Introduction to Climate

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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.

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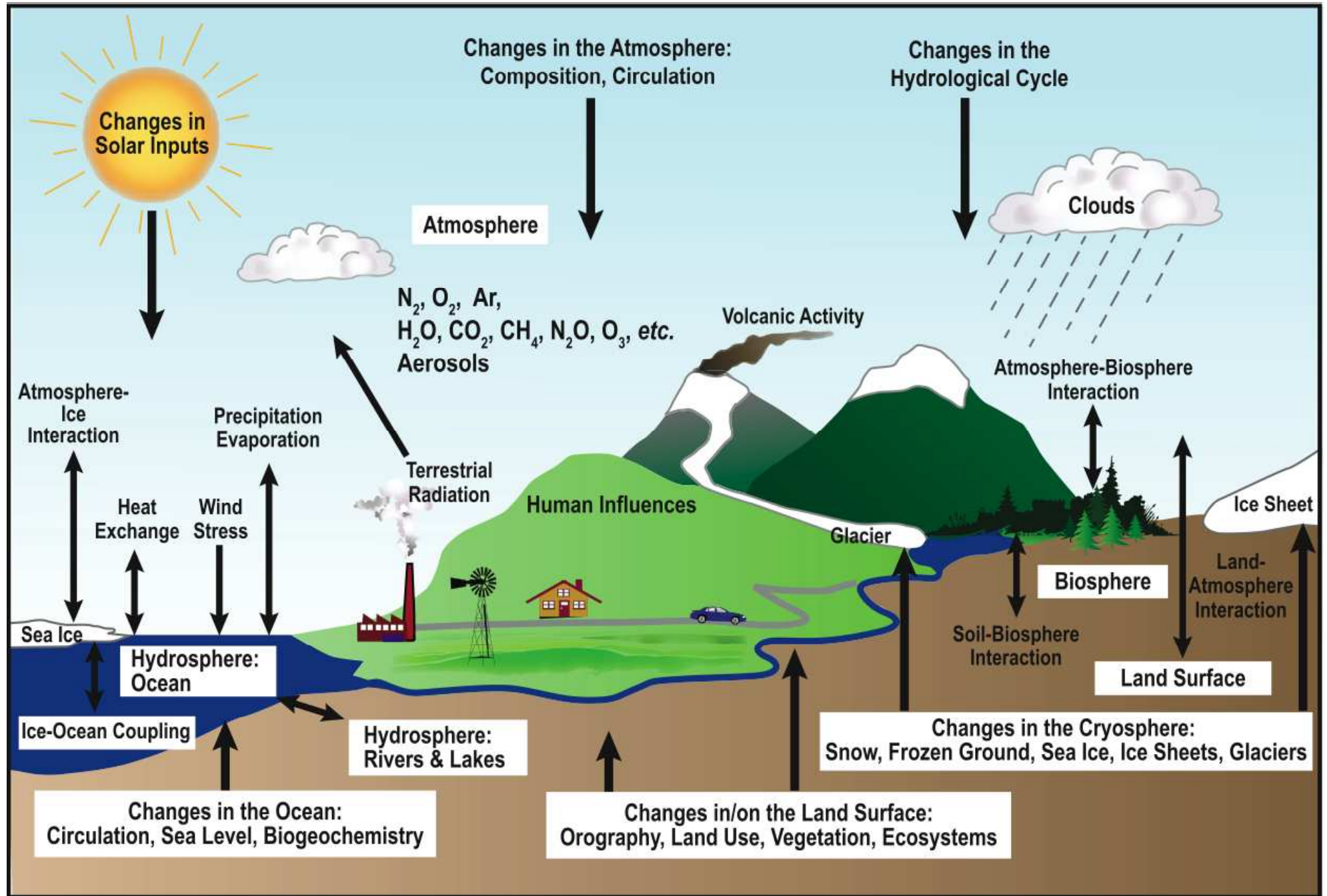
Schematic view of the components of the climate system, their processes and interactions.

Radiative Balance

Radiative Balance between Earth and Space

■ Solar Radiation

- ■ **Solar constant :** $S \approx 1370\, Wm^{-2}$

- ■ **insolation :** $\dfrac{S}{4} \approx 342 Wm^{-2}$

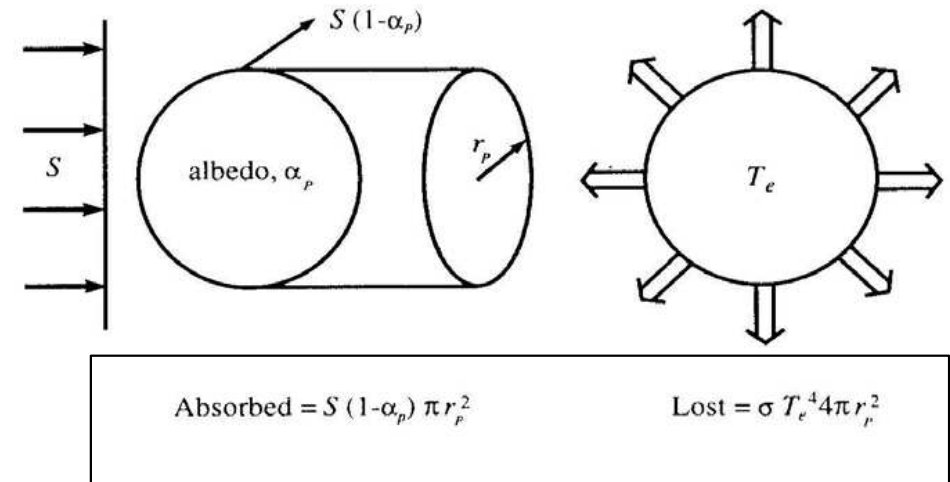- ■ **planetary albedo :** $\alpha_P \approx 0.31$

- ■ **absorption :** $\dfrac{S}{4}(1 - \alpha_P) \approx 235 Wm^{-2}$

■ Terrestrial Radiation

- ■ **equilibrium radiative temperature :** T_e^*

- ■ **emission :** σT_e^{*4}

Earth's temperature observed from the cosmic space

Absorbed $= S(1-\alpha_p)\, \pi r_P^2$ Lost $= \sigma T_e^{4} 4\pi r_P^2$

■ Radiation Balance

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

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

Surface Temperature = 15° C

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
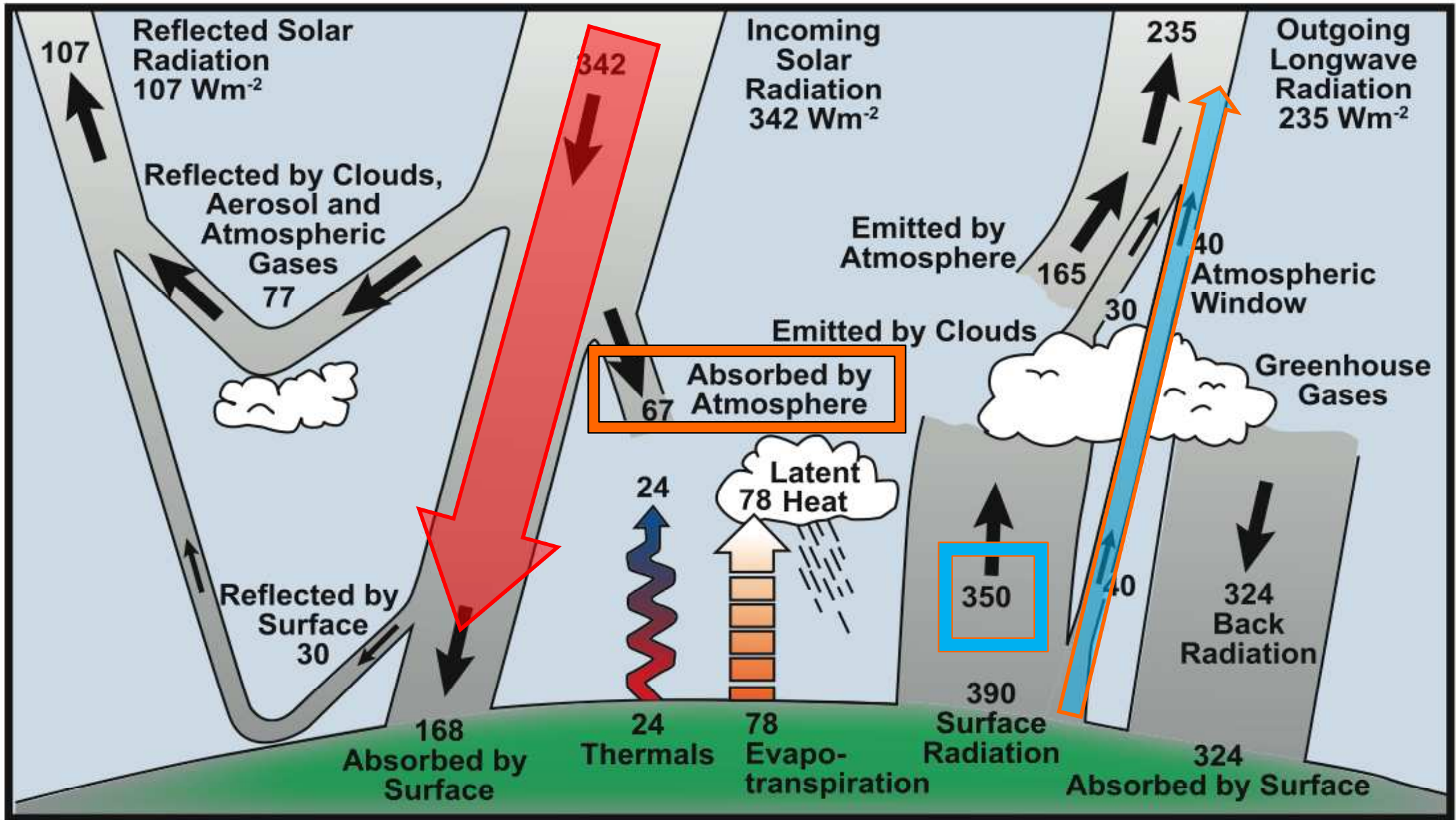
Surface Temperature

-55° C

Surface Pressure

10hPa
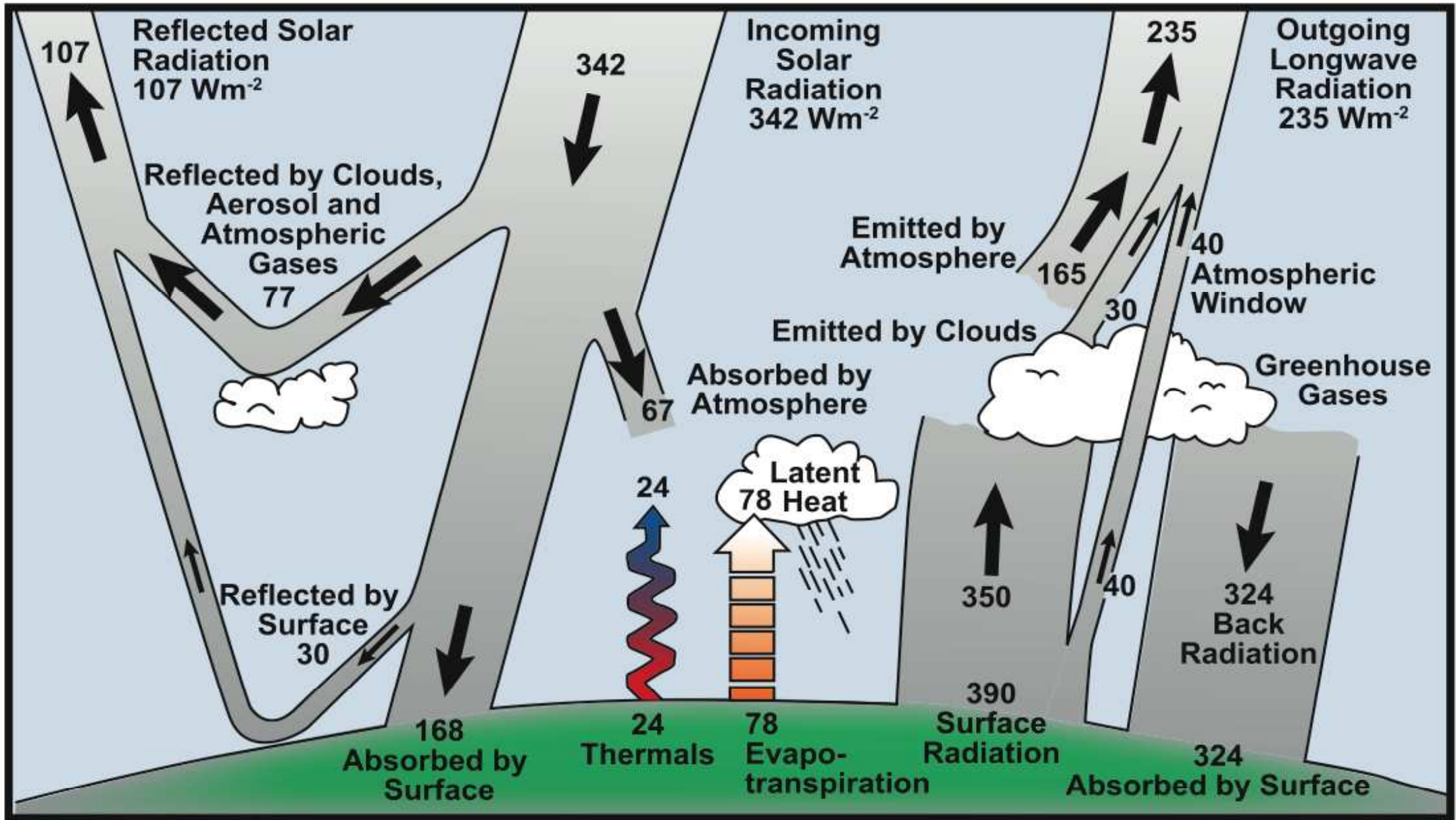
Estimate of the Earth's annual and global mean energy balance. Over the long term, the amount of incoming solar radiation absorbed by the Earth and atmosphere is balanced by the Earth and atmosphere releasing the same amount of outgoing longwave radiation. About half of the incoming solar radiation is absorbed by the Earth's surface. This energy is transferred to the atmosphere by warming the air in contact with the surface (thermals), by evapotranspiration and by longwave radiation that is absorbed by clouds and greenhouse gases. The atmosphere in turn radiates longwave energy back to Earth as well as out to space. Source: Kiehl and Trenberth (1997).

Estimate of the Earth's annual and global mean energy balance. Over the long term, the amount of incoming solar radiation absorbed by the Earth and atmosphere is balanced by the Earth and atmosphere releasing the same amount of outgoing longwave radiation. About half of the incoming solar radiation is absorbed by the Earth's surface. This energy is transferred to the atmosphere by warming the air in contact with the surface (thermals), by evapotranspiration and by longwave radiation that is absorbed by clouds and greenhouse gases. The atmosphere in turn radiates longwave energy back to Earth as well as out to space. Source: Kiehl and Trenberth (1997).

