - 46° C**
- **Earth**
  - **Solar constant :** $S \approx 1370\, Wm^{-2}$
  - **planetary albedo :** $\alpha_P \approx 0.31$
  - **Equilibrium radiative temperature**

$$T_e^* = \sqrt[4]{\frac{S(1-\alpha_P)}{4\sigma}} \approx 254K \approx -19°C$$

- **Mars**
  - **Solar constant : 590 W/m2**
  - **planetary albedo : 0.15**
  - **Equilibrium radiative temperature : -56° C**

Surface Temperature

457° C

Surface Pressure

90,000hPa

Surface Temperature

15° C

Surface Pressure

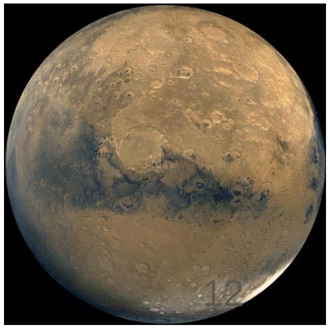
1,000hPa

Surface Temperature

-55° C

Surface Pressure

10hPa



Radiative heating tends to create vertical instability

between heated ground and cooled atmosphere on average

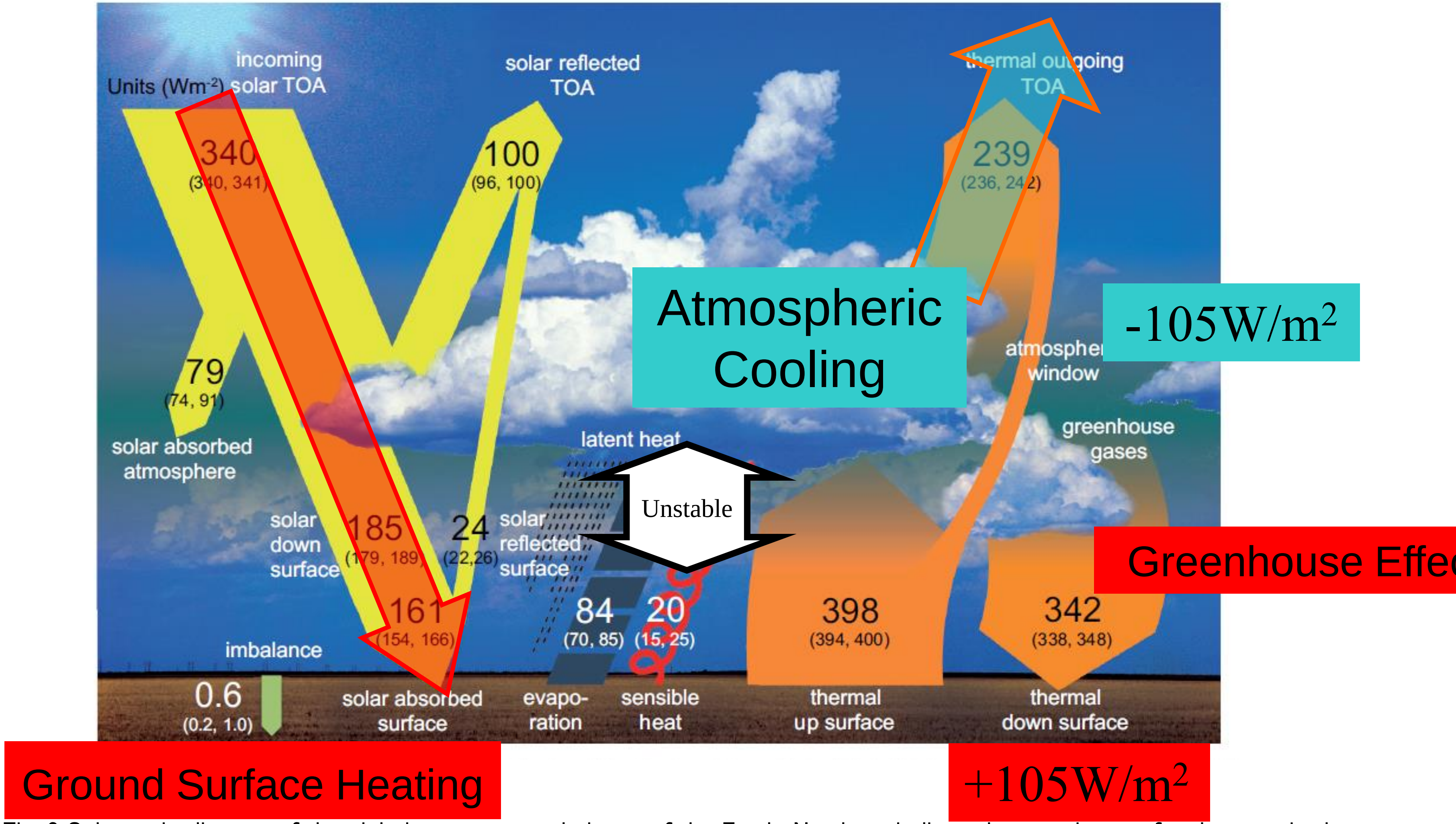


Fig. 2 Schematic diagram of the global mean energy balance of the Earth. Numbers indicate best estimates for the magnitudes of the globally averaged energy balance components together with their uncertainty ranges, representing present day climate conditions at the beginning of the twenty first century. Units Wm−2. Source: Wild et al.(2013.)

One-Dimensional Vertical Profile Model

Thermal Equilibrium of the Atmosphere with a Convective Adjustment

SYUKURO MANABE AND ROBERT F. STRICKLER

General Circulation Research Laboratory, U. S. Weather Bureau, Washington, D. C.

(Manuscript received 19 December 1963, in revised form 13 April 1964)

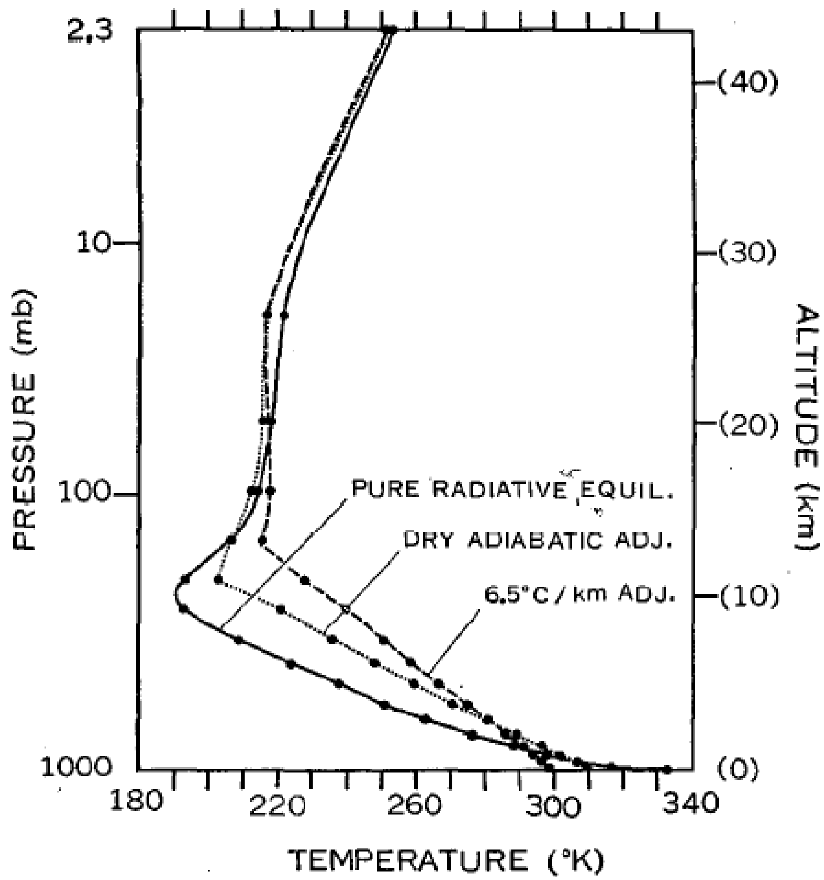


FIG. 4. The dashed, dotted, and solid lines show the thermal equilibrium with a critical lapse rate of 6.5 deg km^{-1}, a dry-adiabatic critical lapse rate (10 deg km^{-1}), and pure radiative equilibrium.

Typical condition at 35N in April

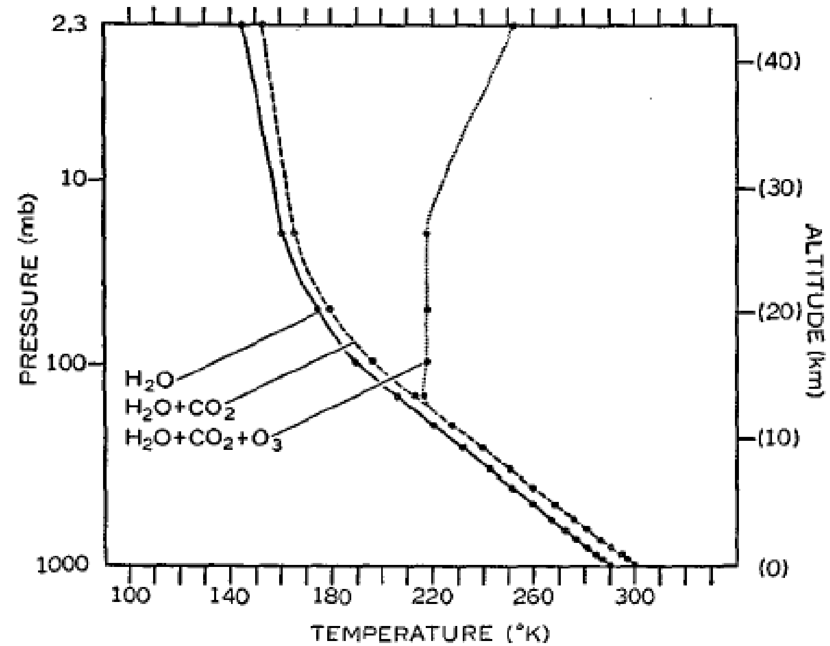


FIG. 6c. Thermal equilibrium of various atmospheres which have a critical lapse rate of 6.5 deg km^{-1}. Vertical distributions of gaseous absorbers at 35N, April, were used. $S_c = 2$ ly min^{-1}, $\cos\bar{\zeta} = 0.5$, $r = 0.5$, no clouds.

One-Dimensional Vertical Profile Model

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Observed Temperature

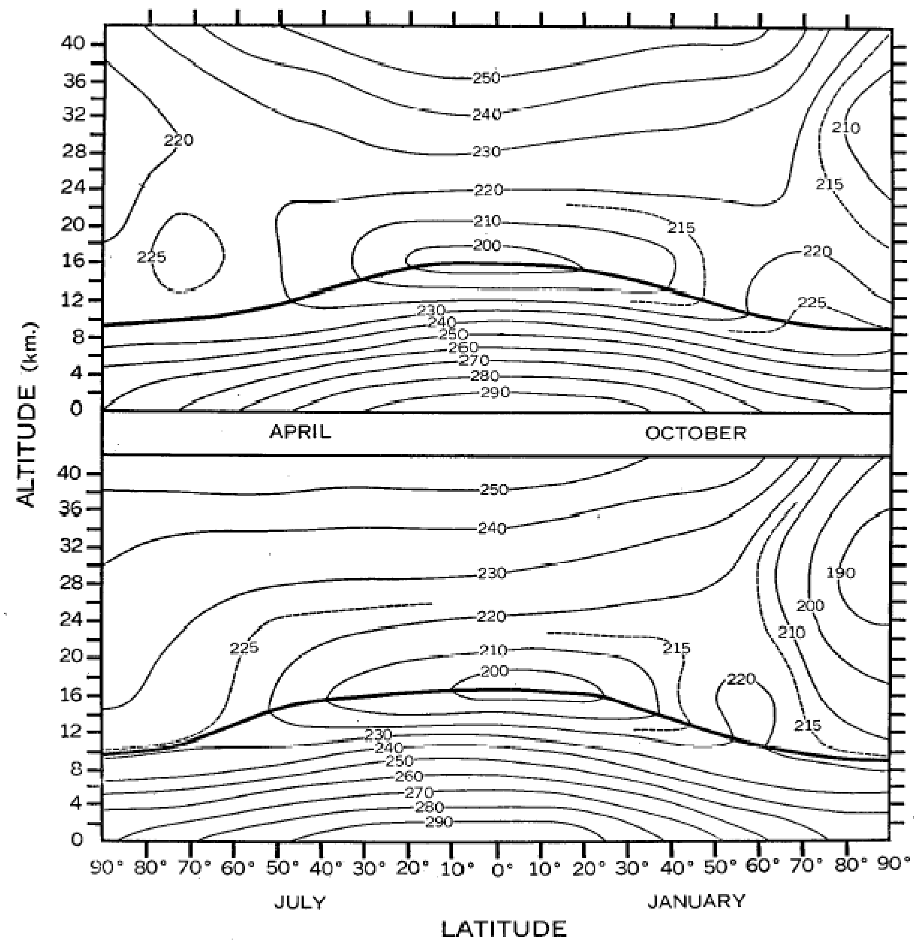


FIG. 12. Distribution of the observed temperature (deg k) in the northern hemisphere for different seasons. From J. London (1956).