Absorption of Radiation from 6000K and 255K Blackbodies

Solar radiation

Terrestrial radiation
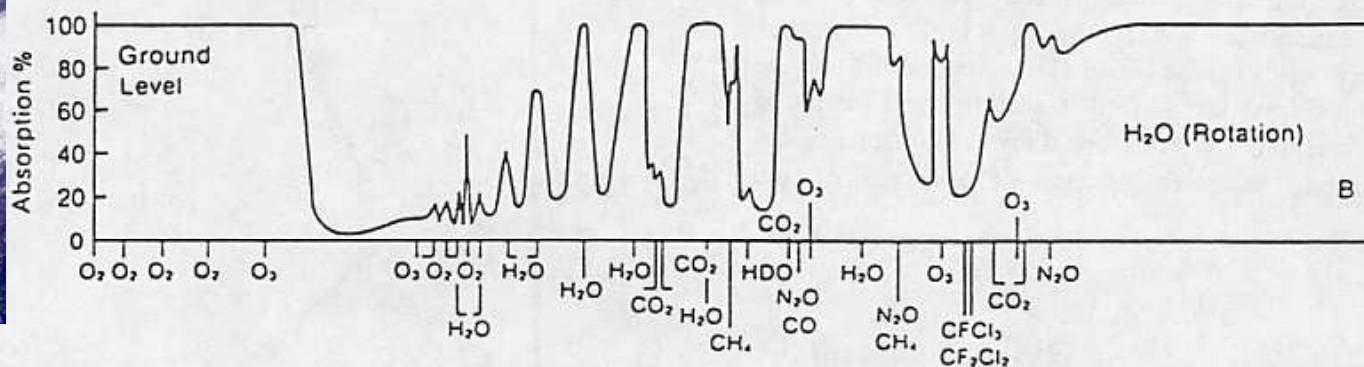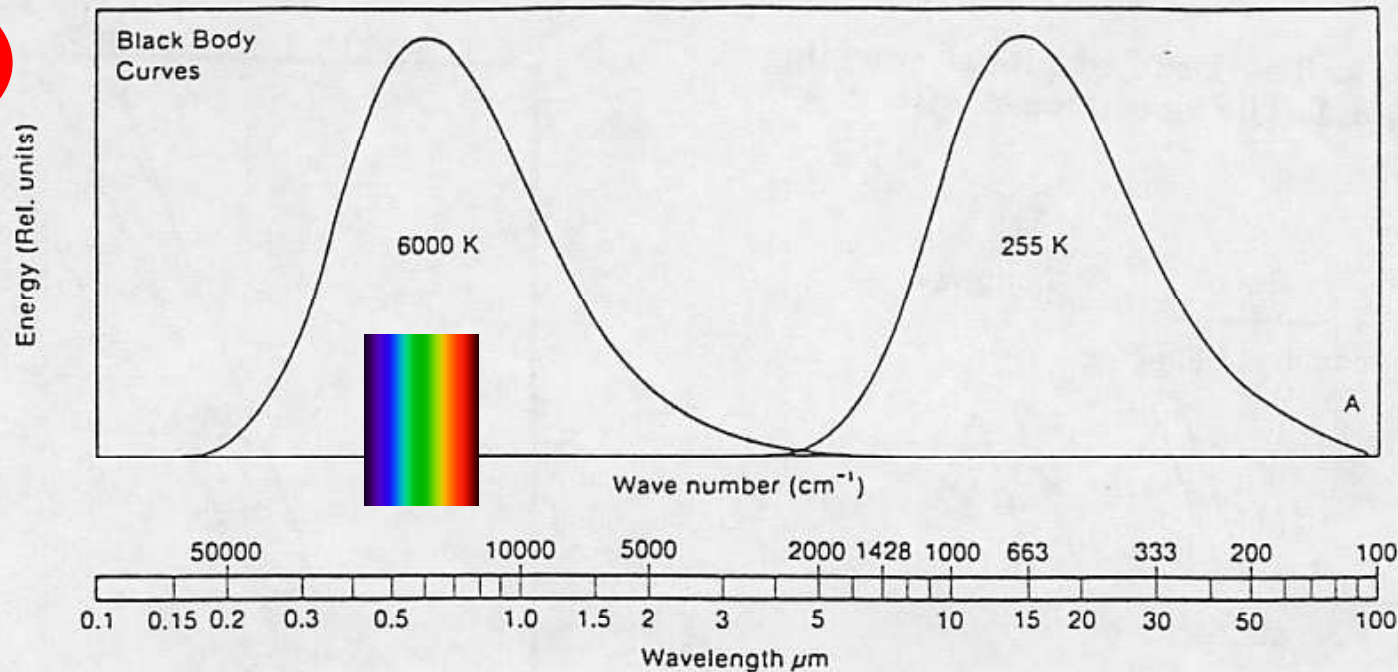
(a) Spectral distribution of long-wave emission from blackbodies at 6000 K and 255 K, corresponding to the mean emitting temperatures of the Sun and Earth, respectively, and (b) percentage of atmospheric absorption for radiation passing from the top of the atmosphere to the surface. Notice the comparatively 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].
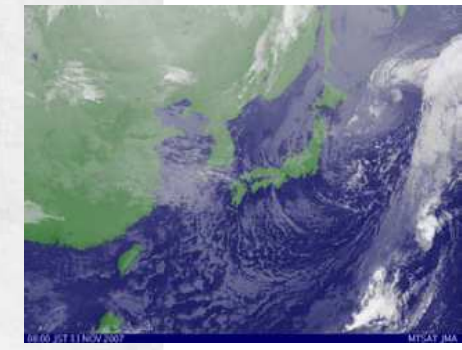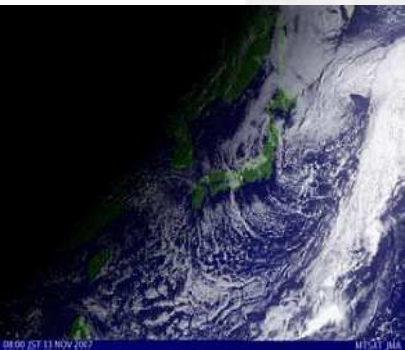
Radiative heating tends to create vertical instability
between heated ground and cooled atmosphere on average

Reflected Solar Radiation 107 Wm^{-2}

107

342

Incoming Solar Radiation 342 Wm^{-2}

235

Outgoing Longwave Radiation 235 Wm^{-2}

Reflected by Clouds, Aerosol and Atmospheric Gases 77

Emitted by Atmosphere

Atmospheric Cooling

40 Atmospheric Window

Greenhouse Gases

67 Abs Atm

Latent 78 Heat

24

Unstable

350

40

324 Back Radiation

Reflected by Surface 30

24 Thermals

78 Evapo-transpiration

390 Surface Rad

Ground Surface Heating

Greenhouse Effect

Estimate of the Earth's annual and global mean energy balance. Over the long term, the amount of incoming solar radiation absorbed by the Earth and atmosphere is balanced by the Earth and atmosphere releasing the same amount of outgoing longwave radiation. About half of the incoming solar radiation is absorbed by the Earth's surface. This energy is transferred to the atmosphere by warming the air in contact with the surface (thermals), by evapotranspiration and by longwave radiation that is absorbed by clouds and greenhouse gases. The atmosphere in turn radiates longwave energy back to Earth as well as out to space. Source: Kiehl and Trenberth (1997).

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)

Typical condition at 35N in April

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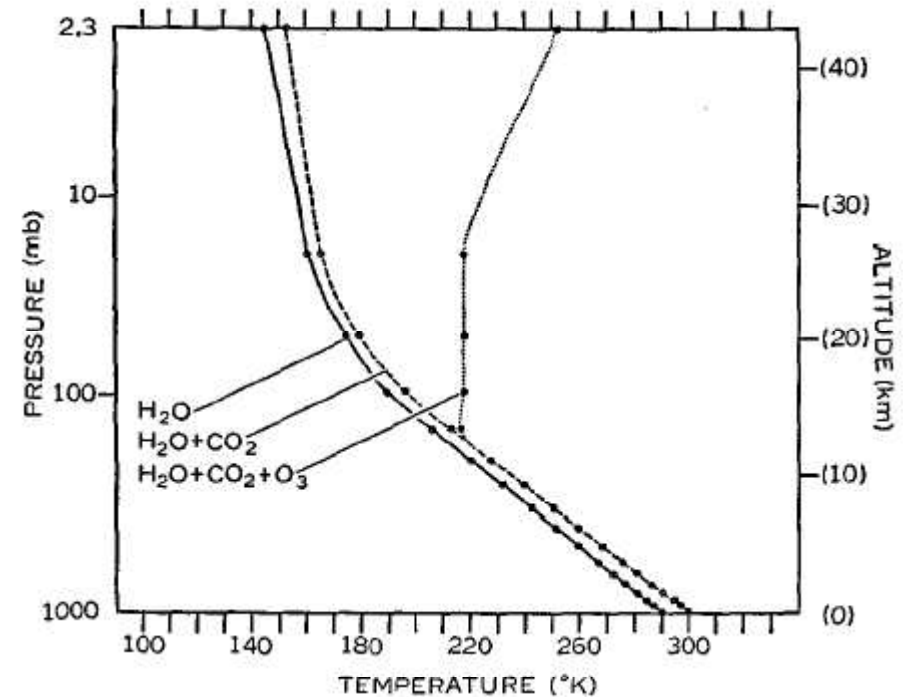
FIG. 6c. Thermal equilibrium of various atmospheres which have a critical lapse rate of 6.5 deg km^{-1}. Vertical distributions of gaseous absorbers at 35N, April, were used. $S_c = 2$ ly min^{-1}, $\cos\bar{\zeta} = 0.5$, $r = 0.5$, no clouds.

One-Dimensional Vertical Profile Model

Thermal Equilibrium of the Atmosphere with a Convective Adjustment

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Observed Temperature

1-D model Simulations for each latitudes
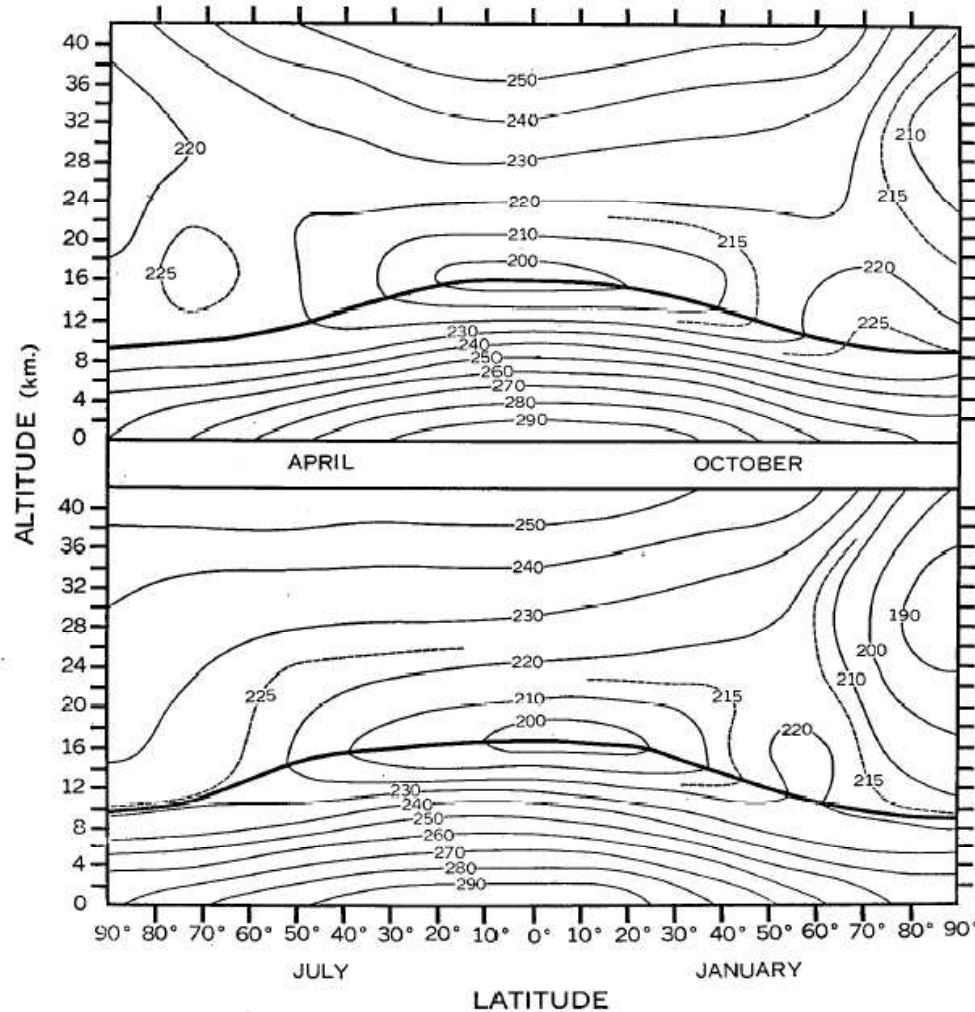
FIG. 12. Distribution of the observed temperature (deg k) in the northern hemisphere for different seasons. From J. London (1956).
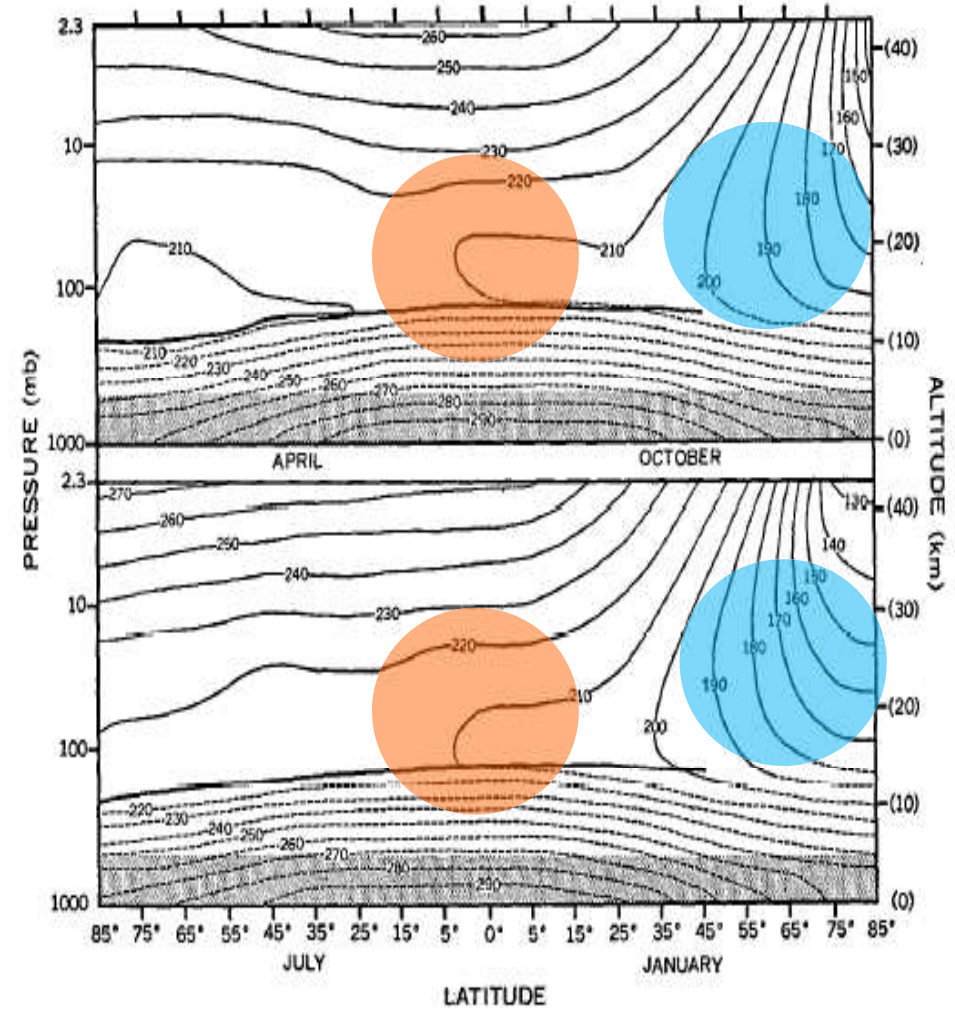
FIG. 13. The local radiative equilibrium temperature of the stratosphere. The shaded area is the region where the temperature was fixed at the observed value. Above the shaded area the state of the convective equilibrium, whose critical lapse rate for convective adjustment is deg km⁻¹, is shown. The region covered by solid lines is in local radiative equilibrium.