1-D model Simulations for each latitudes

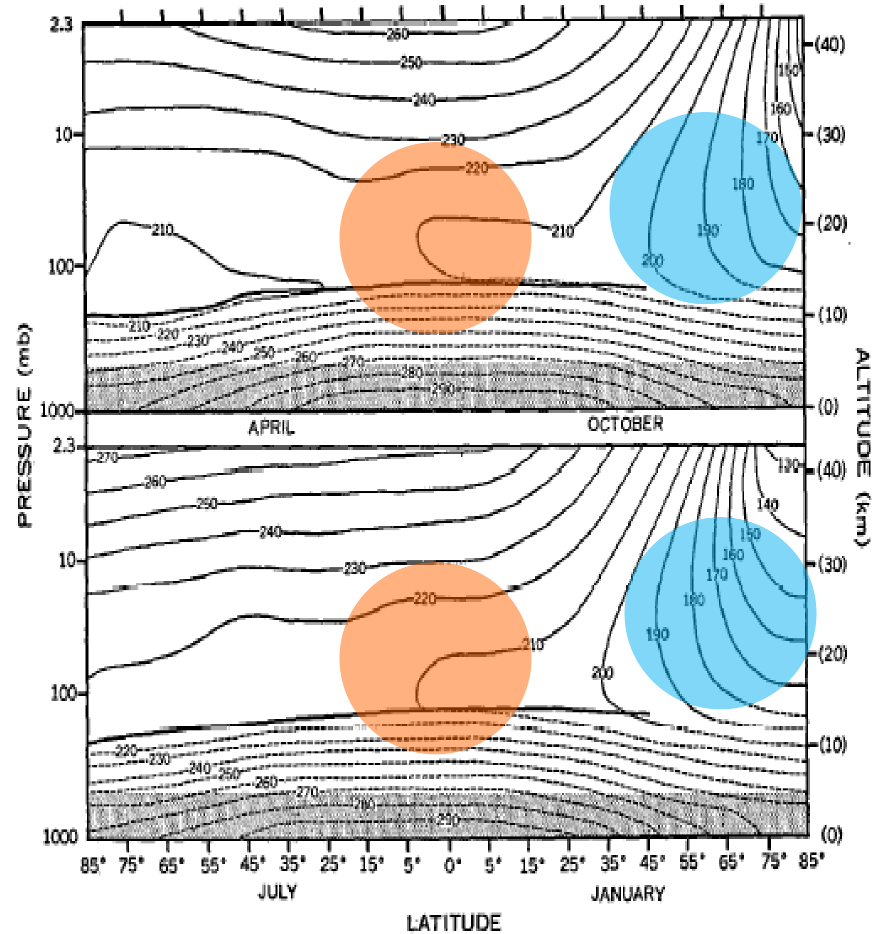


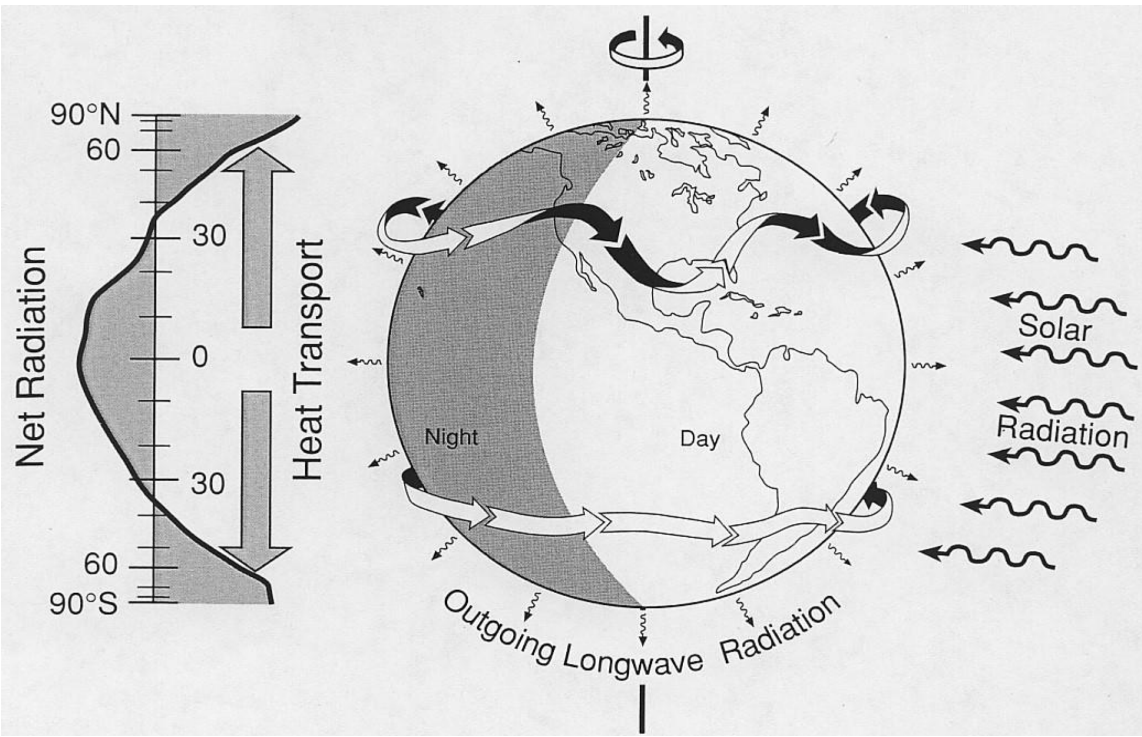
FIG. 13. The local radiative equilibrium temperature of the stratosphere. The shaded area is the region where the temperature was fixed at the observed value. Above the shaded area the state of the convective equilibrium, whose critical lapse rate for convective adjustment is deg km^{-1}, is shown. The region covered by solid lines is in local radiative equilibrium.

1.3 Horizontal Radiative Imbalance and Circulations

Imbalanced horizontal distribution of radiative heating

(1) Latitudinal Imbalance between Pole and Tropics Timescale=1year

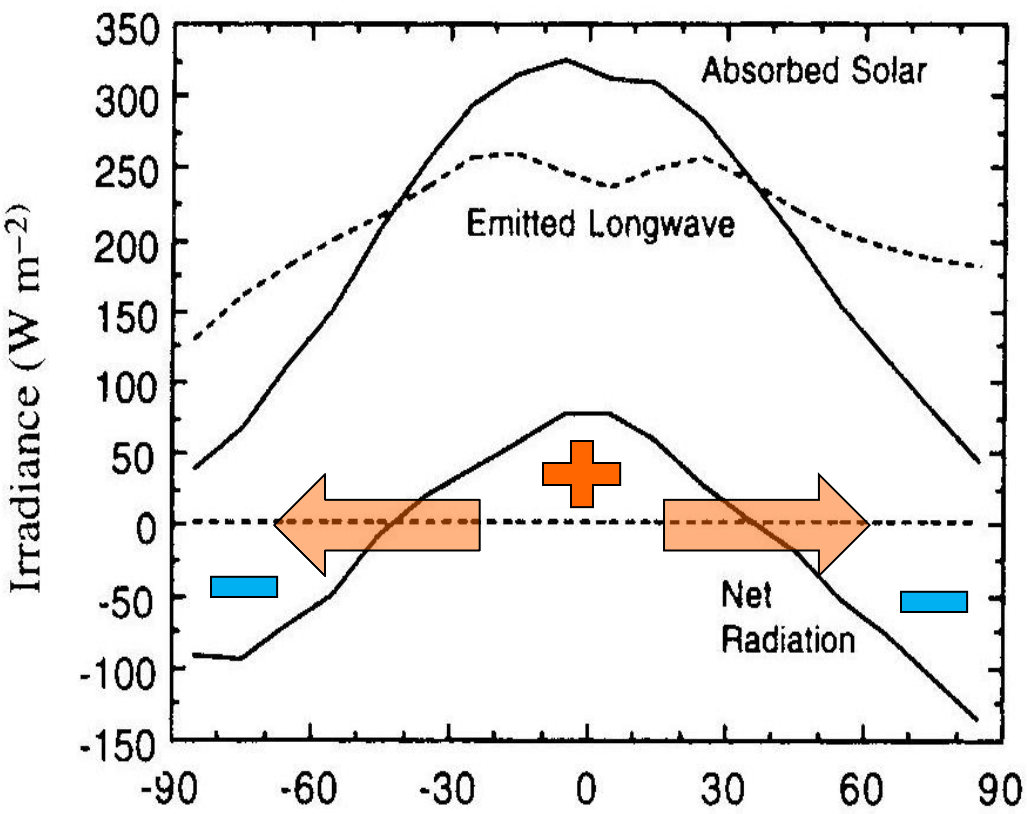
Driving Forces of Climate



- ⌘ Relaxation time to radiative equilibrium temperature (radiative equilibrium timescale) is estimated as about 30 days.
- ⌘ Radiative imbalance between Pole and Tropics drives global circulations.
- ⌘ Radiative imbalance between day and night has small influence on global circulations directly.

(2) Longitudinal Imbalance between Day and Night Timescale=1day

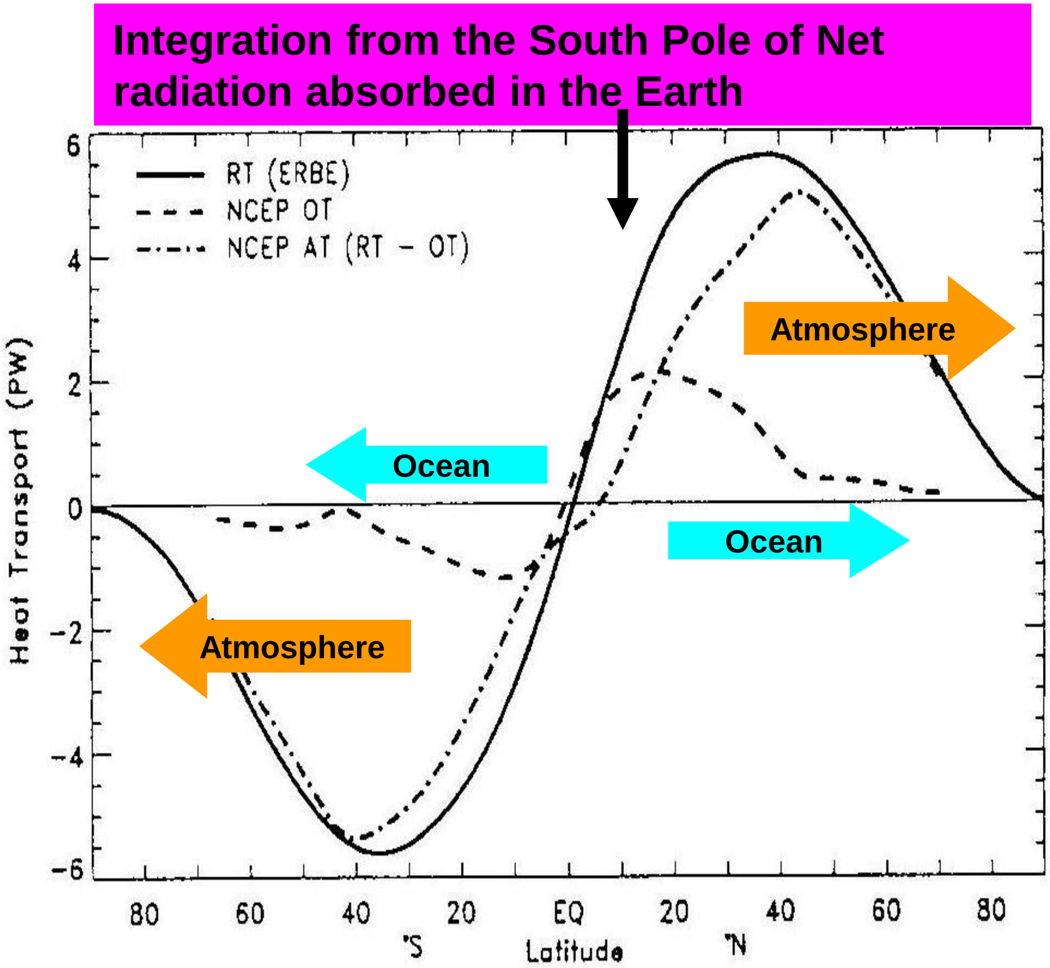
Meridional distribution of Annual mean radiation balance



Hartmann (1994)

- **Solar radiation**
  - **Global mean : 235 Wm^{-2}**
  - **Low latitude : over 300 Wm^{-2}**
  - **Poles : about 50 Wm^{-2}**
- **Terrestrial radiation**
  - **Global mean : 235 Wm^{-2}**
  - **Less gradient between low latitudes and poles compared to that in solar radiation**
- **Net radiation**
  - **Global mean : 0 Wm^{-2}**
  - **Positive in low latitudes, negative in high latitudes**
- **Poleward heat transport by the atmosphere and ocean balances this meridional heat imbalance**

Heat transport by the atmosphere and ocean



Trenberth and Caron (2001)

- Implied heat transport from top of atmosphere (TOA) radiation balance
  - Integrate net radiation from pole to pole
- Both the atmosphere and ocean are responsible for heat transport
- Atmospheric transport is larger, particularly in the mid and high latitudes
- Oceanic heat transport is large in low-latitudes

From Wallace, J.M. and P. V. Hobbs, 2006: Atmospheric Science. Academic Press, 483pp.