Horizontal Radiative Imbalance and Circulations

Imbalanced horizontal distribution of radiative heating

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

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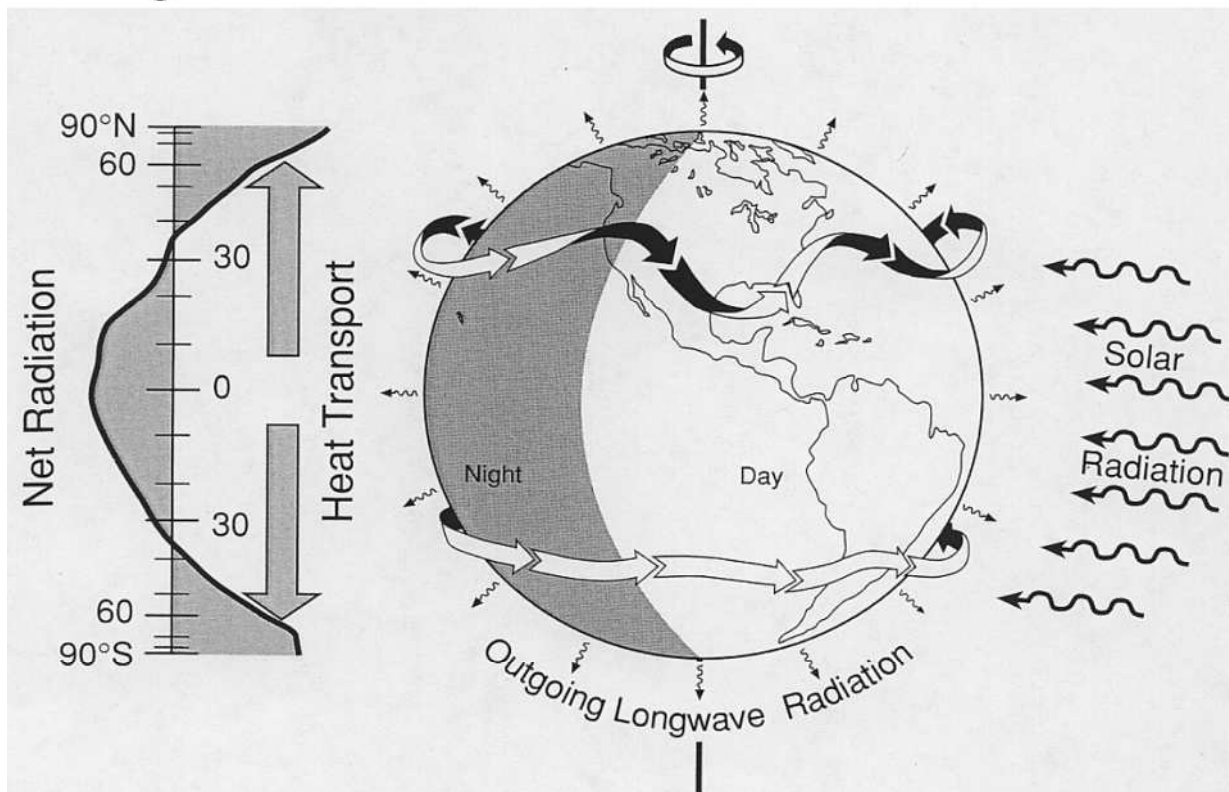
- Relaxation time to radiative equilibrium temperature (radiative equilibrium timescale) is estimated as about 30 days.
- Radiative imbalance between Pole and Tropics drives global circulations.
- Radiative imbalance between day and night has small influence on global circulations directly.

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

Diurnal Cycle of Precipitation from TRMM

From Takayabu, Y.N., 2002: GEOPHYSICAL RESEARCH LETTERS, VOL. 29, NO. 12, 1584, 10.1029/2001GL014113.

From Arakawa, O. and A. Kitoh at MRI/JMA
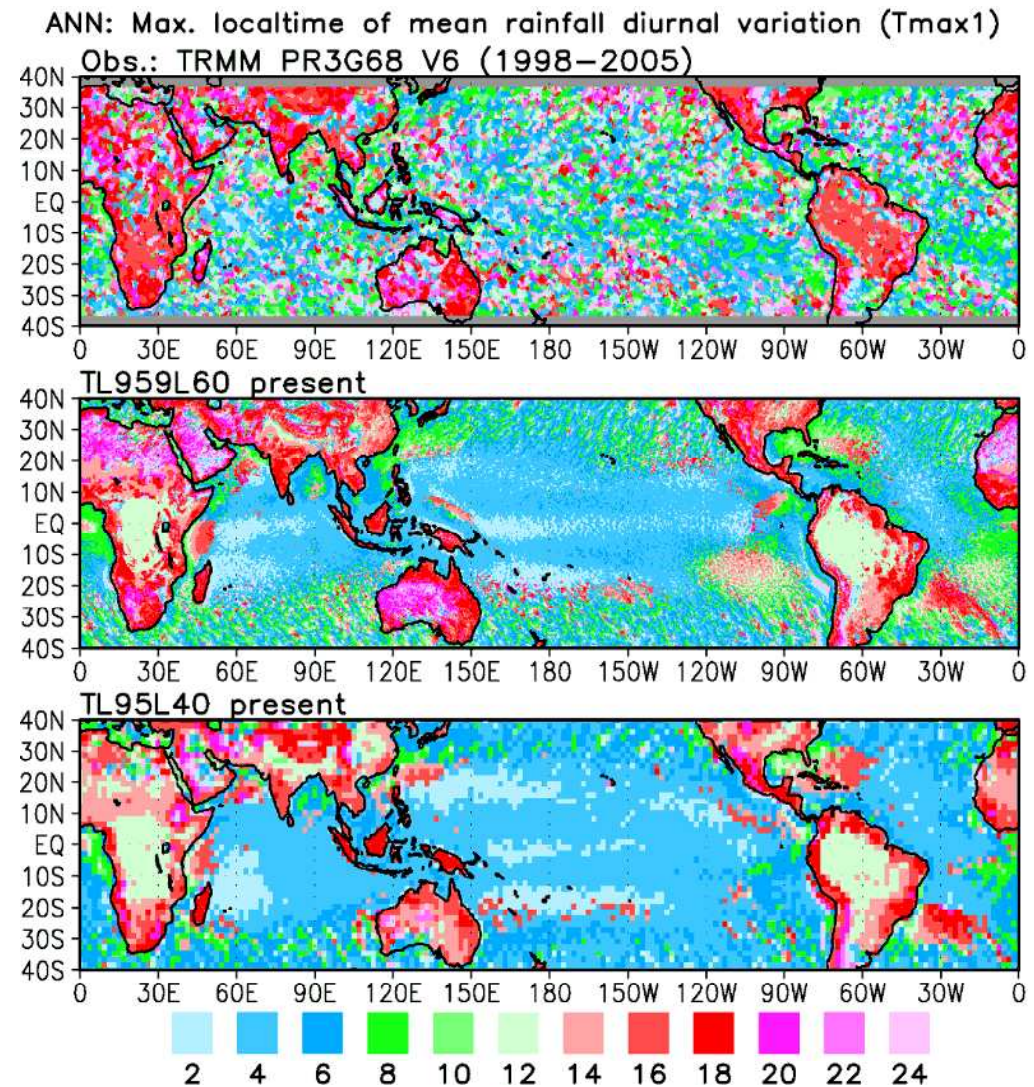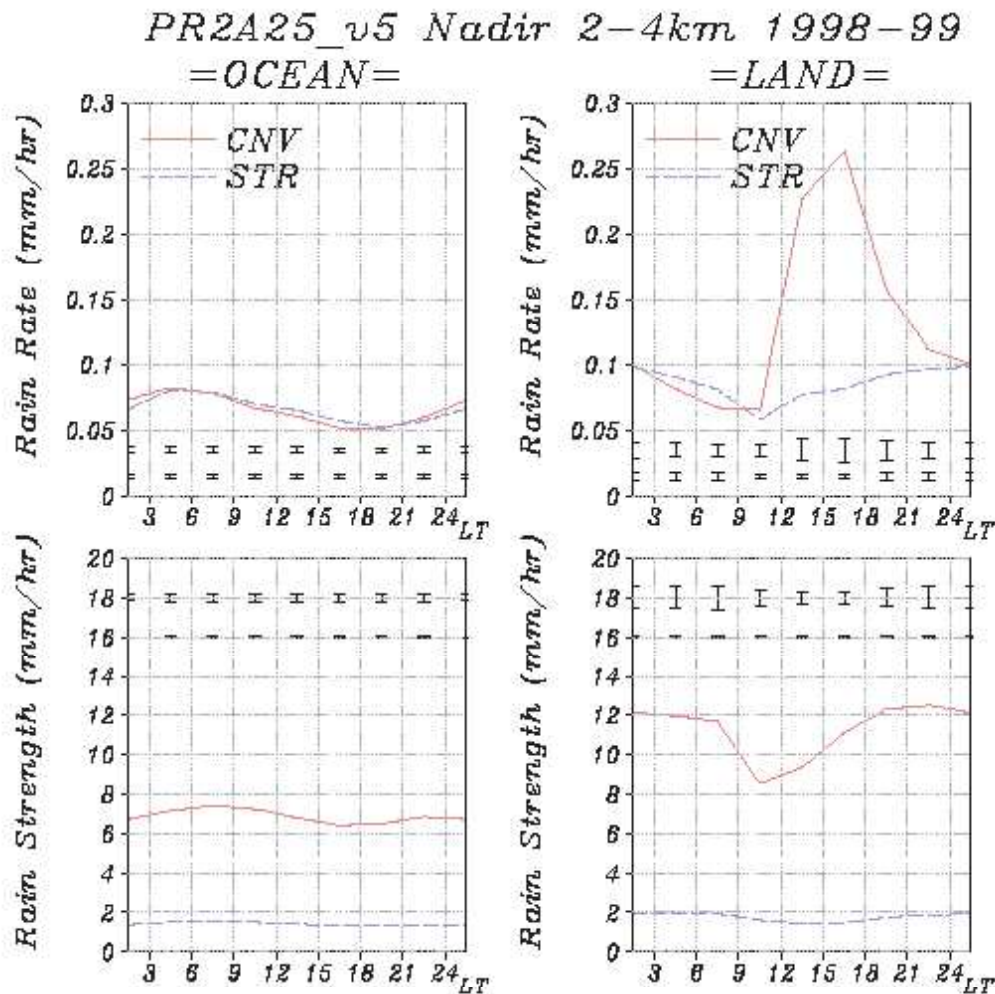
Diurnal Precipitation near Coastal Area

Diurnal Precipitation Regimes in the Global Tropics*

KAZUYOSHI KIKUCHI AND BIN WANG+

Department of Meteorology, and International Pacific Research Center, University of Hawaii at Manoa, Honolulu, Hawaii
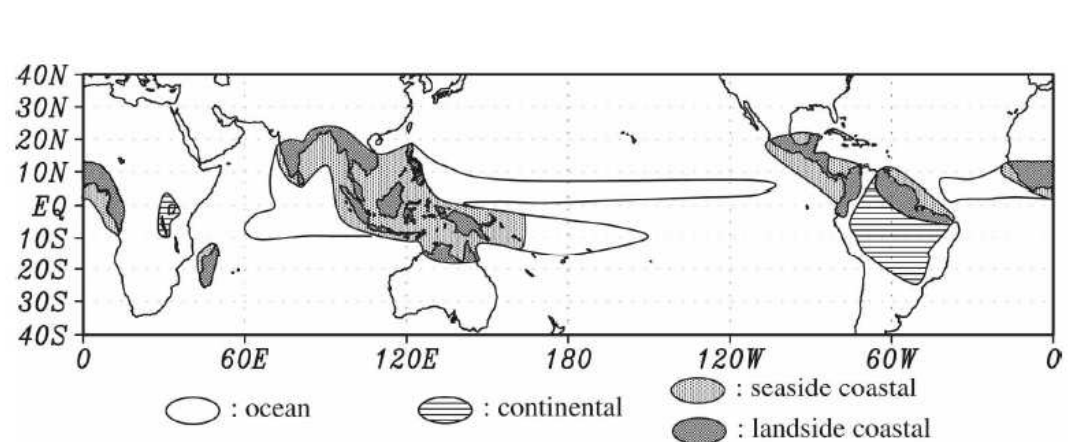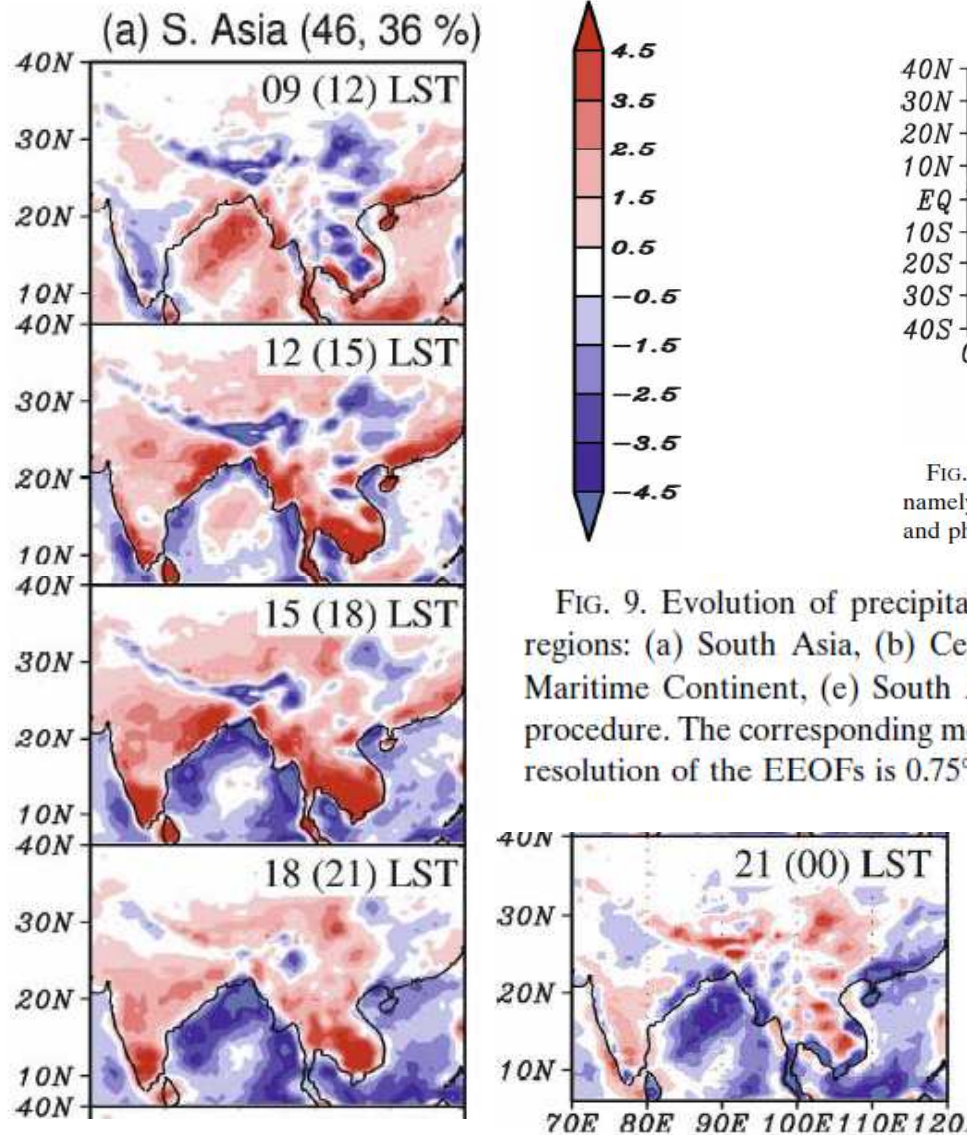
FIG. 10. Schematic diagram showing the global tropical diurnal precipitation regimes. Three regimes, namely oceanic, continental, and coastal regimes, are identified according to the amplitude, peak time, and phase propagation characteristics of the diurnal precipitation.

FIG. 9. Evolution of precipitation represented by the combination of $EEOF_1$ and $EEOF_2$ for the following regions: (a) South Asia, (b) Central America and northwest part of America, (c) West Africa, (d) Indonesia Maritime Continent, (e) South America, and (f) Madagascar. See the text for the complete description of the procedure. The corresponding modified LSTs are shown at the right corner of panels in (a) and (d). The horizontal resolution of the EEOFs is 0.75° × 0.75°.

Meridional distribution of Annual mean radiation balance
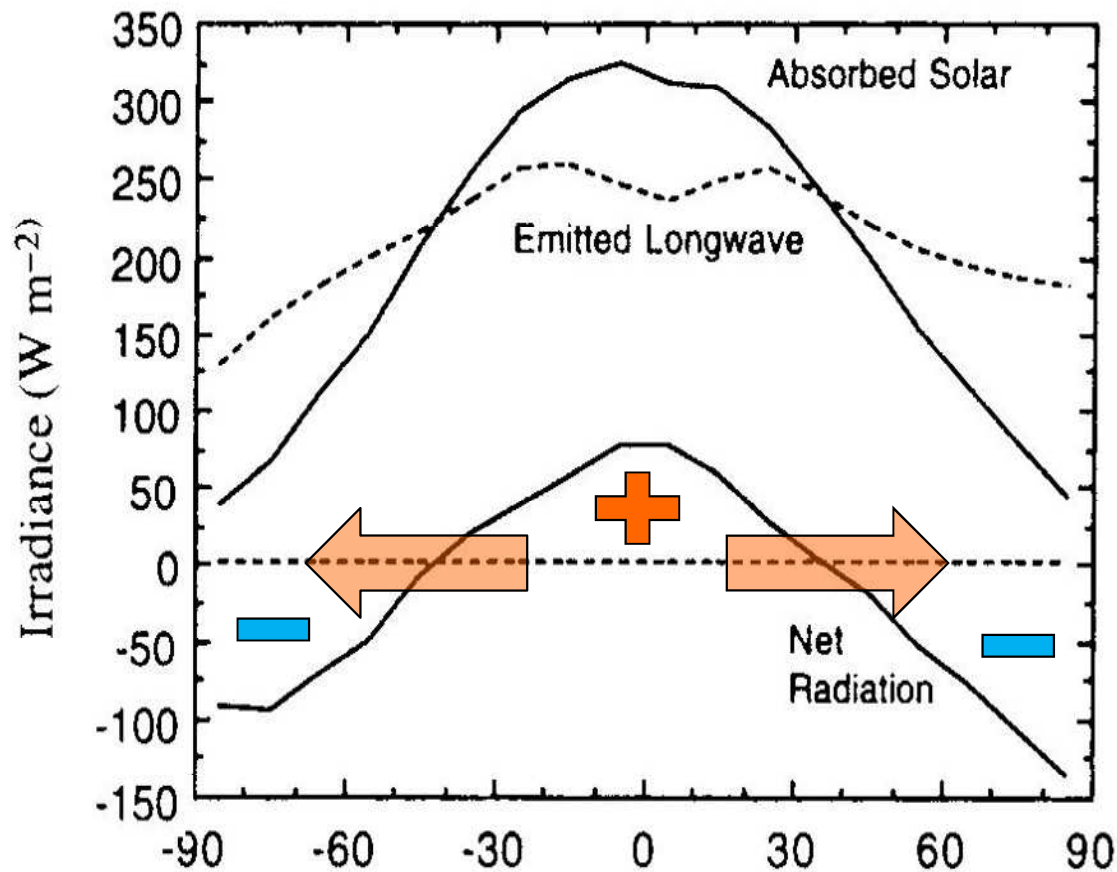
Hartmann (1994)

- **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**

Observed annual mean SST, surface wind, precipitation

(a) Sea Surface Temperature & Wind   ANN
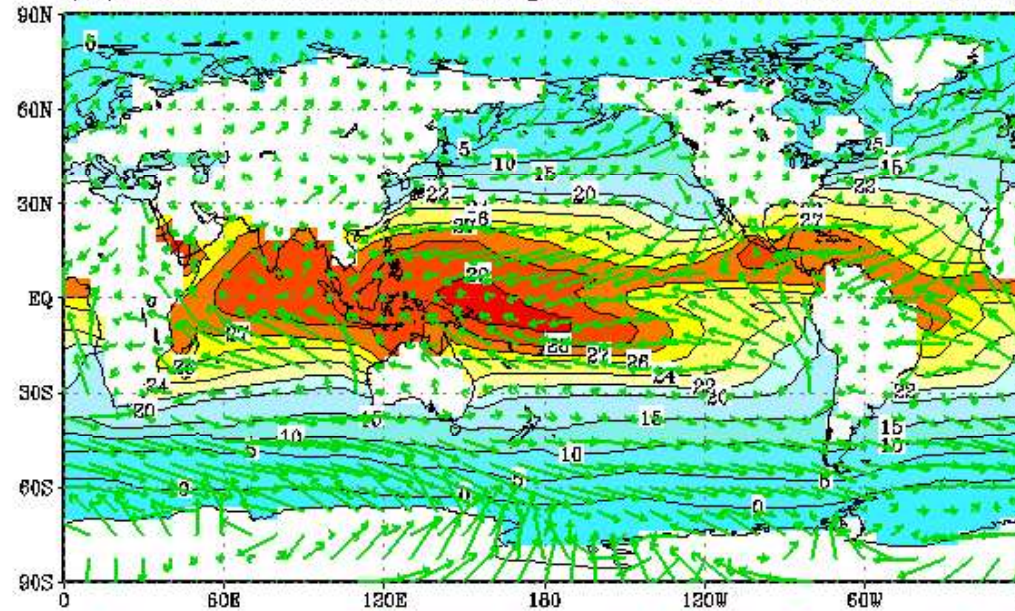
(b) Precipitation   ANN
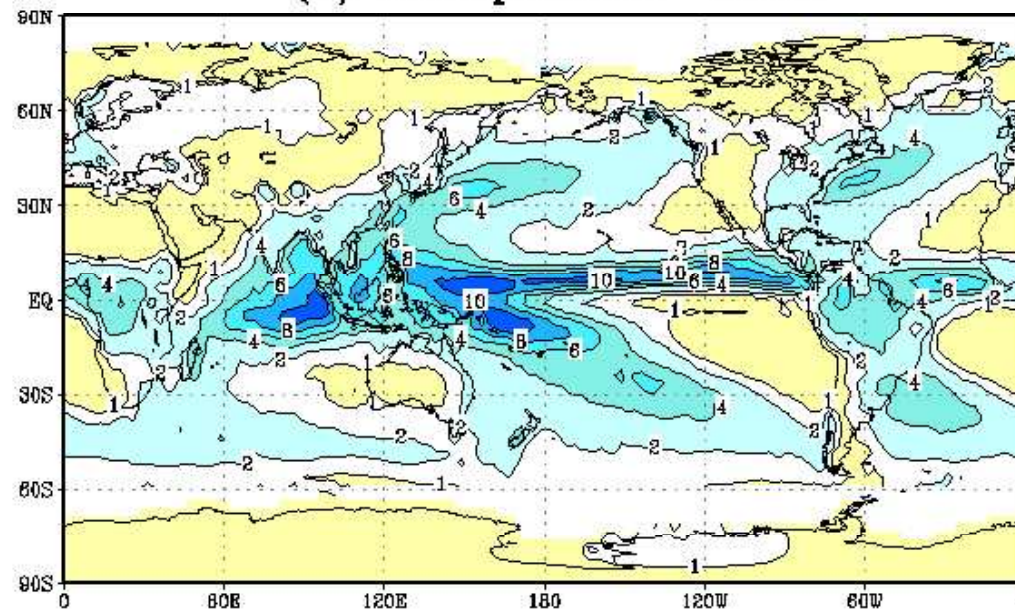
Sea surface temperature (SST) and Cumulus
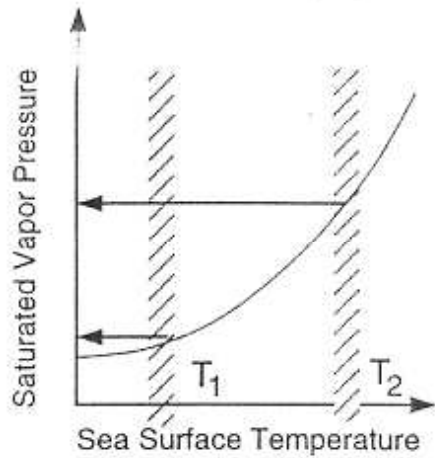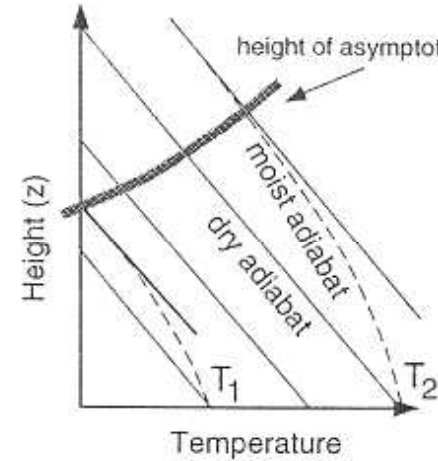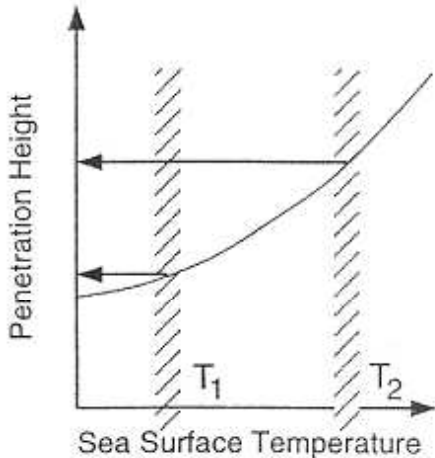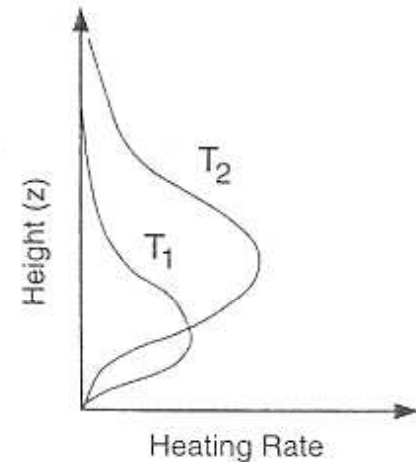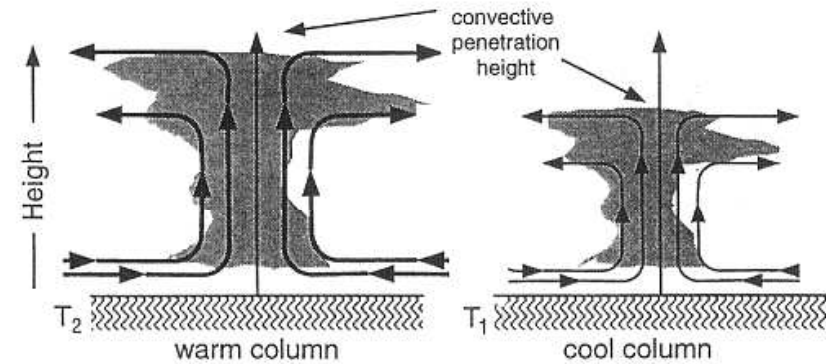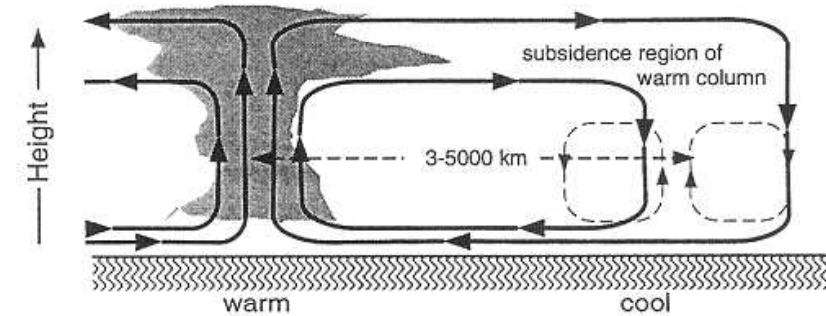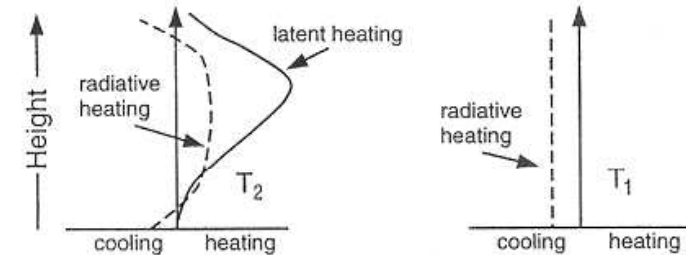
Energy Transport by Atmospheric Circulation

High energy air parcel → Energy Loss

Tropopause

Net energy Transport →

Energy Gain ← **Low energy air parcel**

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
$$H=C_pT+gZ+Lq$$

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.