Energy Transport by Atmospheric Circulation

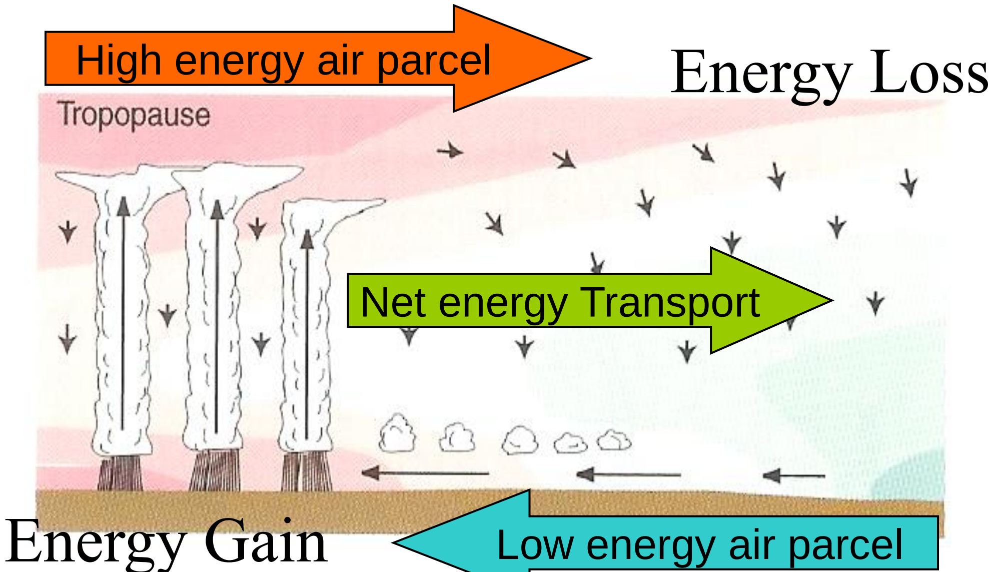


Fig. 10.3 Schematic of air parcels circulating in the atmosphere. The Colored shading represents potential temperature or moist static energy, with pink indicating higher values and blue lower values. Air parcels acquire latent and sensible heat during the time that they reside within the boundary layer, raising their moist static energy. They conserve moist static energy as they ascend rapidly in updrafts in clouds, and they cool by radiative transfer as they descend much more slowly in clear air.

- Moist Static Energy

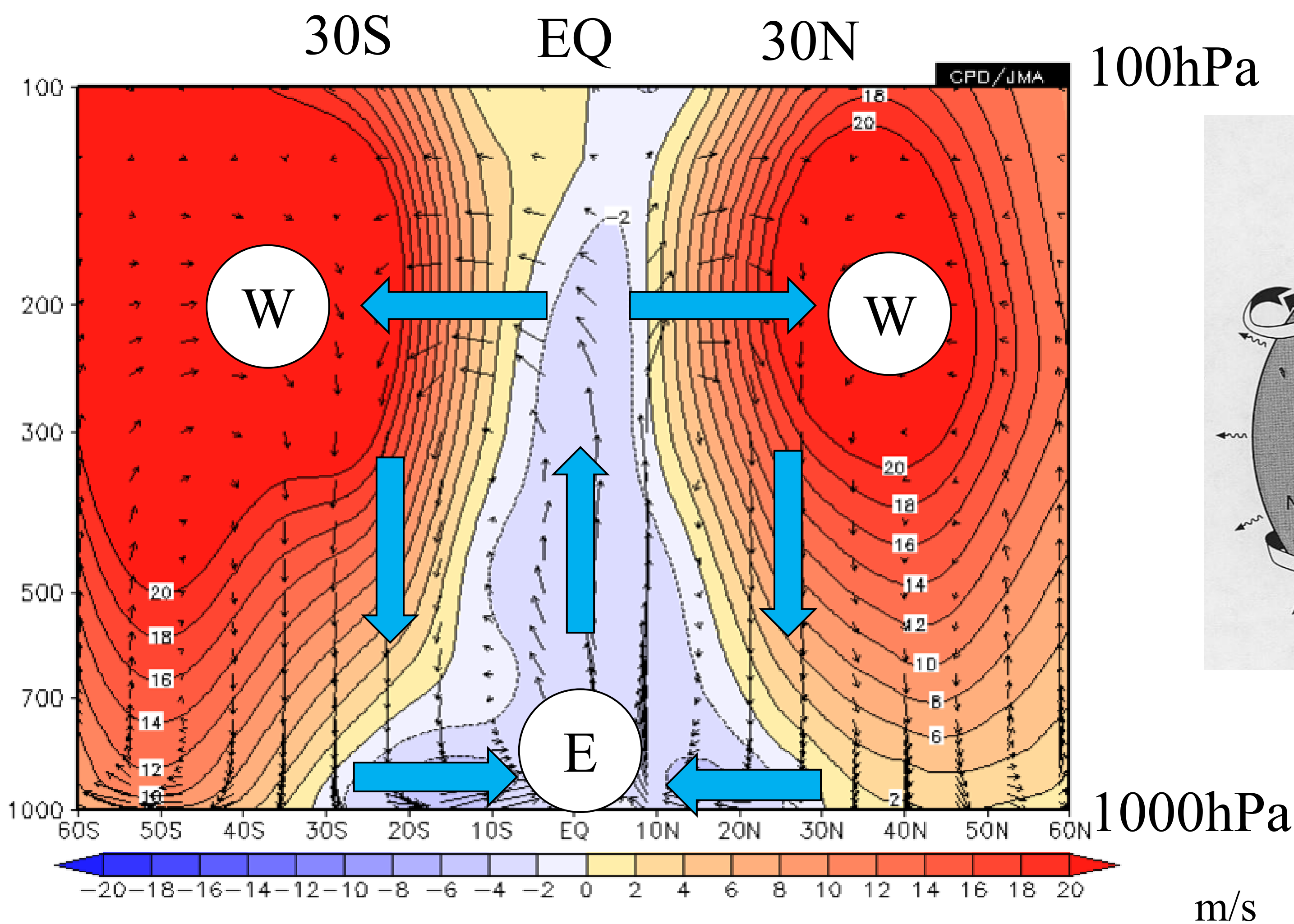
$$H=CpT+gZ+Lq$$

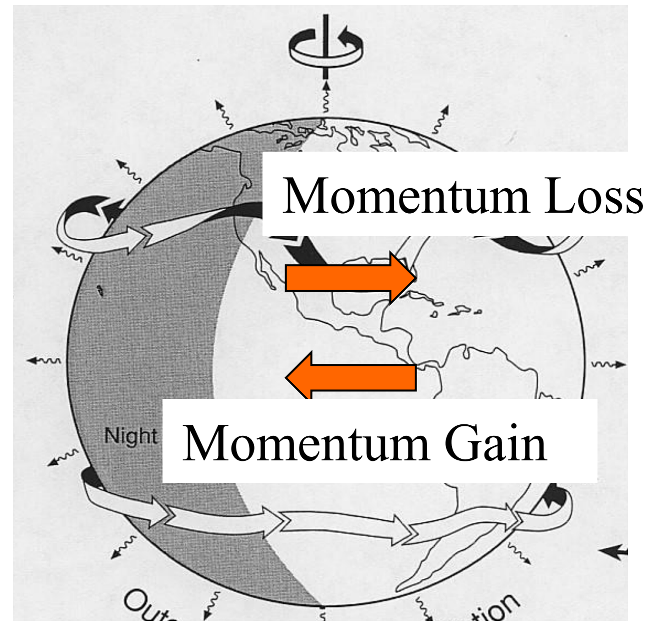
T: Temperature
Z: Height
q: Specific Humidity

H of air parcels is conserved even through adiabatic process and/or condensation process,

but, not conserved through the processes of radiation, heat and moisture supply from ground surface.

Hadley (direct) circulation and Zonal flow





Annual and zonal mean circulation. Color : Zonal wind, arrow : meridional circulation

Energy (left) and Momentum (right) Transport by Atmospheric Circulation

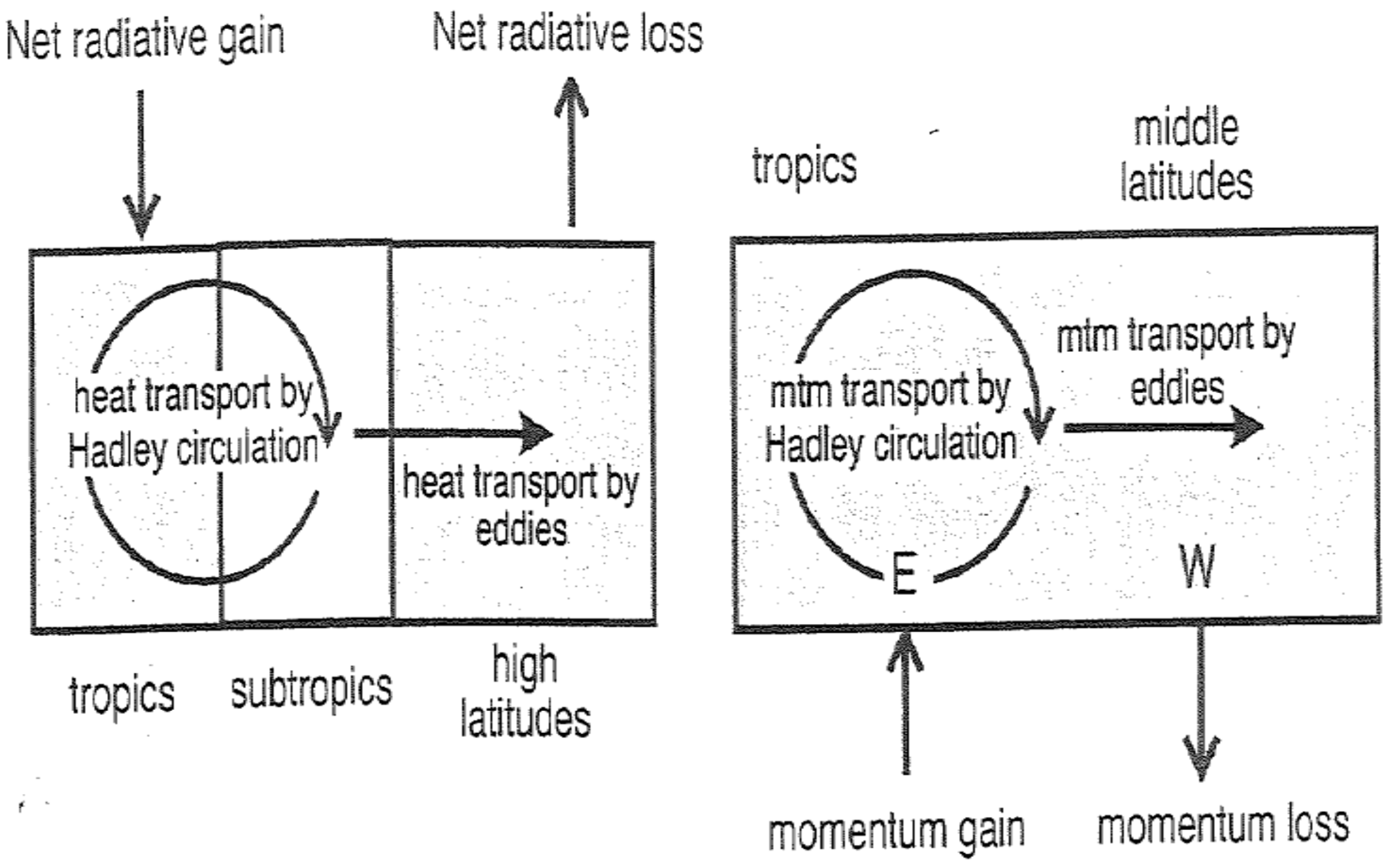


FIGURE 8.12. Schematic of the transport of (left) energy and (right) momentum by the atmospheric general circulation. Transport occurs through the agency of the Hadley circulation in the tropics, and baroclinic eddies in middle latitudes (see also Fig. 8.1).

From Marshall J., and R. A. Plumb, 2008: Atmosphere, Ocean, and Climate Dynamics, Academic Press, 319pp.