Heat transport by the atmosphere and ocean

Integration from the South Pole of Net radiation absorbed in the Earth

Atmosphere

Ocean
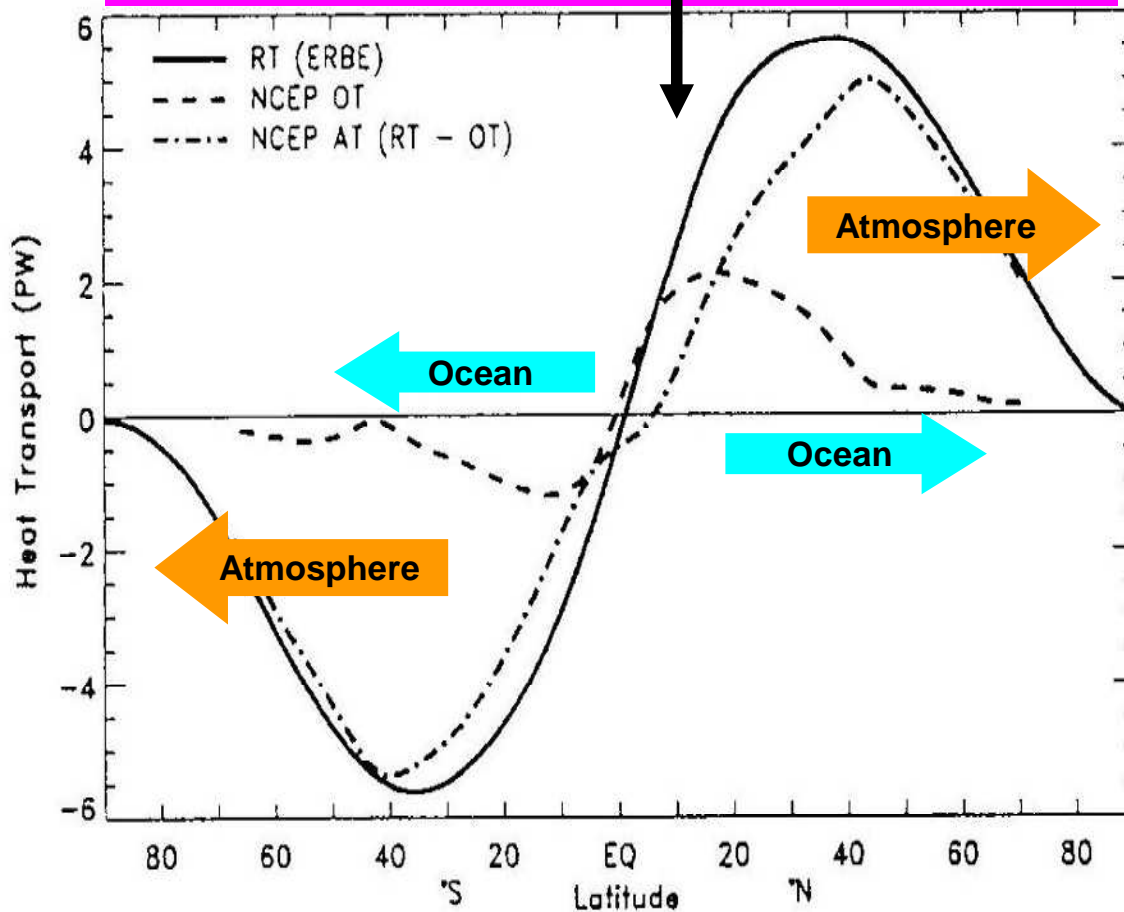
Ocean

Atmosphere

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**

Atmospheric global circulations driven by latitudinal heating contrast

From Wallace, J.M. and P. V. Hobbs, 2006: Atmospheric Science. Academic Press, 483pp.
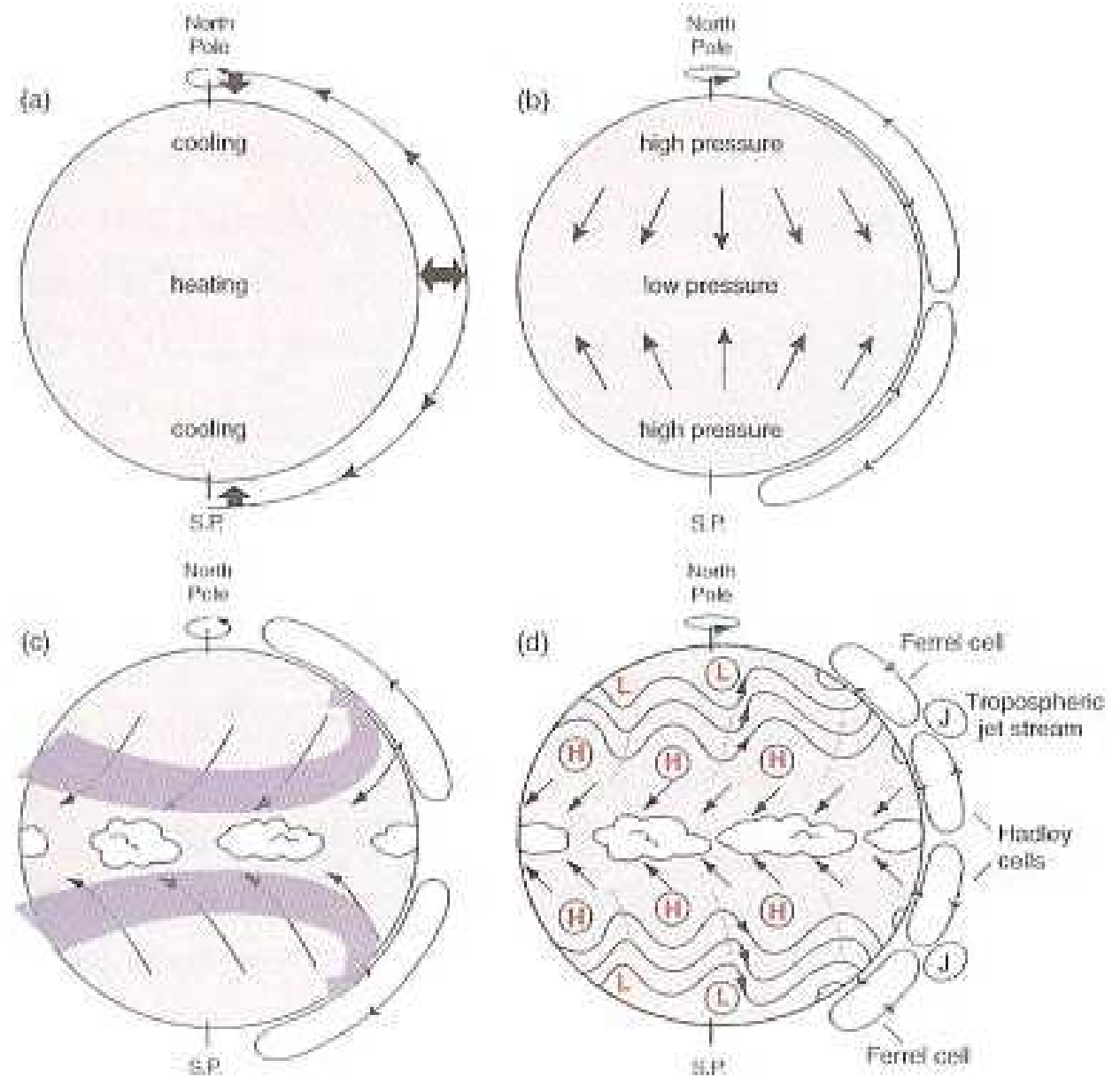
Fig. 7.21 Schematic depiction of the general circulation as it develops from a state of rest in a climate model for equinox conditions in the absence of land–sea contrasts. See text for further explanation.

Hadley (direct) circulation

These are model results.

(1) There are two direct and one indirect circulations.

Stream-function

Zonal wind
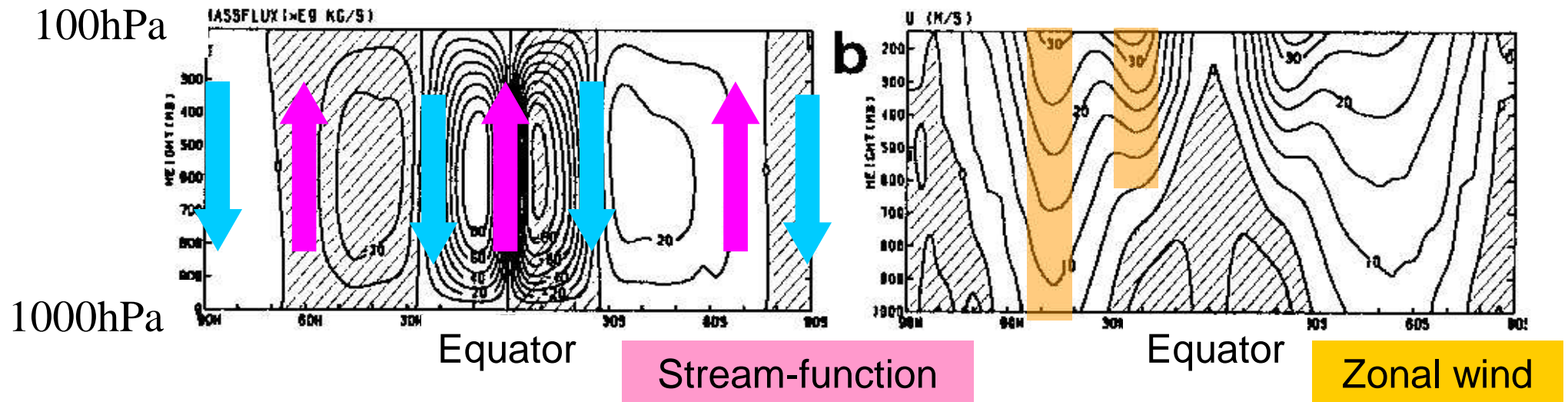
(2) After zonal waves are removed in the model, Hadley circulations and subtropical jets are left.
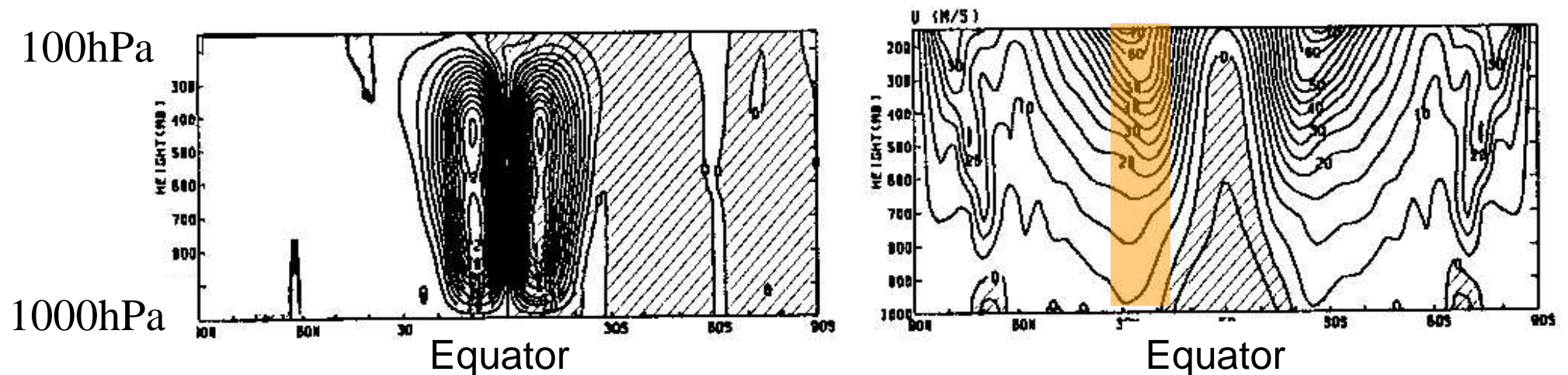
Ferrel (in-direct) circulation.

Three mechanisms to drive meridional mass circulations

Heating contrast

Heating

Cooling

Heat Transport by Waves

Heat Transport by Waves

Vertical contrast of Momentum Transport by Waves and/or Friction

Momentum Transport by Waves

Jet

Momentum Transport by Waves

$du/dt = 0$

Friction

Primary mechanism for

Hadley Circulation

Primary mechanism for

Ferrel Circulation
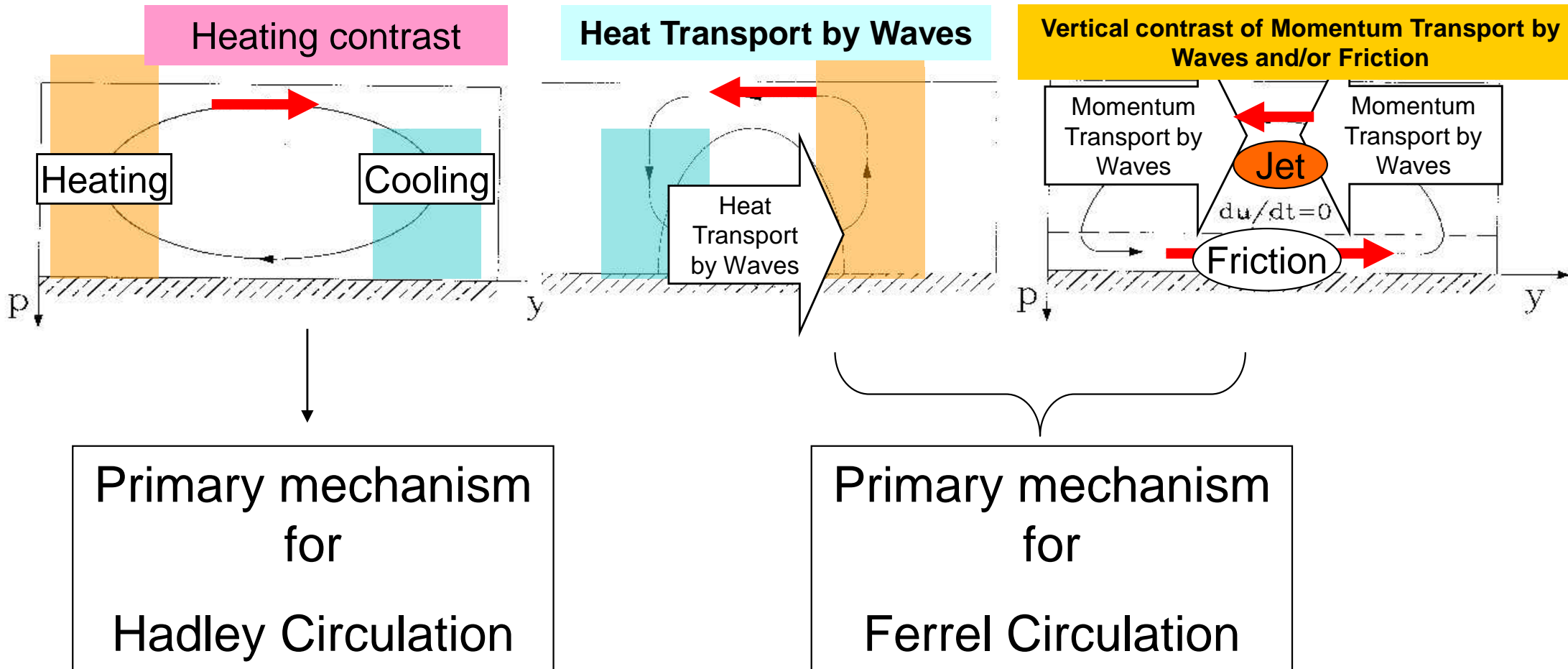
Mean meridional Circulations depend on vertical coordinates.