Observed atmospheric general circulation

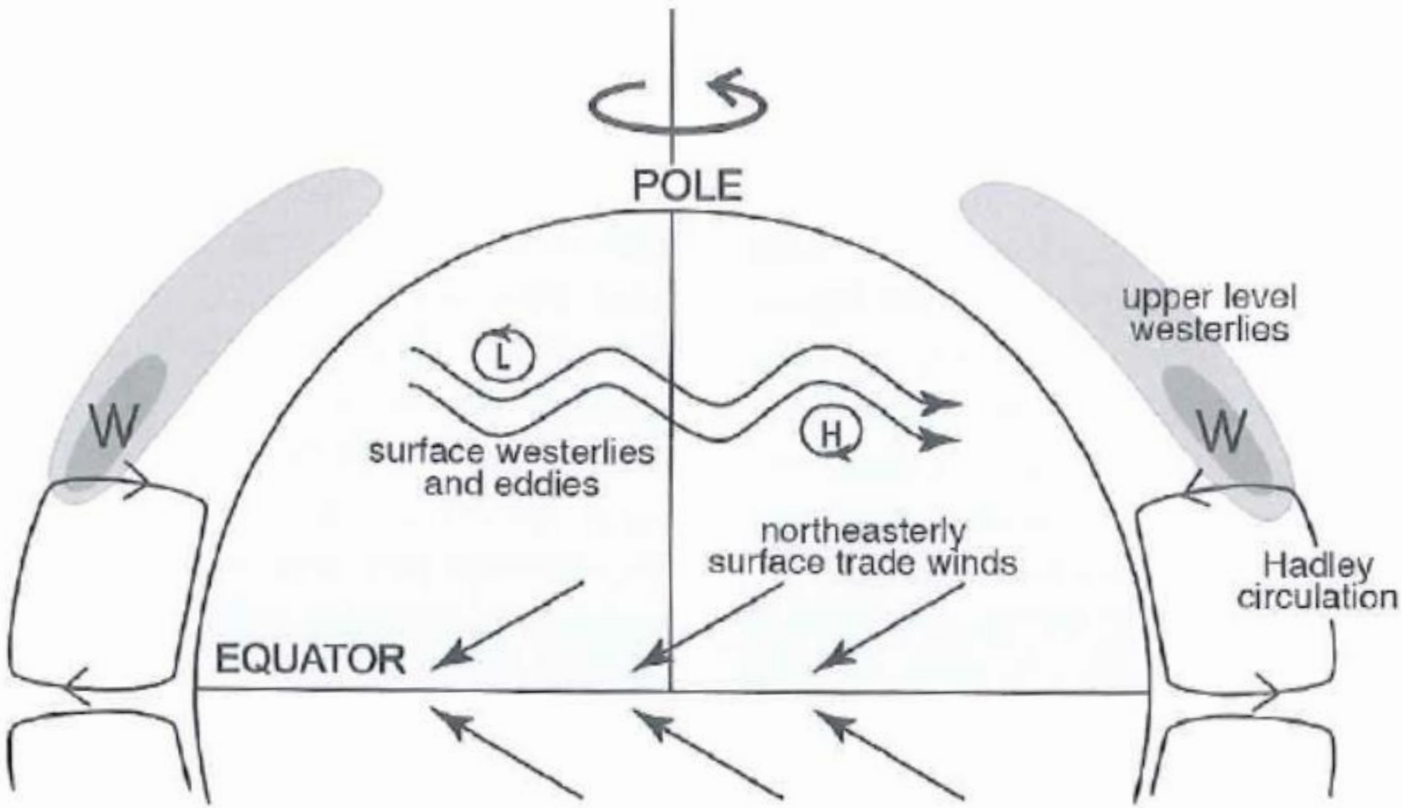


FIGURE 8.2. Schematic of the observed atmospheric general circulation for annual-averaged conditions. The upper level westerlies are shaded to reveal the core of the subtropical jet stream on the poleward flank of the Hadley circulation. The surface westerlies and surface trade winds are also marked, as are the highs and lows of middle latitudes. Only the northern hemisphere is shown. The vertical scale is greatly exaggerated.

1.4 Seasonal Change

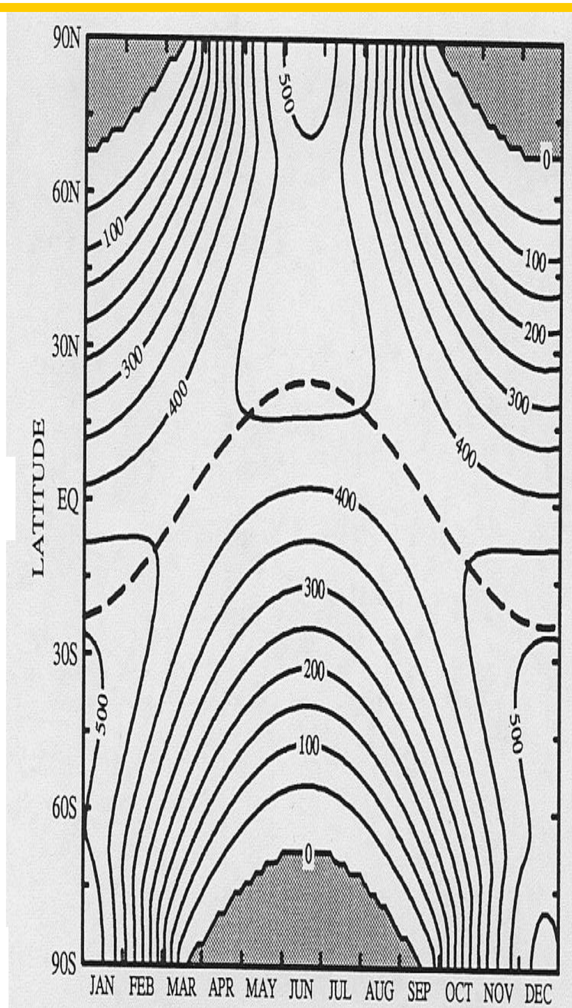
Seasonal Change of Solar Insolation and Temperature

Solar Insolation W/m²

90N

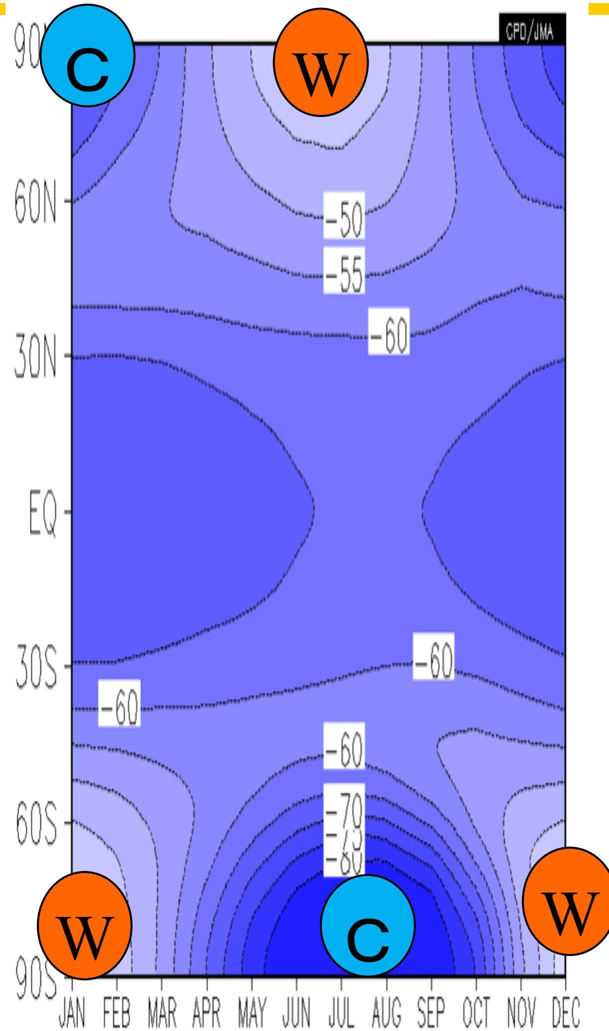
EQ.

90S



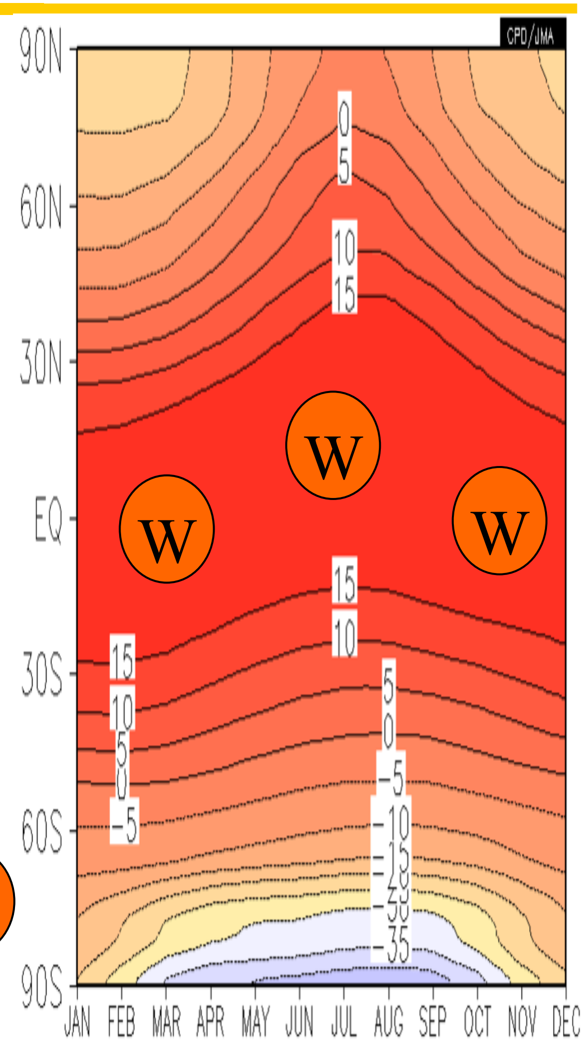
JAN JUL DEC

Temperature in the stratosphere (50hPa) ℃



JAN JUL DEC

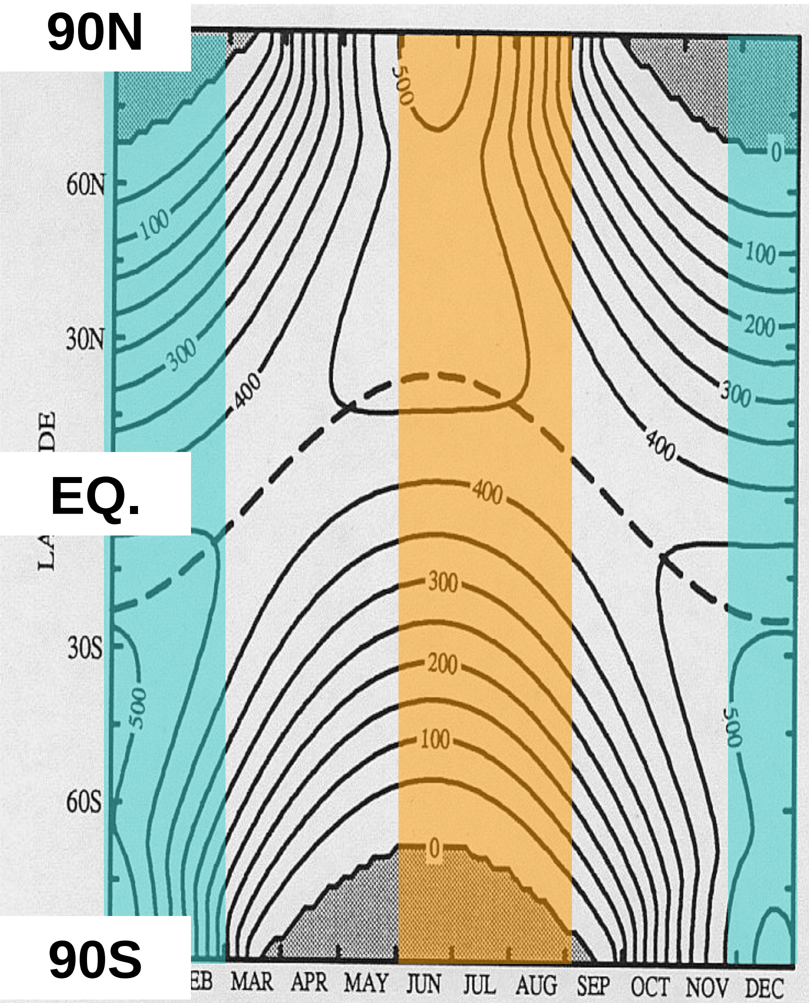
Temperature in the troposphere (850hPa) ℃



JAN JUL DEC

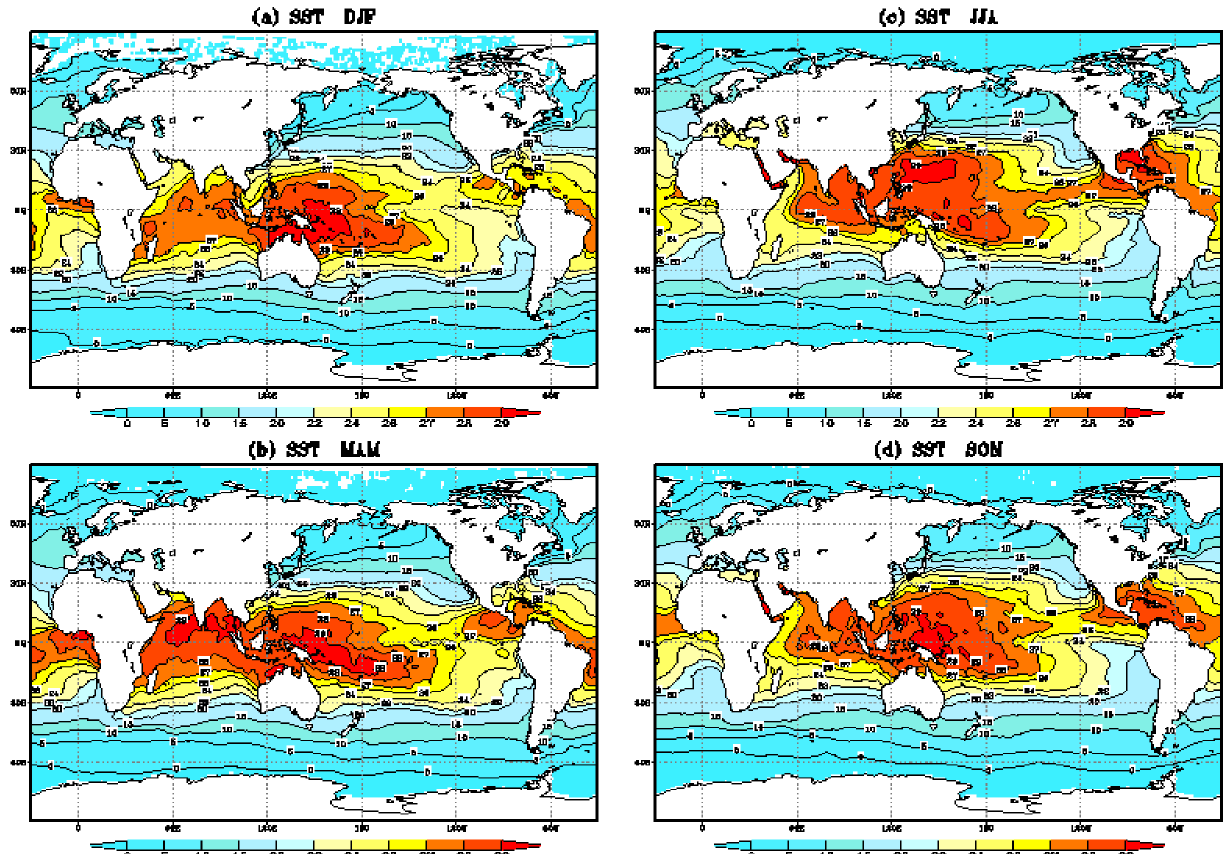
Seasonal Change of Sea Surface Temperature (SST)

Solar Insolation



Dec.-Jan.-Feb.

Jun-July-August



Mar.-Apr.-May

Sep.-Oct.-Nov.

Heat Capacity of atmosphere and ocean

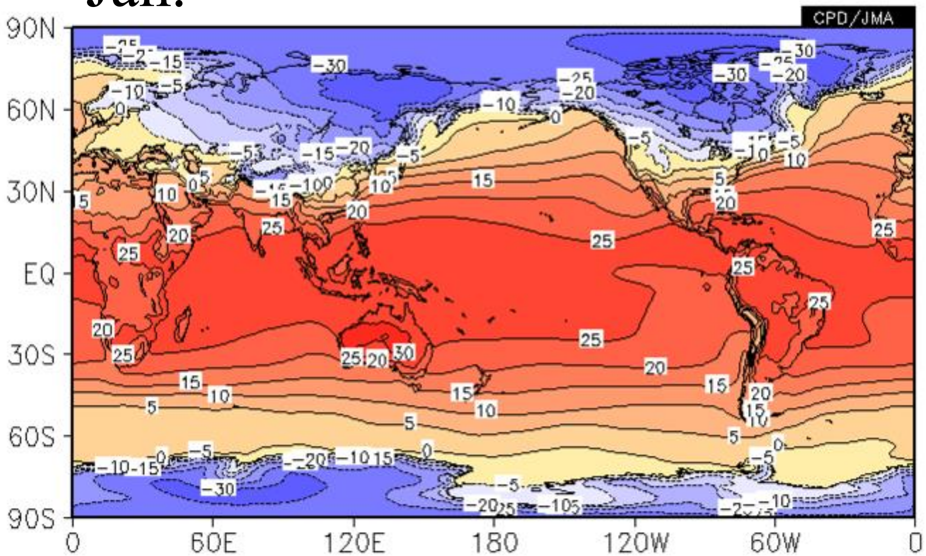
	Atmosphere	Ocean
Density	1.2-1.3kgm^{-3}	$10^3 kgm^{-3}$ ：atom. X 800
Mass(per 1 m^2)	（Top ～ Surface） $10^4 kgm^{-2}$	（Surface ～10m depth） $10^4 kgm^{-2}$ ：Mass of the atmosphere is the same as that of ocean with 10m depth
Specific heat	$10^3 Jkg^{-1}K^{-1}$	$4 \times 10^3 Jkg^{-1}K^{-1}$ ：atom. X 4
Heat capacity (per 1 m^2)	（Top ～ Surface） $10^7 JK^{-1}m^{-2}$	（Surface ～2.5m depth） $10^7 JK^{-1}m^{-2}$ ：Heat capacity of the atmosphere is the same as that of ocean with 2.5m depth

"1K in 250m depth ocean" is near equal to "100K in the atmosphere"

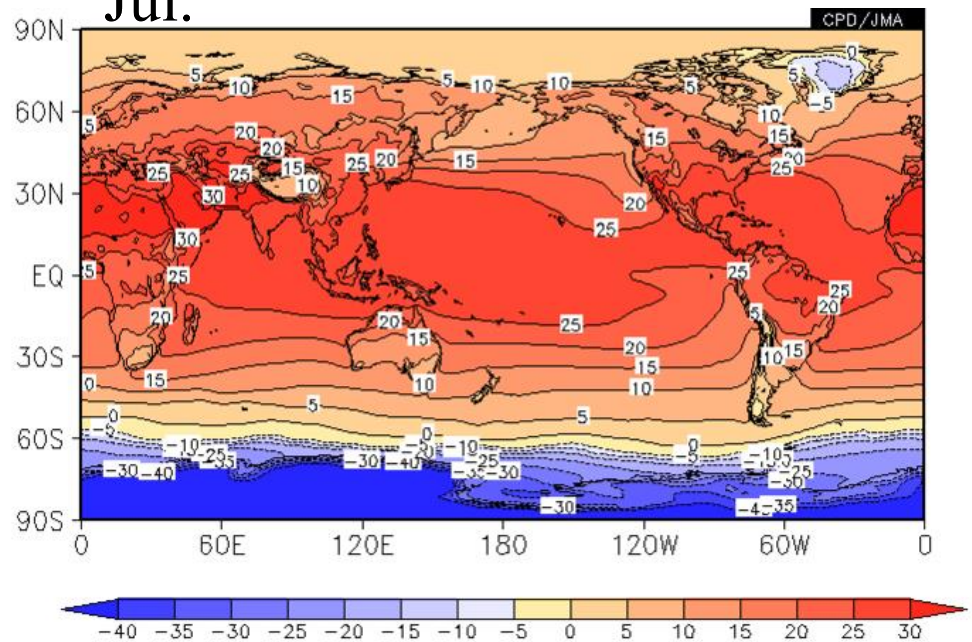
＊from Gill 1982

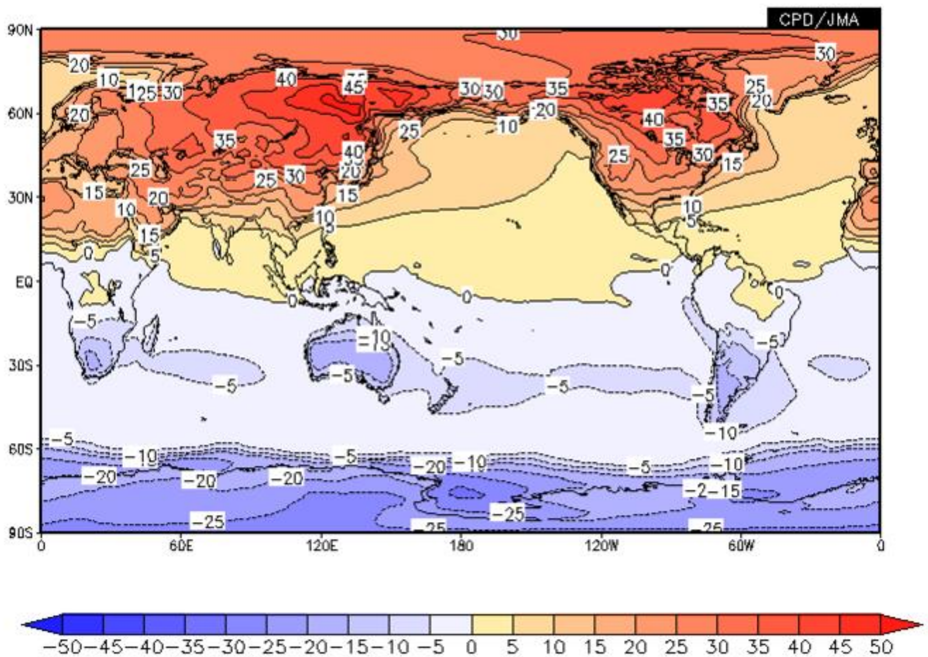
Jan-Jul contrast of surface temperature (℃)

Jan.



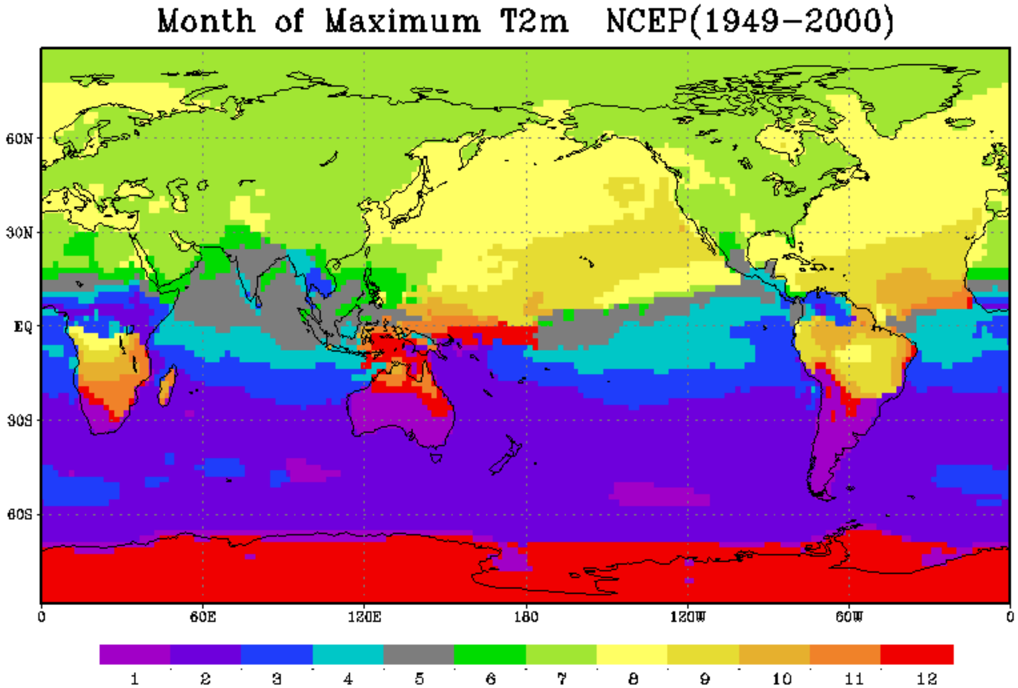
Jul.

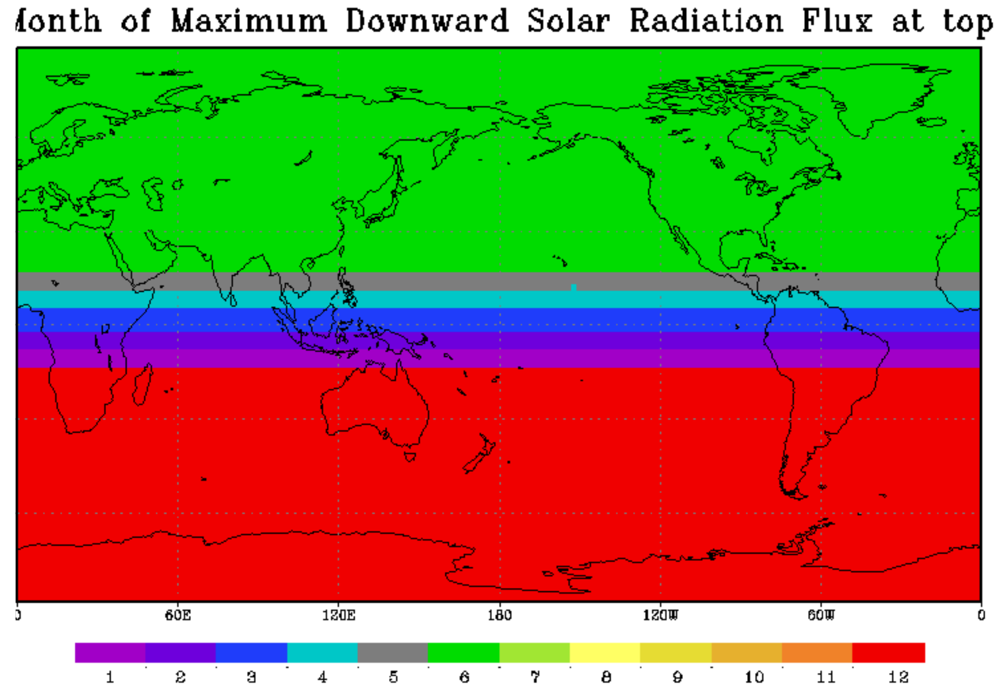




Jul.- Jan.

Month of maximum monthly mean temperature





(Right) Downward solar radiation at the top of the atmosphere is maximum in June (December) poleward of about 15° latitude in the NH (SH). In the tropics, it is January, February, March, April and May at 10° S, 4° S, 2° N, 8° N and 14° N, respectively.

(Left) Actual month of maximum monthly mean temperature is quite different due to inertia of atmosphere, land and ocean. It is July over the continents and August over the oceans in the NH, but its distribution is not simple.