Iwasaki, T., 1989: A diagnostic formulation for wave–mean flow interactions and Lagrangian-mean circulation with a hybrid vertical coordinate of pressure and isentropes. *J. Meteor. Soc. Japan,* **67,** 293–312.
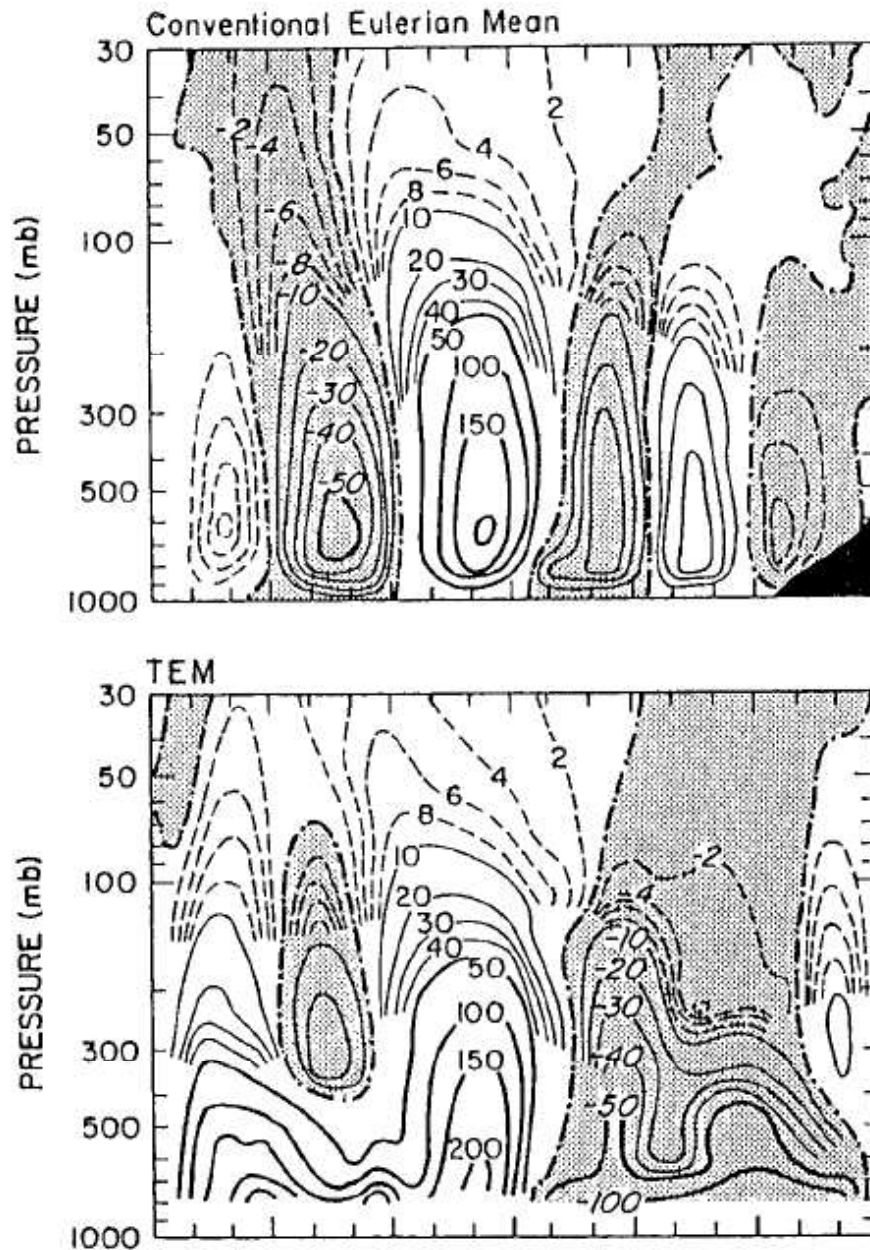
Lagrangian Mean

Net mass transport including wave effect
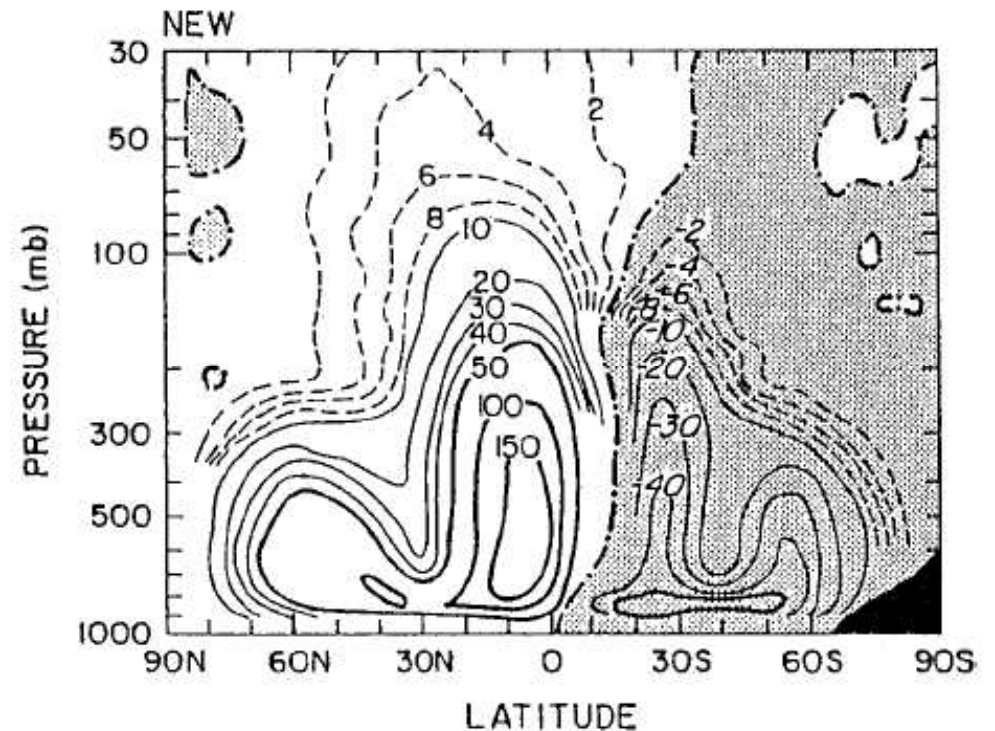
FIG. 1. Mass streamfunctions of the perpetual Jan simulation by the National Center for Atmospheric Research Community Climate Model version 1, diagnosed with the conventional Eulerian mean, the TEM, and the p_+ analyses. Contour values are shown with units of $\times 10^9$ kg s^{-1} (after I89).

Seasonal Change

orbit

March 21

December 21

June 21

September 23

23.5°

After Phim and Imbrie (1986/1987)
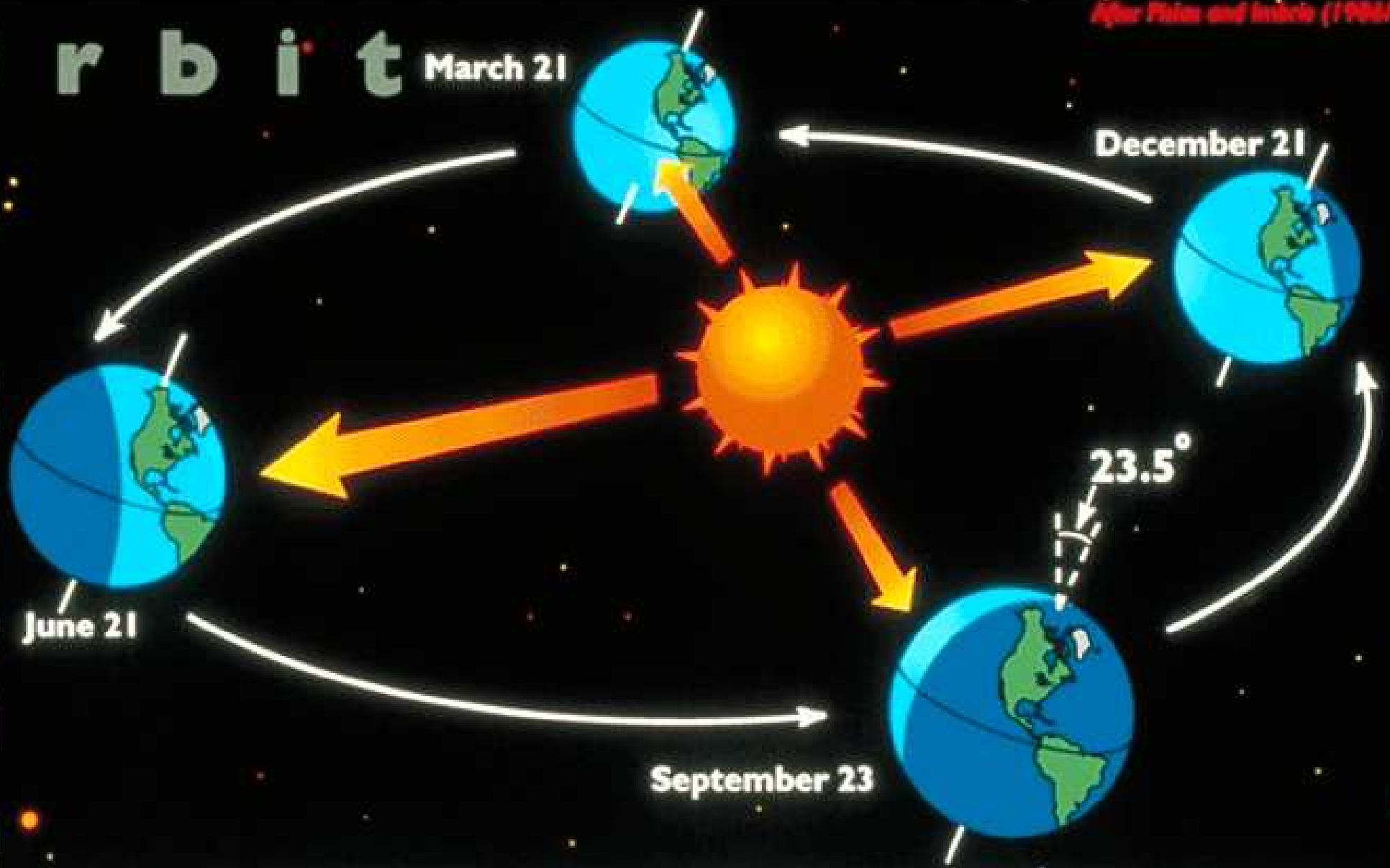
Viewed in the present, the tilted Earth revolves around the Sun on an elliptical path. The orientation of the axis remains fixed in space, producing changes in the distribution of solar radiation over the course of the year. These changes in the pattern of radiation reaching Earth's surface cause the succession of the seasons. The Earth's orbital geometry, however, is not fixed over time. Indeed, long-term variations in the Earth's orbit help explain the waxing and waning of global climate in the last several million years.

month-latitude diagram of TOA solar insolation
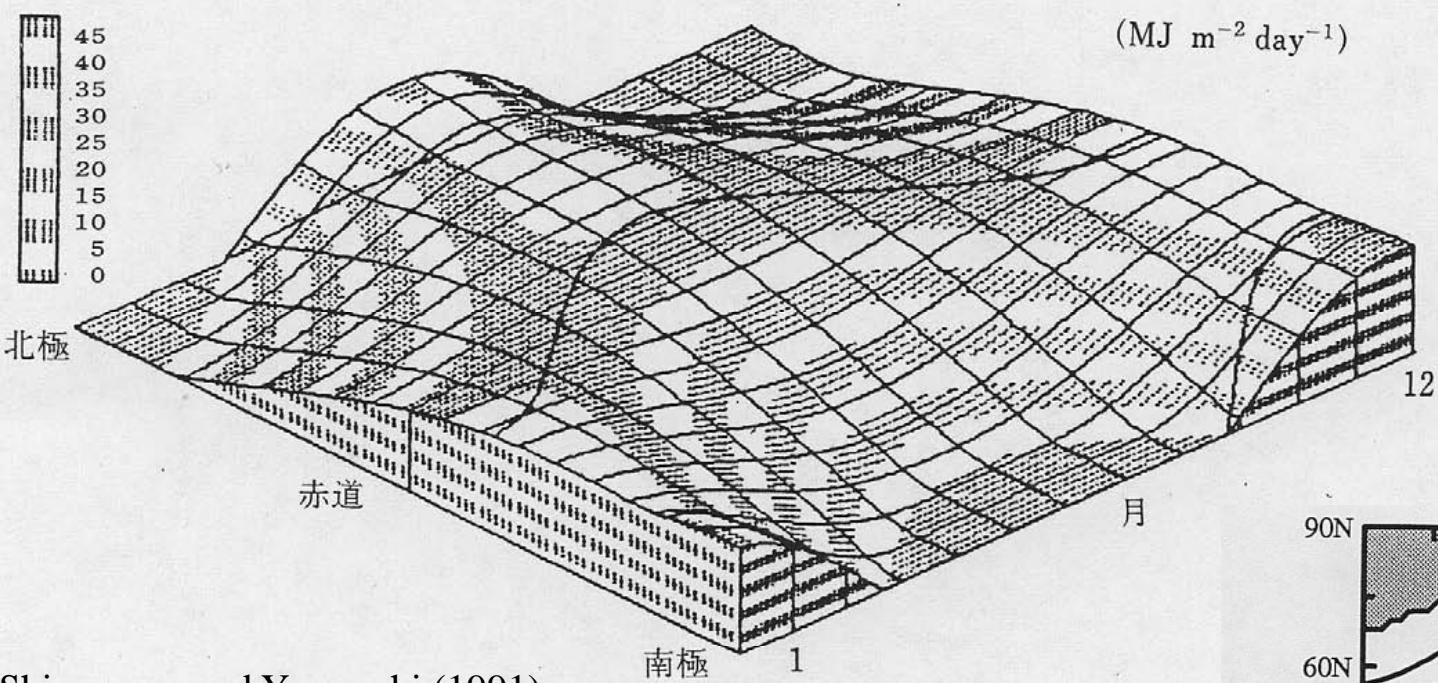
Shimamura and Yamauchi (1991)

Hartmann (1994)

Earth's orbit is not circular, and currently Earth is somewhat closer to the sun during SH summer than during NH summer. As a result, the maximum insolation in the SH is about 6.9% higher than that in the NH. Note that at the summer solstice the insolation in high latitudes is actually greater than that near the equator. This results from the very long days during summer and in spite of the relatively large solar zenith angles at high latitudes.
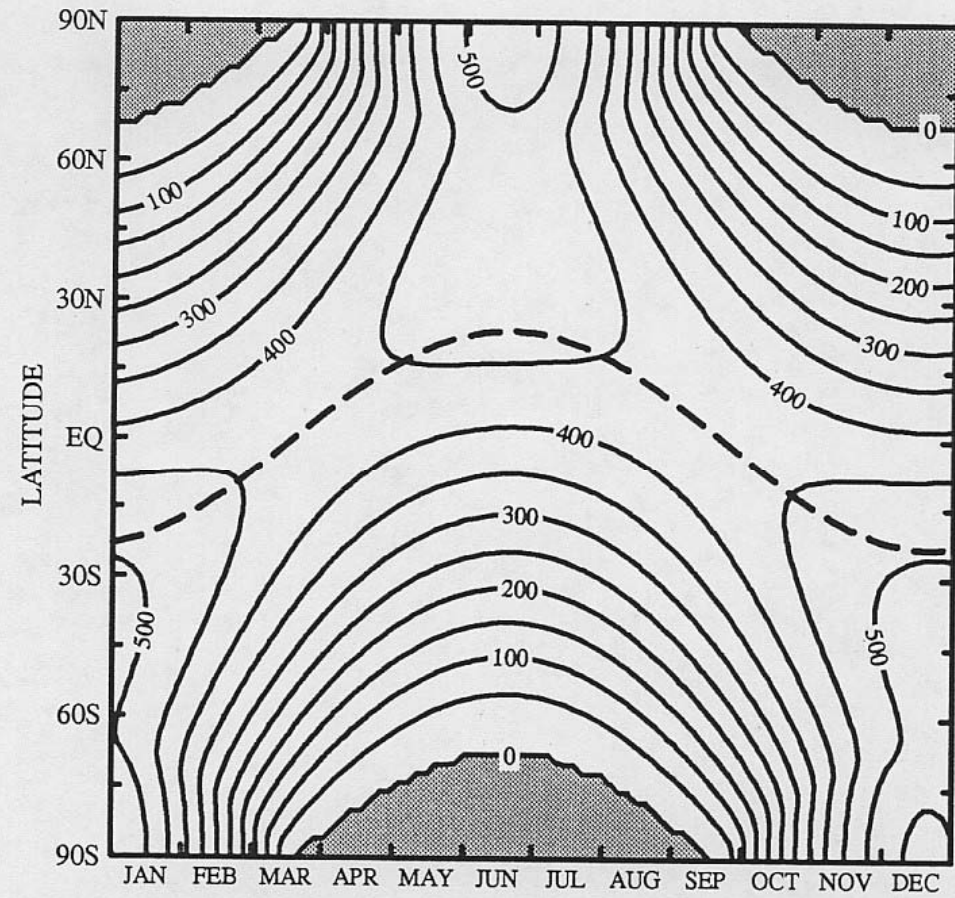
Seasonal Change of Sea Surface Temperature (SST)