From Wallace, J.M. and P. V. Hobbs, 2006: Atmospheric Science. Academic Press, 483pp.

Monsoon circulation

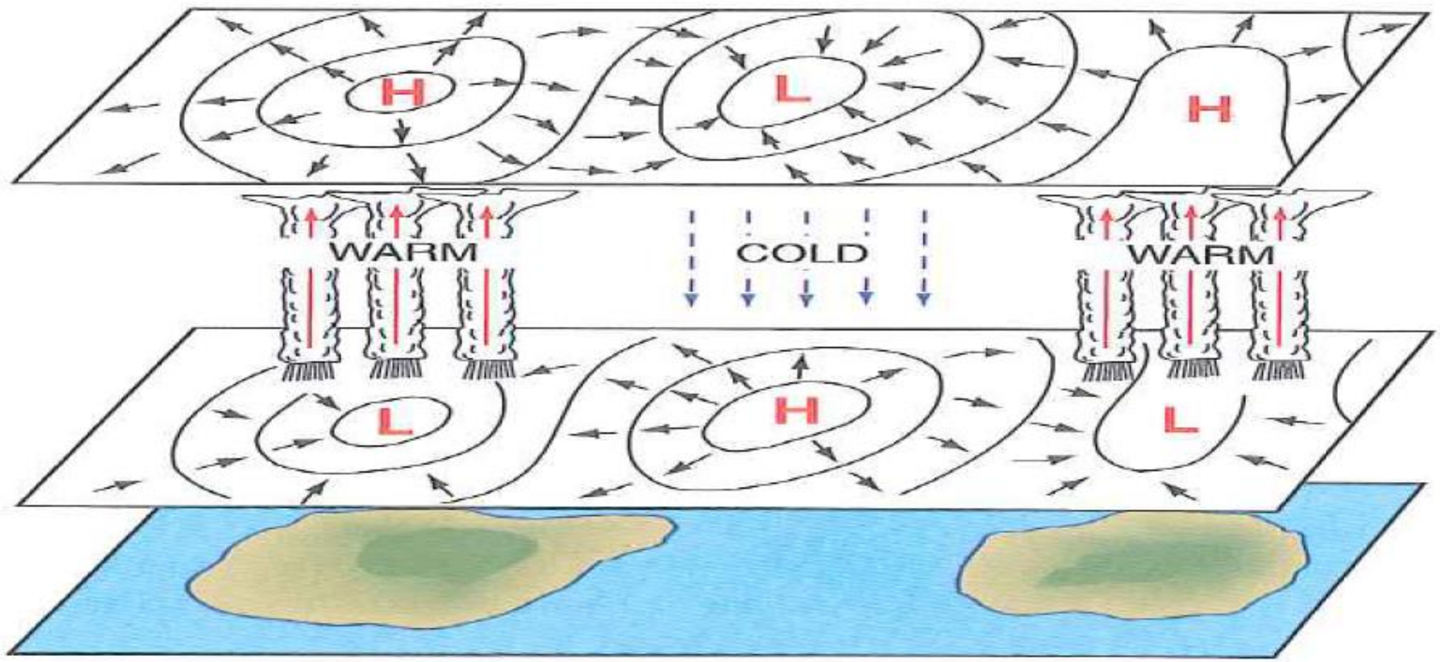
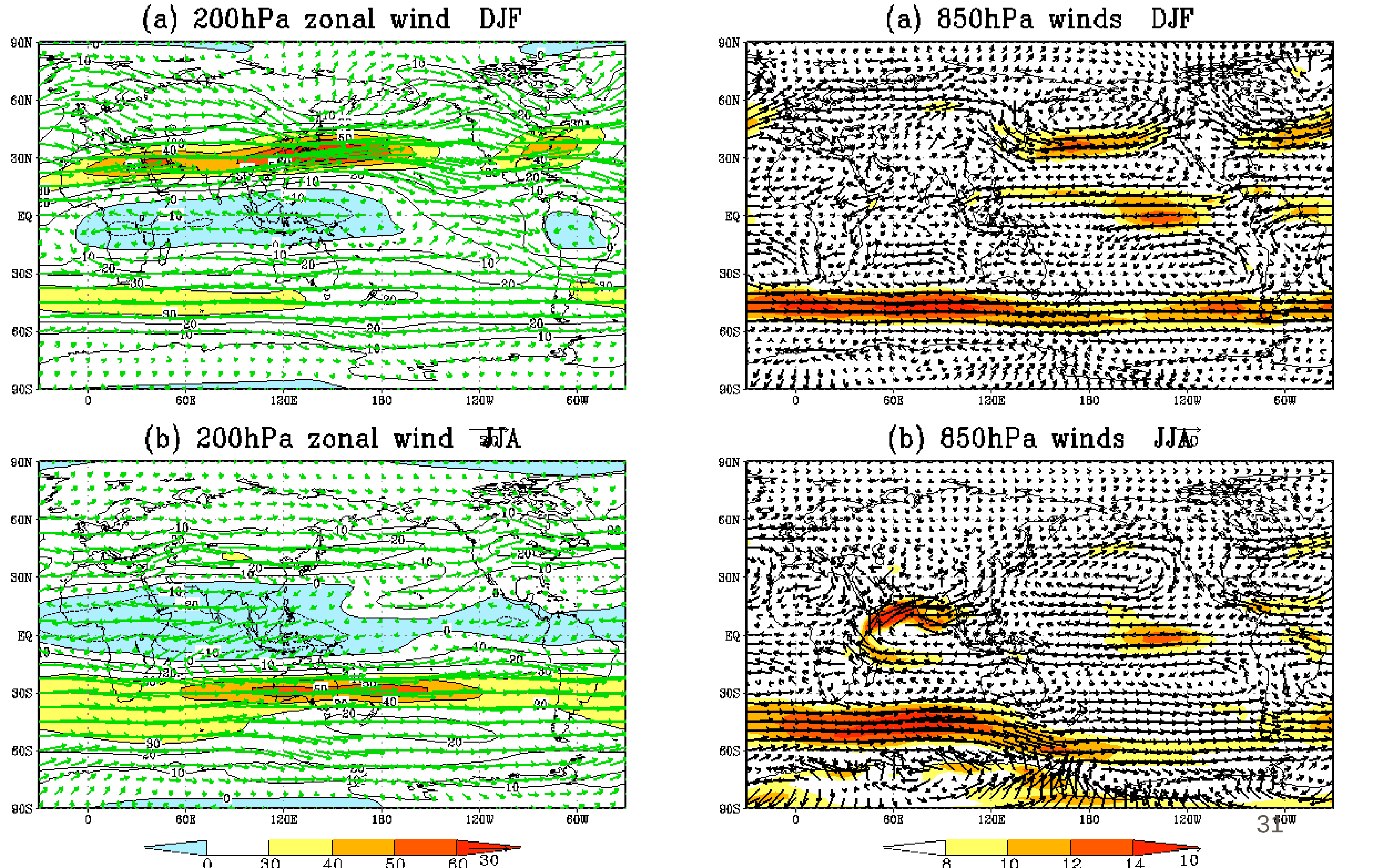
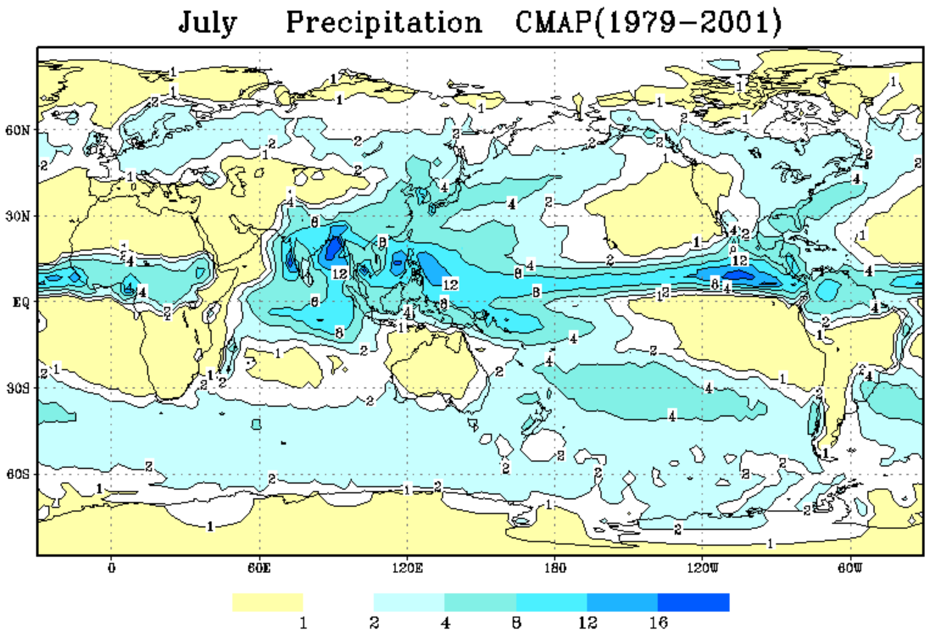


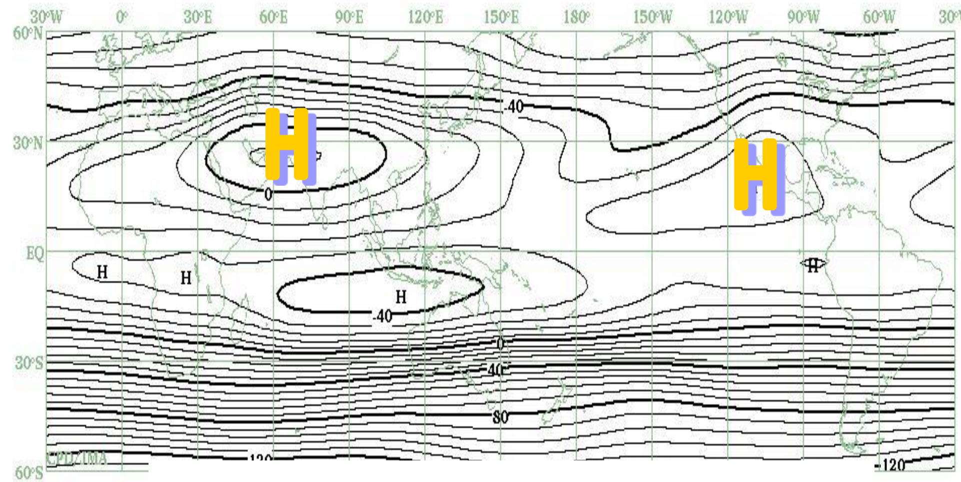
Fig. 10.9 Idealized representation of the monsoon circulations. The islands represent the subtropical continents in the summer hemisphere. Solid lines represent isobars or height contours near sea level (lower plane) and near 14 km or 150 hPa (upper plane). Short solid arrows indicate the sense of the cross-isobar flow. Vertical arrows indicate the sense of the vertical motions in the middle troposphere. Regions that experience of summer monsoon rainfall are also indicated.

200 hPa and 850 hPa winds in JJA and DJF

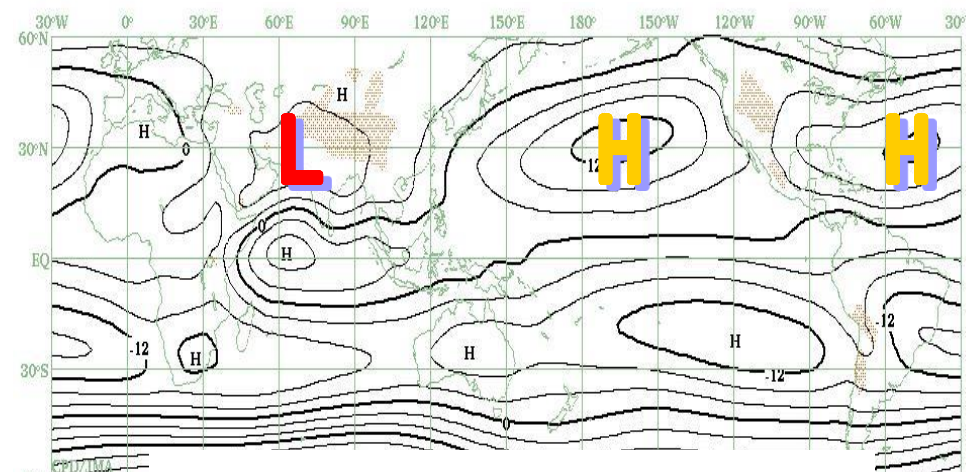


Northern Summer Monsoon



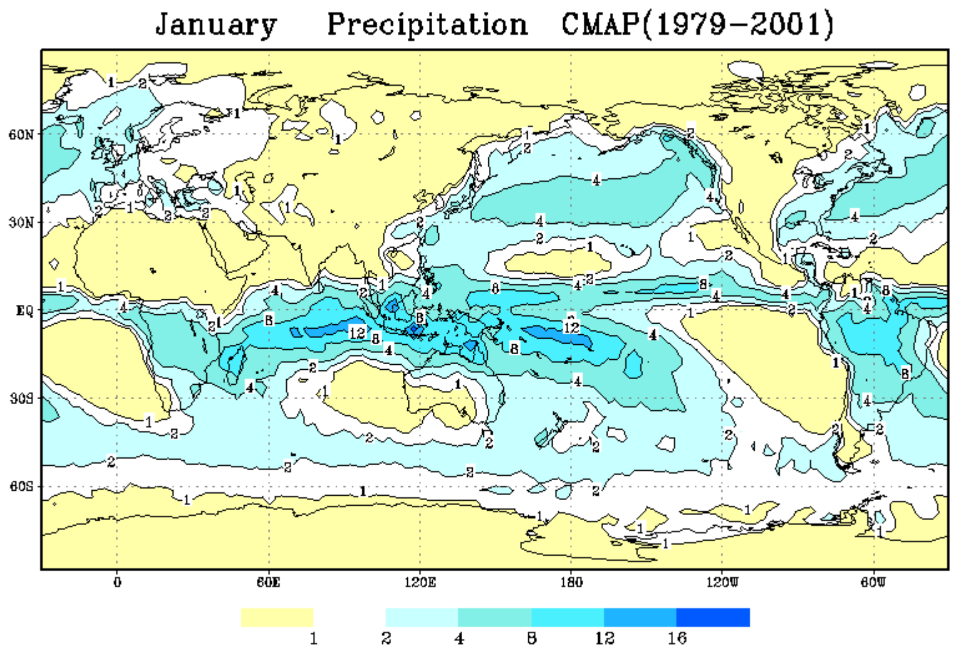


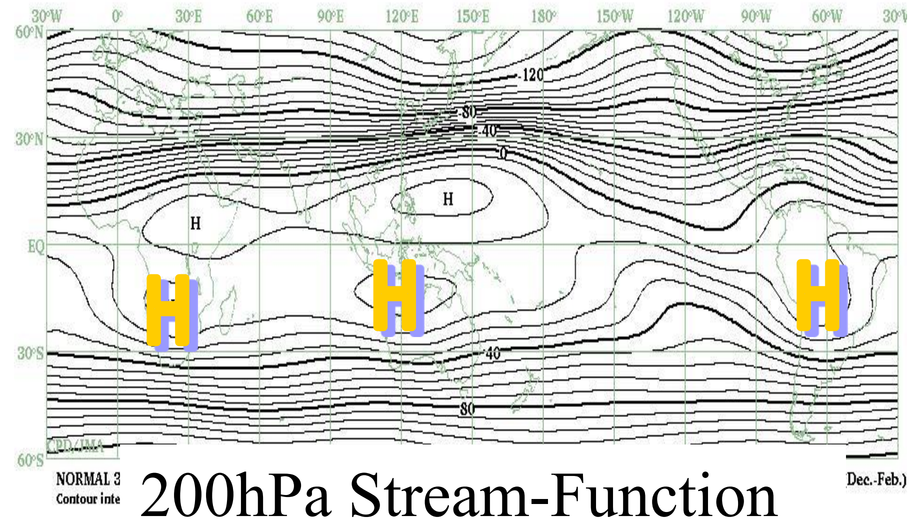
200hPa Stream-Function



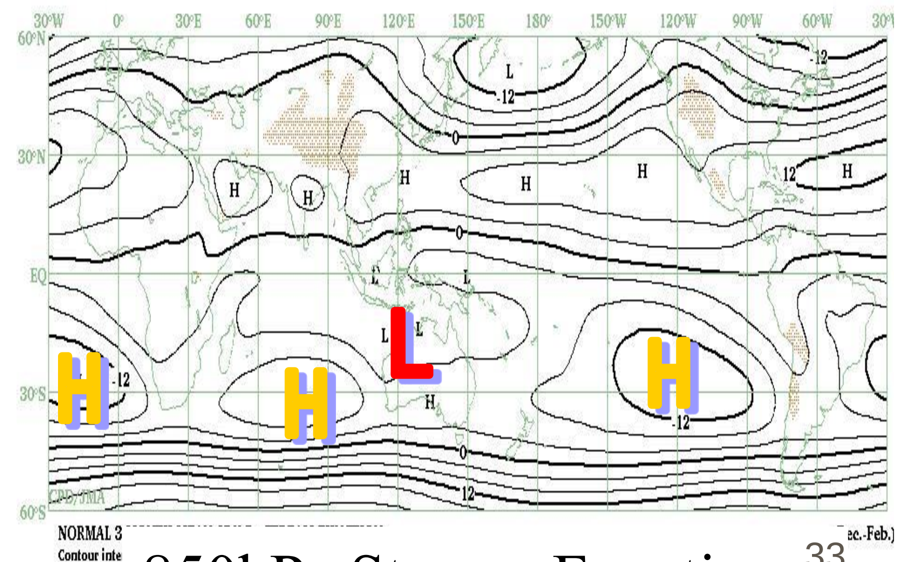
850hPa Stream-Function

Southern Summer Monsoon



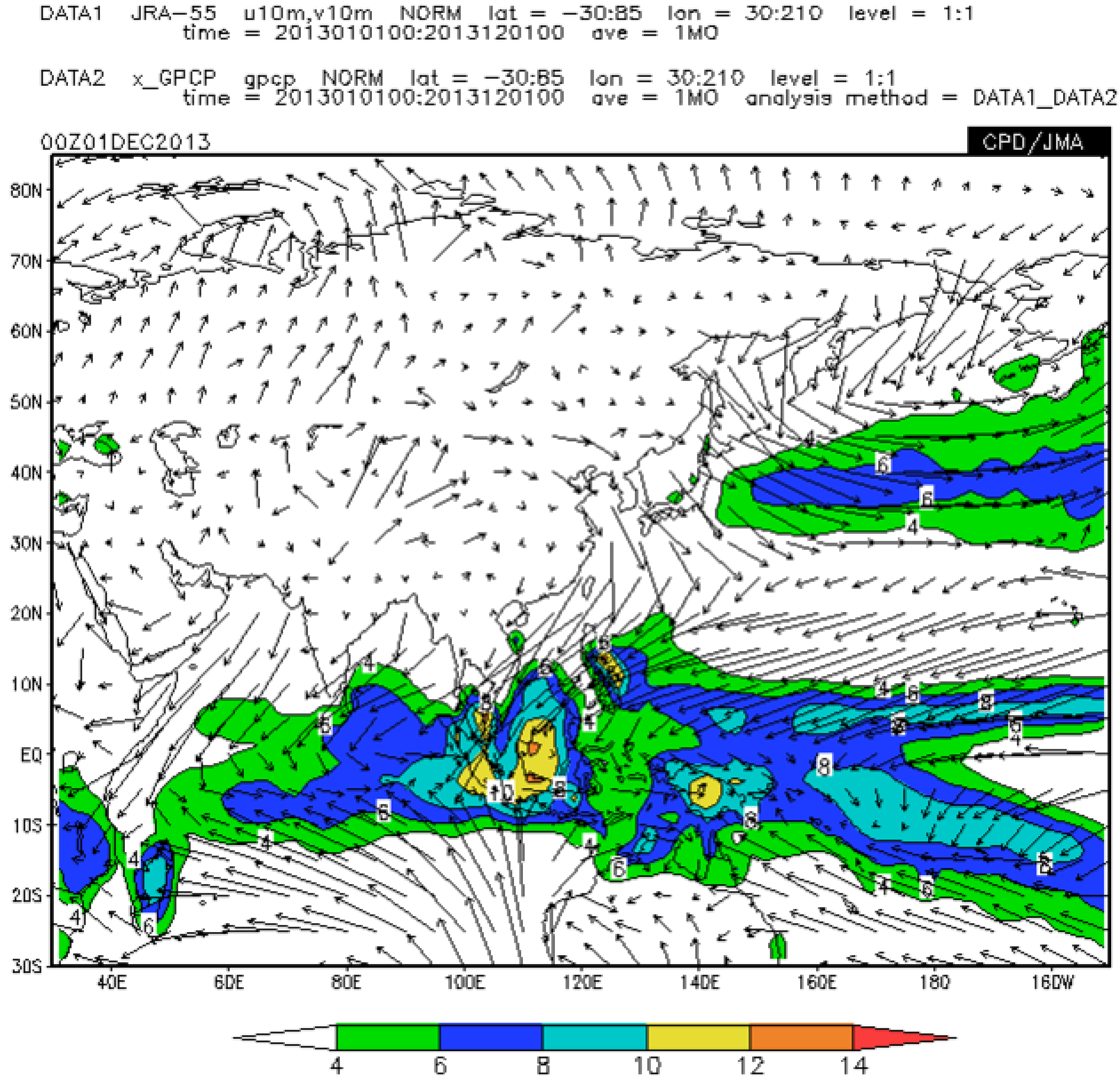


200hPa Stream-Function

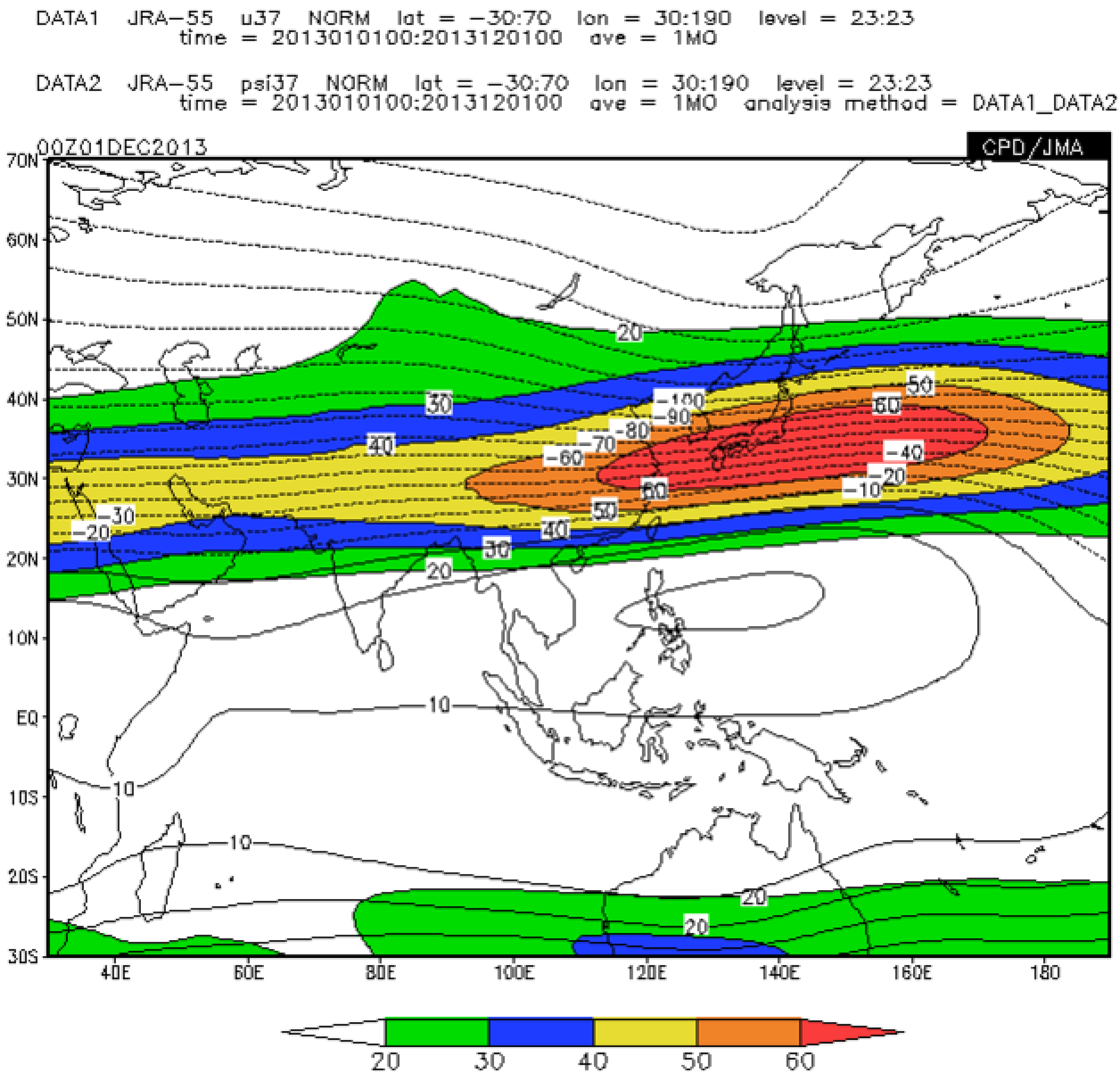


850hPa Stream-Function

Seasonal change of precipitation and surface wind



Seasonal change of zonal wind and stream function at 200hPa



Q. J. R. Meteorol. **SOC. (1996), 122,** pp. **1385-1404** Monsoons and the dynamics of deserts. By MARK J. RODWELL' and BRIAN **I.** HOSKINS