Solar Insolation
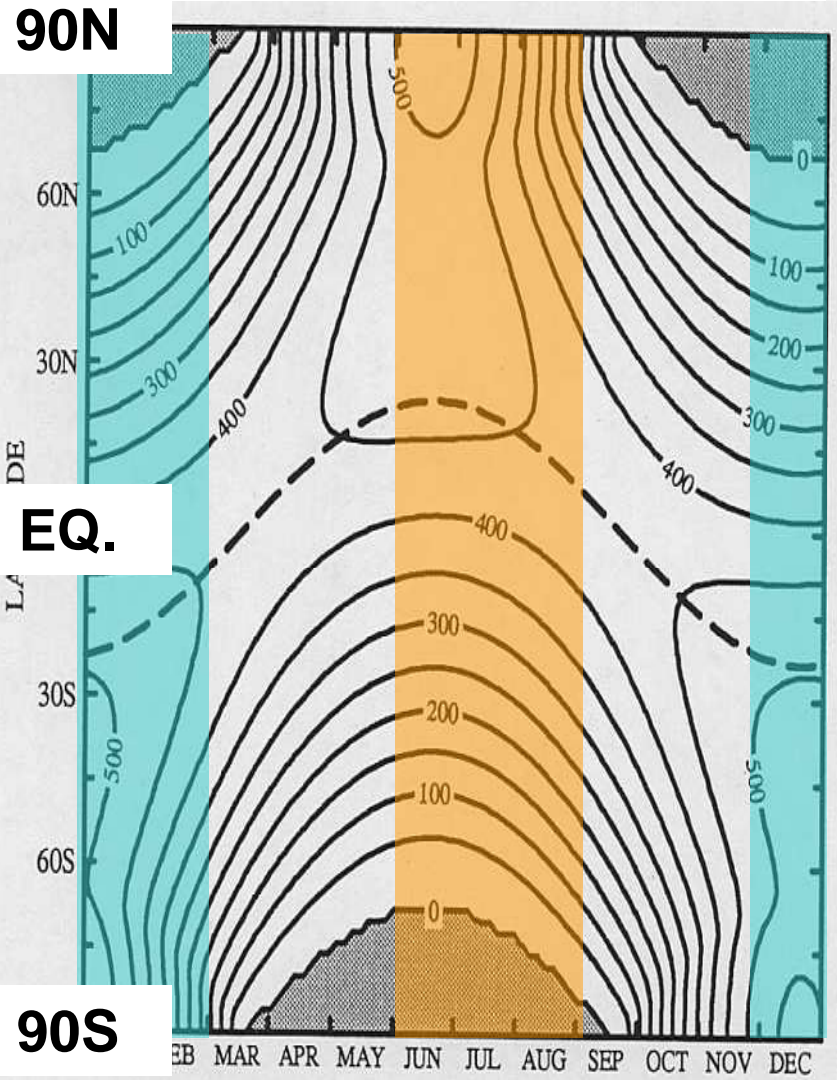
Dec.-Jan.-Feb.
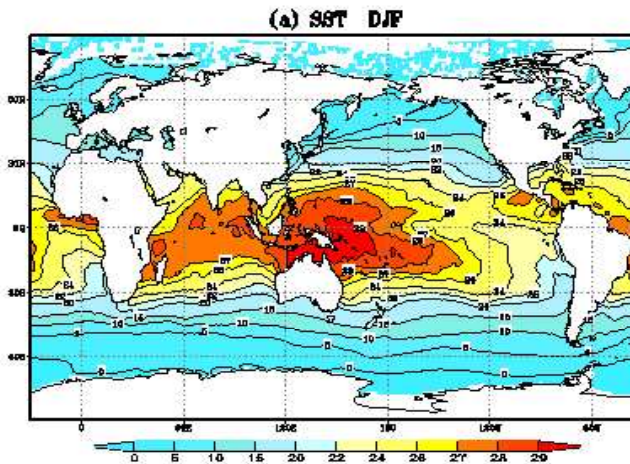
Jun-July-August
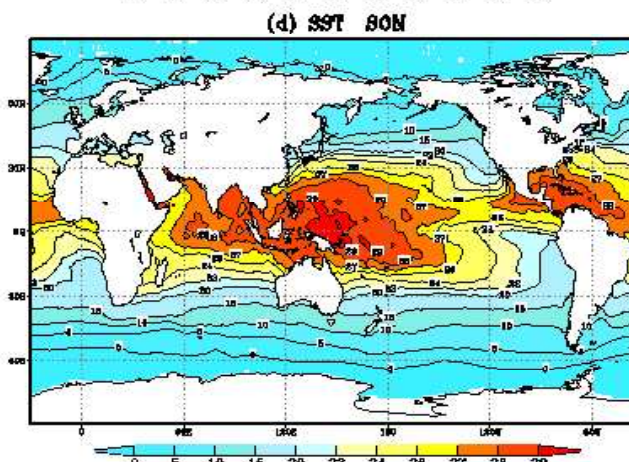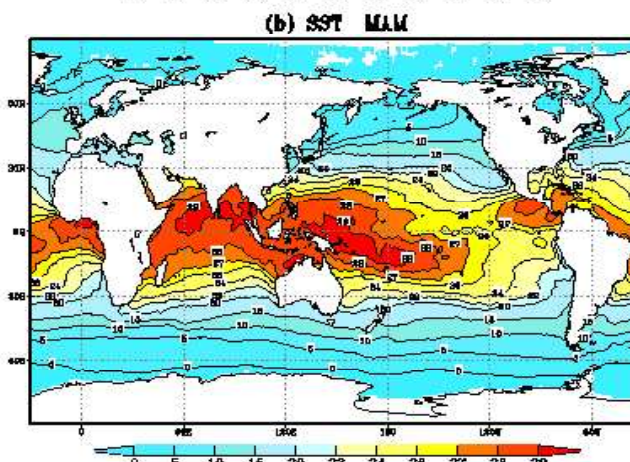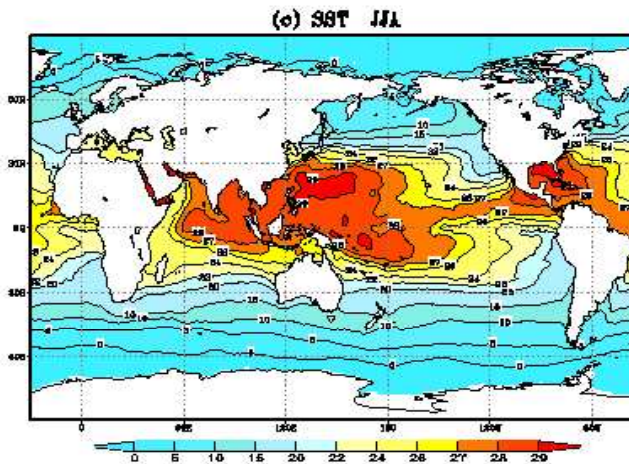
Mar.-Apr.-May

Sep.-Oct.-Nov.

SST and Precipitation in each season
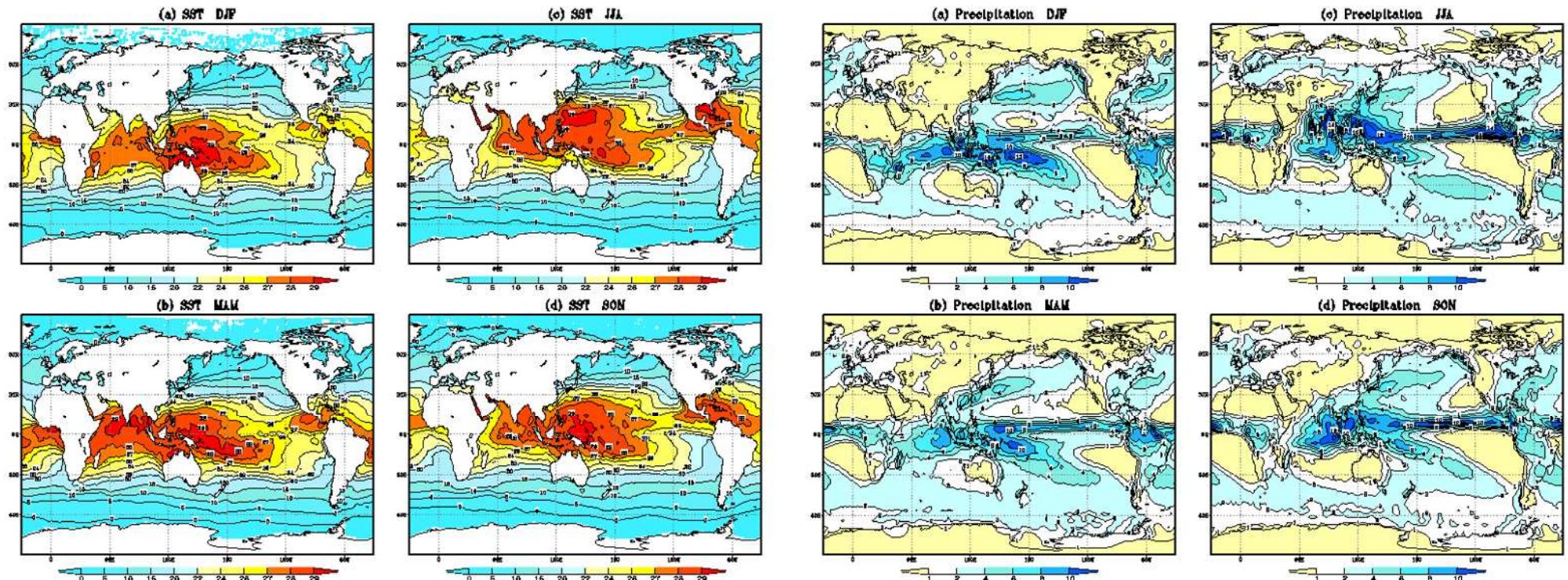
Seasonal Change of Temperature and Zonal Wind