25N

Precipitation

10N

Precipitation

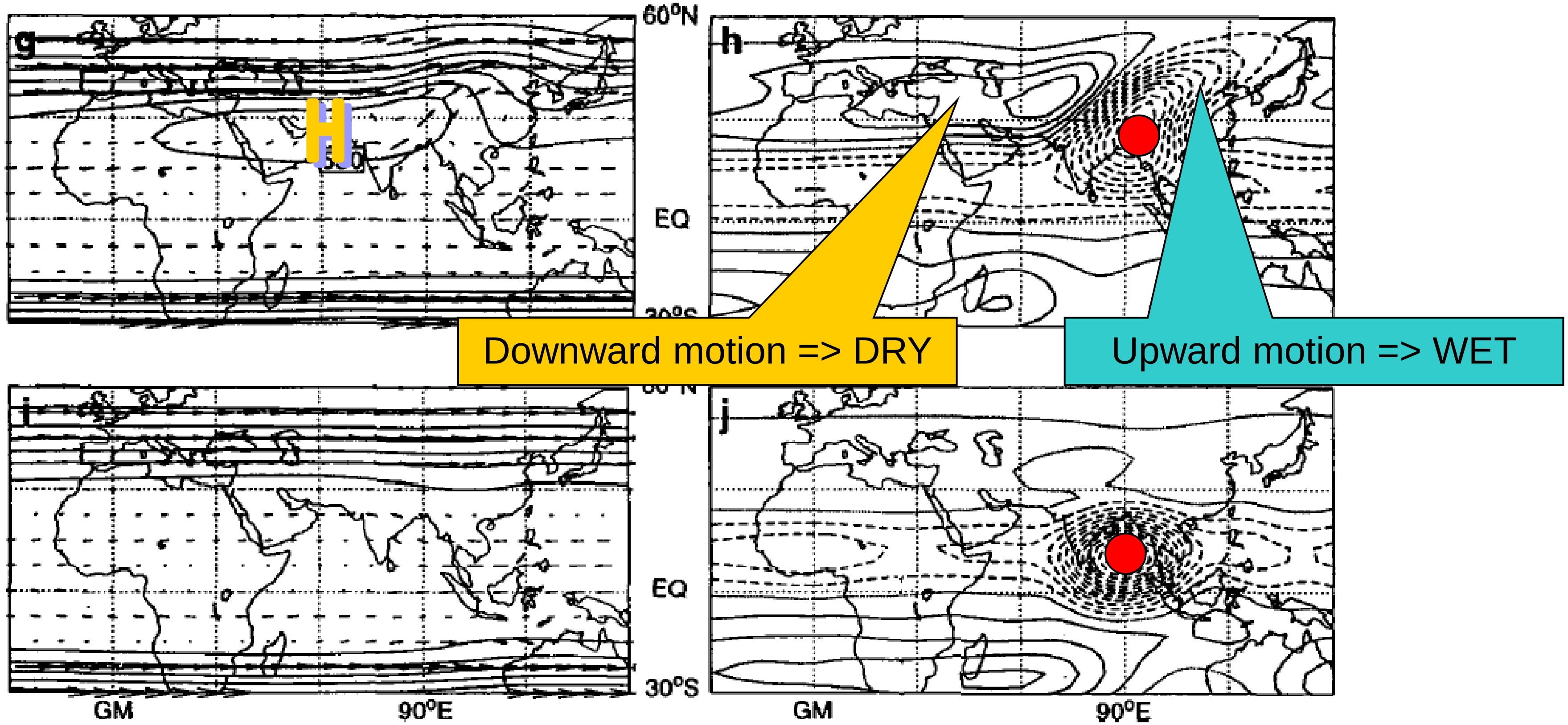


Figure 8. Day 11 of a series of integrations without orography. (a), (c), (e), (g) and (i) pressure and horizontal winds on the 325 K isentropic surface, with contour interval 40 hPa; (b), (d), (f), (h) and (j) vertical velocity at 477 hPa, with contour interval 0.25 hPa hr^{-1}; (a) and (b) integration linearized about a resting basic-state and forced with heating at 90°E, 25°N; (c) and (d) integration linearized about a resting basic-state and forced with heating at 25°N superimposed on the June to August zonal-mean flow; (e) and (f) integration linearized about the zonal-mean basic-state and forced with heating at 25°N; (g) and (h) non-linear integration forced with heating at 25°N; (i) and (j) nonlinear integration forced with heating at 90°E, 10°N.

1.5 Role of orography on climate

Effect of mountains on climate

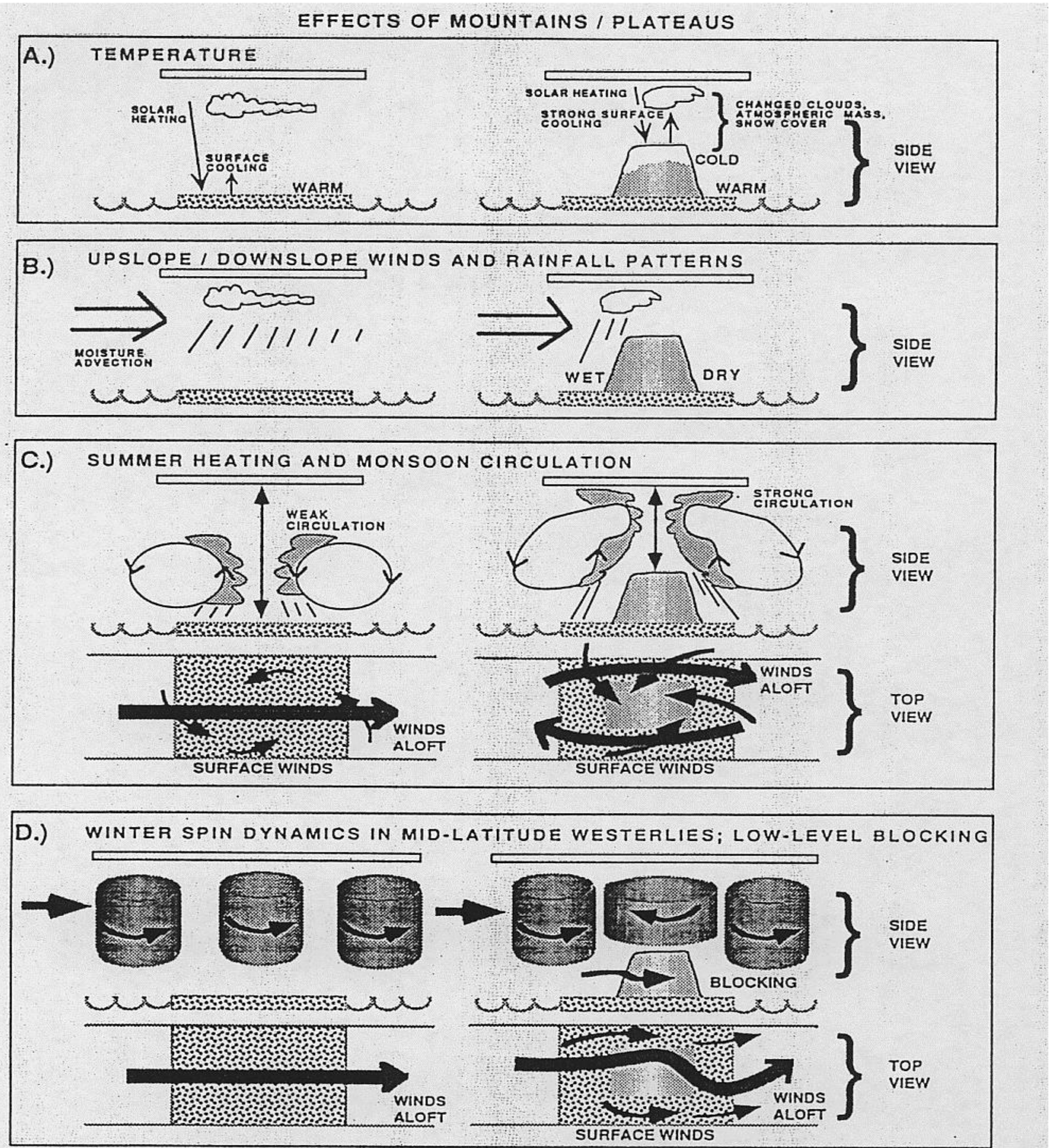
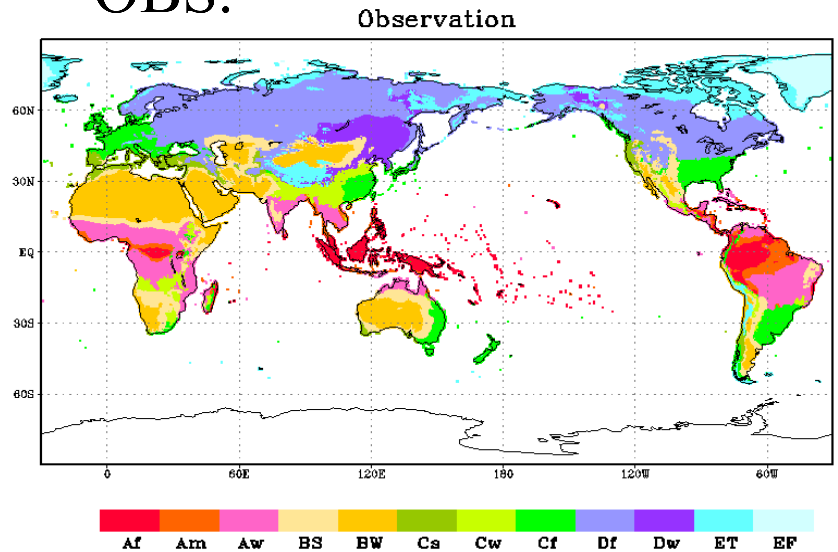


Figure 1. Effects of mountains/plateaus on climate: (*a*) temperature, (*b*) upslope/downslope winds and rainfall patterns, (*c*) summer heating and monsoon circulation, and (*d*) winter spin dynamics in mid-latitude westerlies, and low-level blocking. See text for explanation.

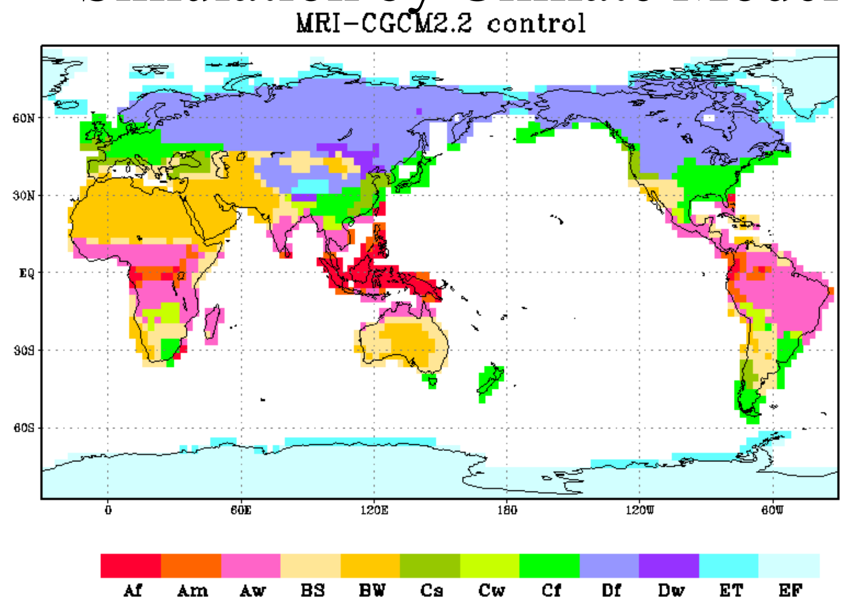
Kutzbach et al. (1993) J.Geology

Effect of mountain: Koppen climate

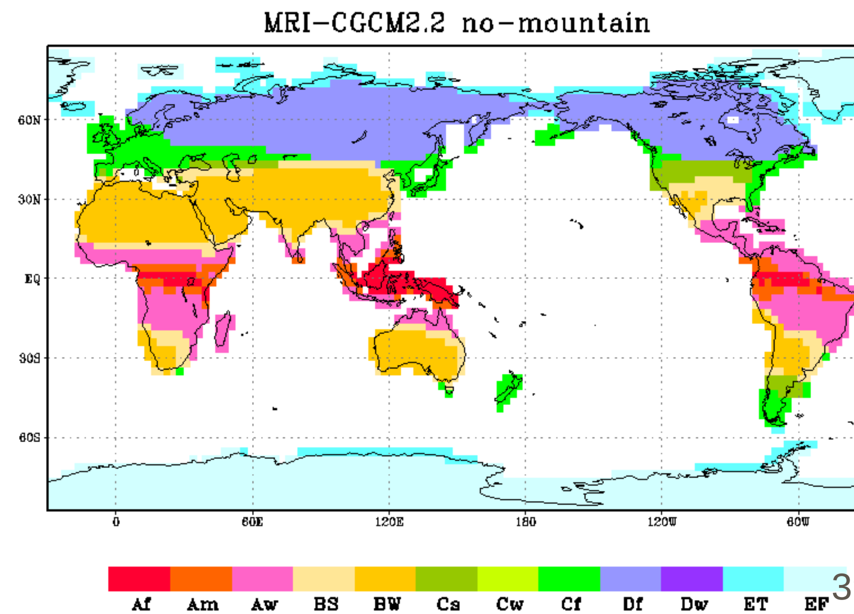
OBS.



Simulation by Climate Model



Simulation by Climate Model with mountain



Outline of the lecture

1. Climate System (90 min. + α)
   1.1 Introduction
   1.2 Radiative Balance
   1.3 Horizontal Radiative Imbalance and Circulations
   1.4 Seasonal Change
   1.5 Role of Orography on Climate
2. Climate Variability (120 min. + α)
   2.1 Introduction
   2.2 Intraseasonal Variability: Quasi-stationary Rossby wave, MJO and equatorial waves
   2.3 Interannual Variability: ENSO, El Nino Modoki, IOD
   2.4 Decadal Variability: PDO, ENSO-Monsoon relation
3. JMA's latest one-month prediction