Solar Insolation
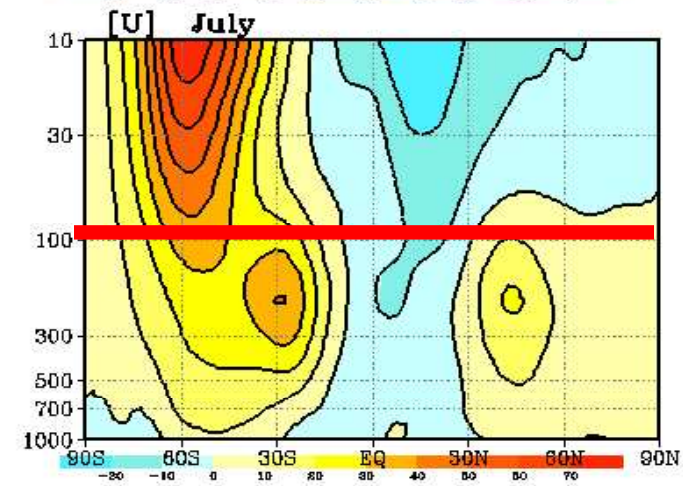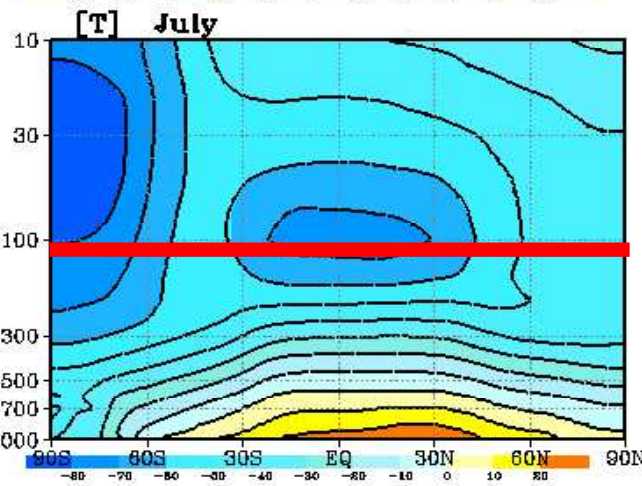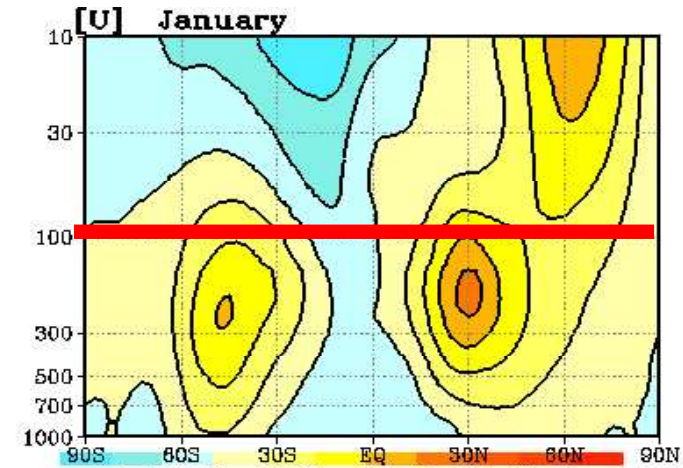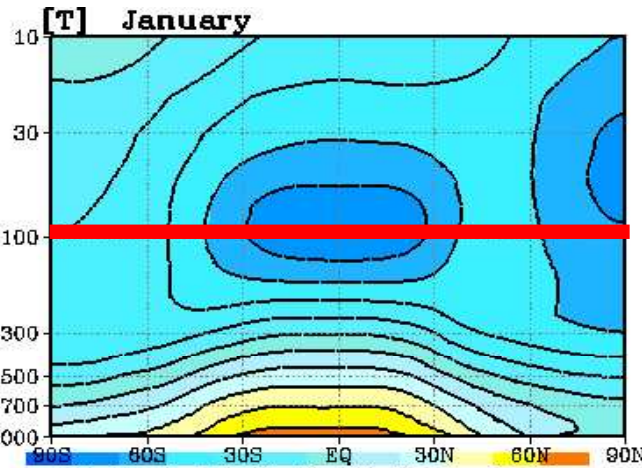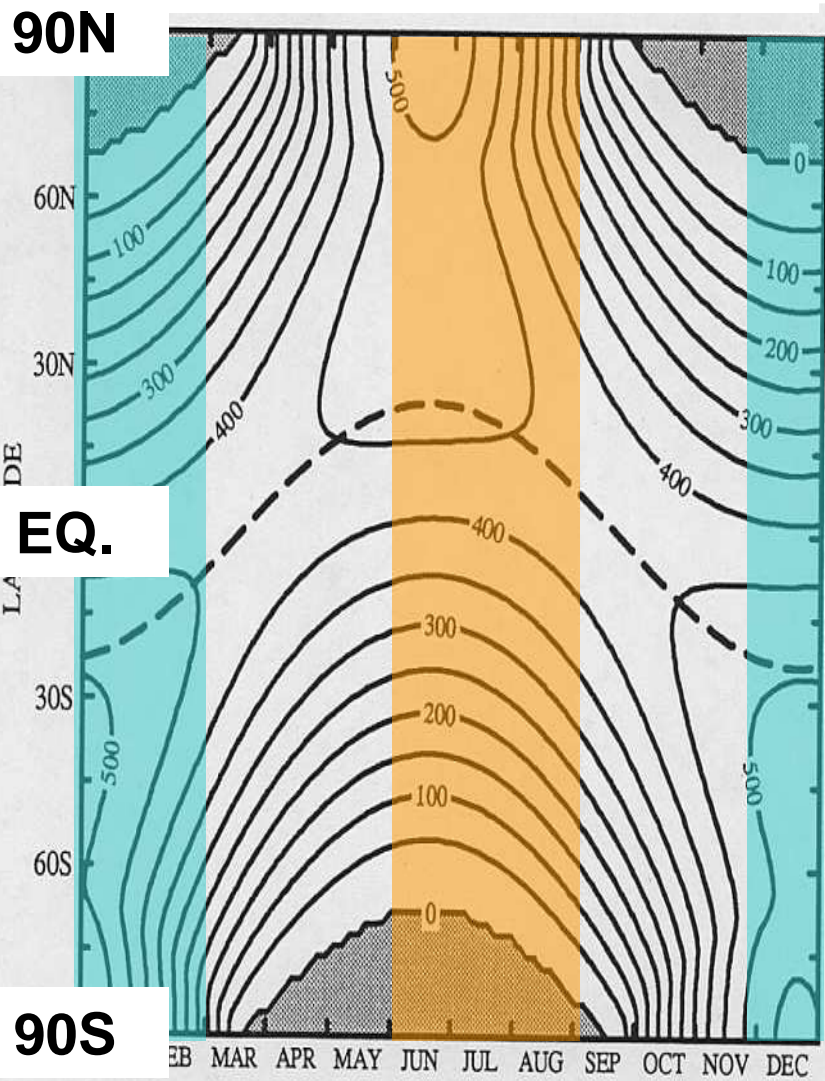
200 hPa and 850 hPa winds in JJA and DJF
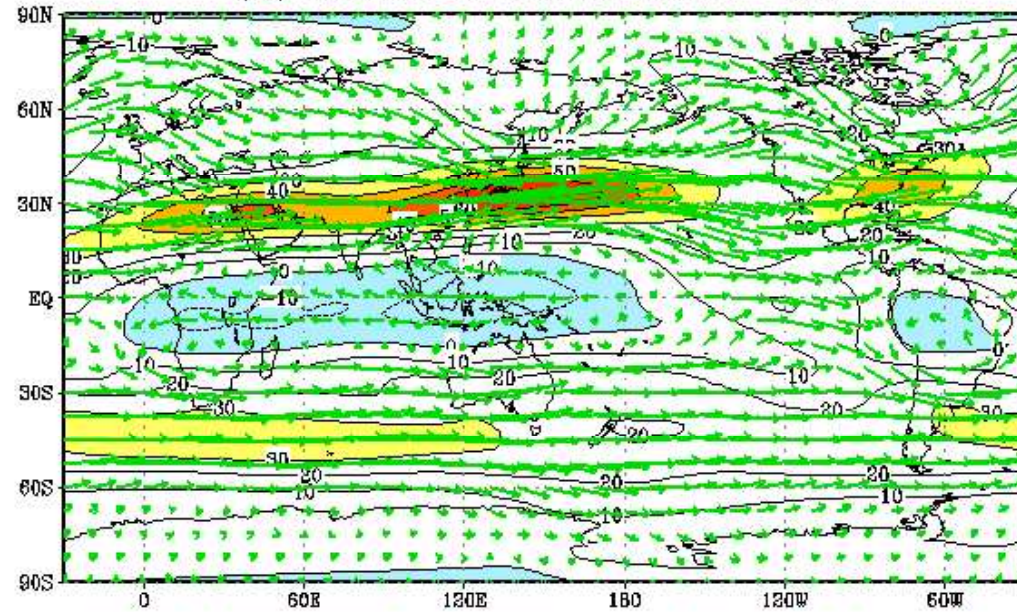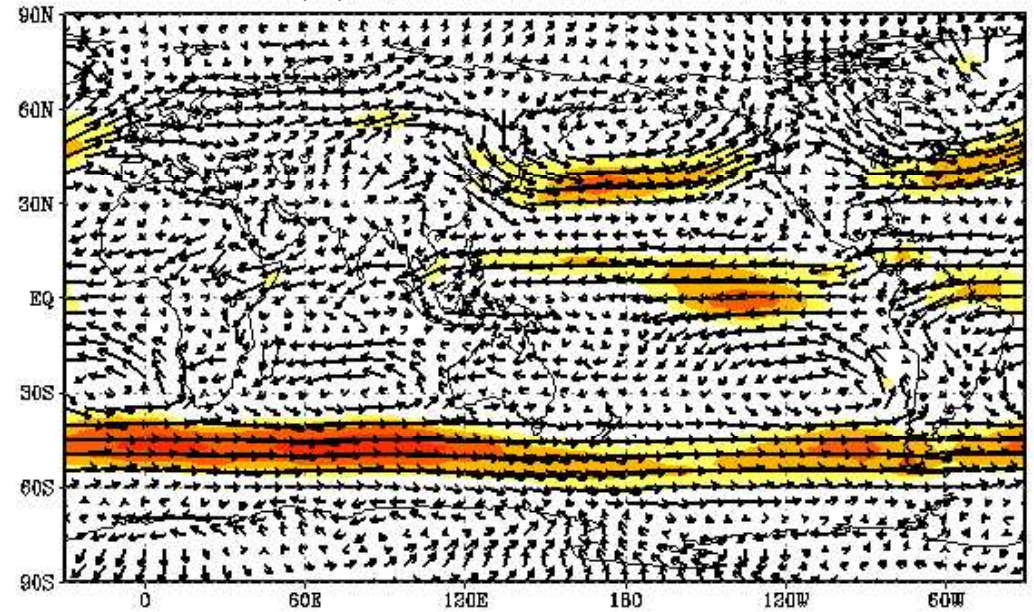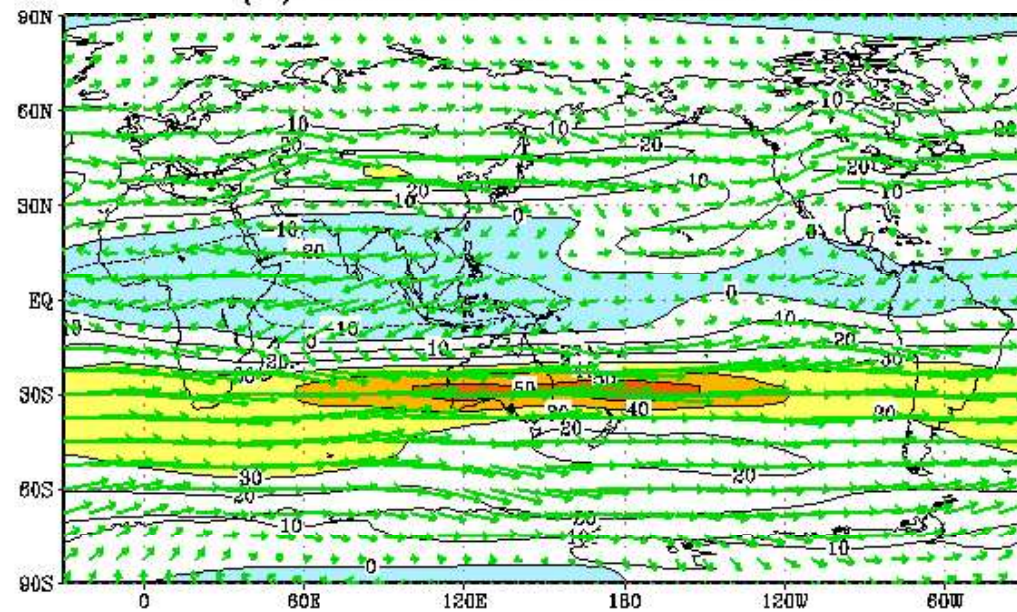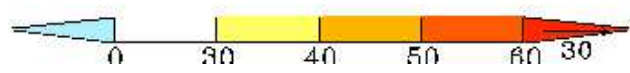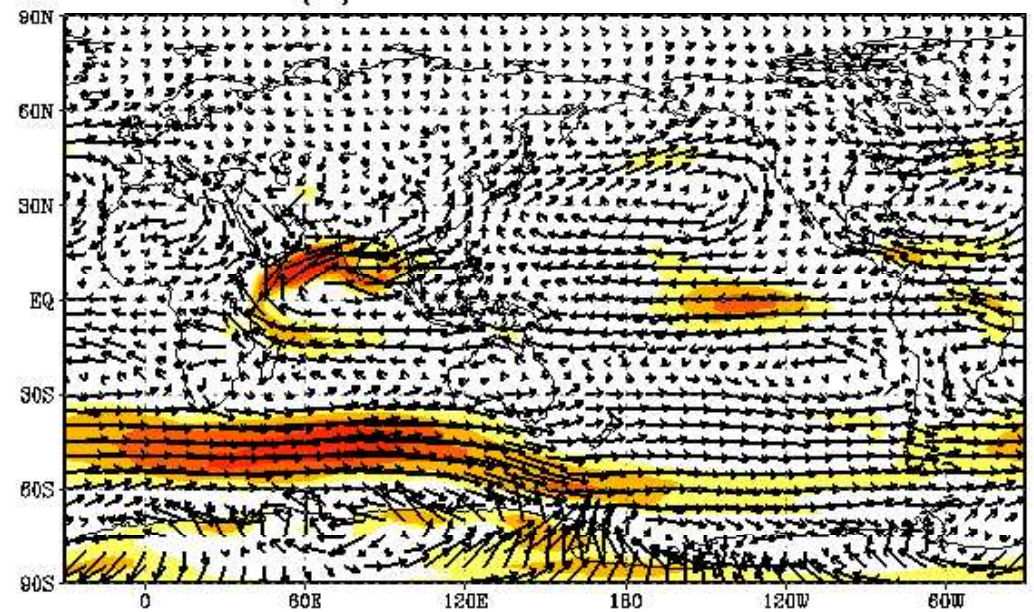
Jan-Jul contrast of surface temperature/precipitation

Northern Summer Monsoon

July   Precipitation   CMAP(1979-2001)

200hPa Stream-Function

850hPa Stream-Function

Monsoons and the dynamics of deserts.   By MARK J. RODWELL' and BRIAN **I.** HOSKINS

25N

Precipit ation

10N
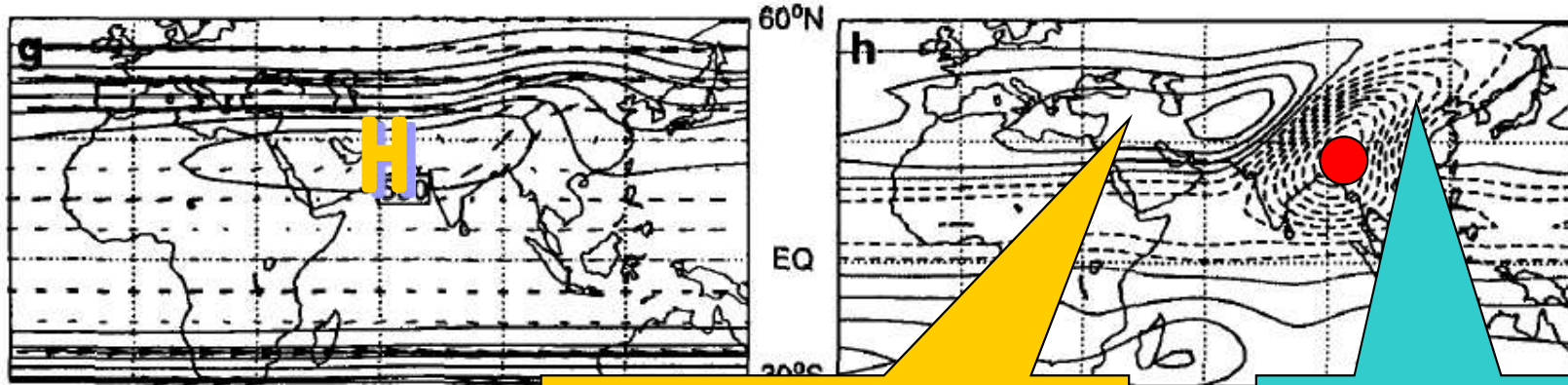
Precipit ation

Downward motion => DRY

Upward motion => WET
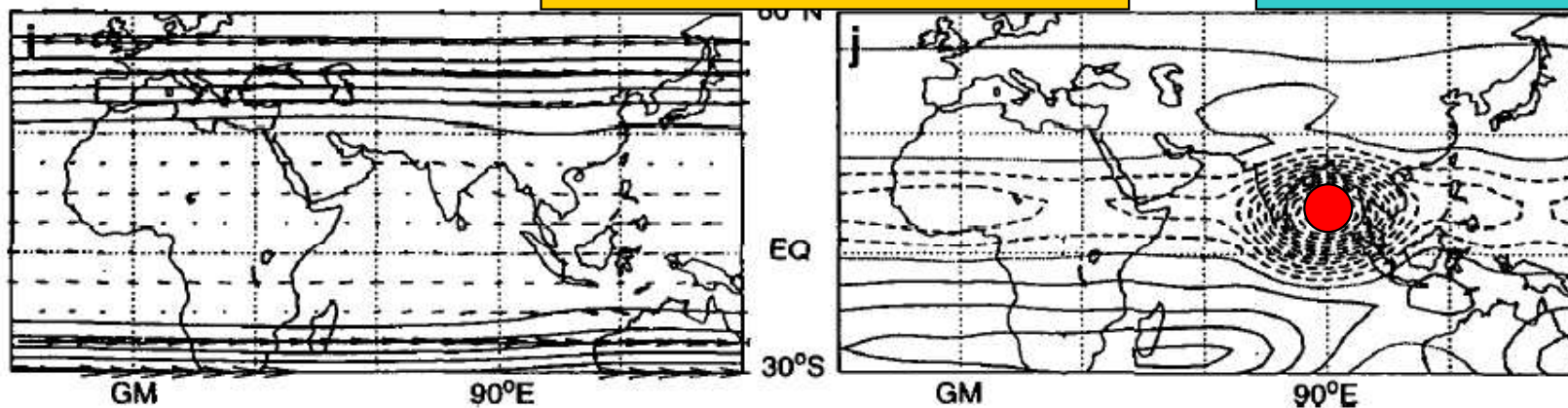
Figure 8.   Day 11 of a series of integrations without orography. (a), (c), (e), (g) and (i) pressure and horizontal winds on the 325 K isentropic surface, with contour interval 40 hPa; (b), (d), (f), (h) and (j) vertical velocity at 477 hPa, with contour interval 0.25 hPa hr^{-1}; (a) and (b) integration 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.

Precipitation and Mountain

Role of Narrow Mountains in Large-Scale Organization of Asian Monsoon Convection*

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International Pacific Research Center, SOEST, University of Hawaii at Manoa, Honolulu, Hawaii

W. TIMOTHY LIU

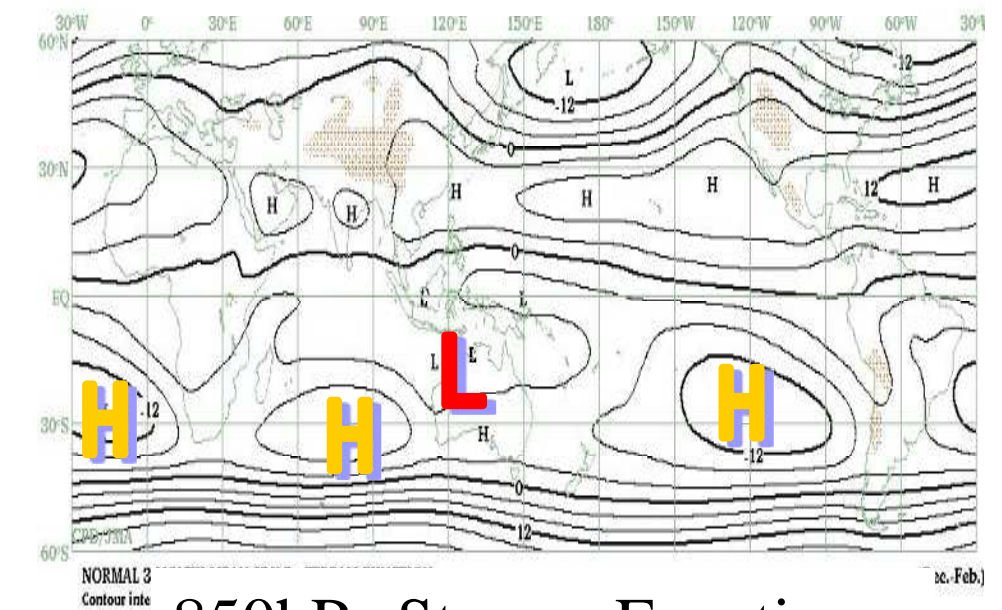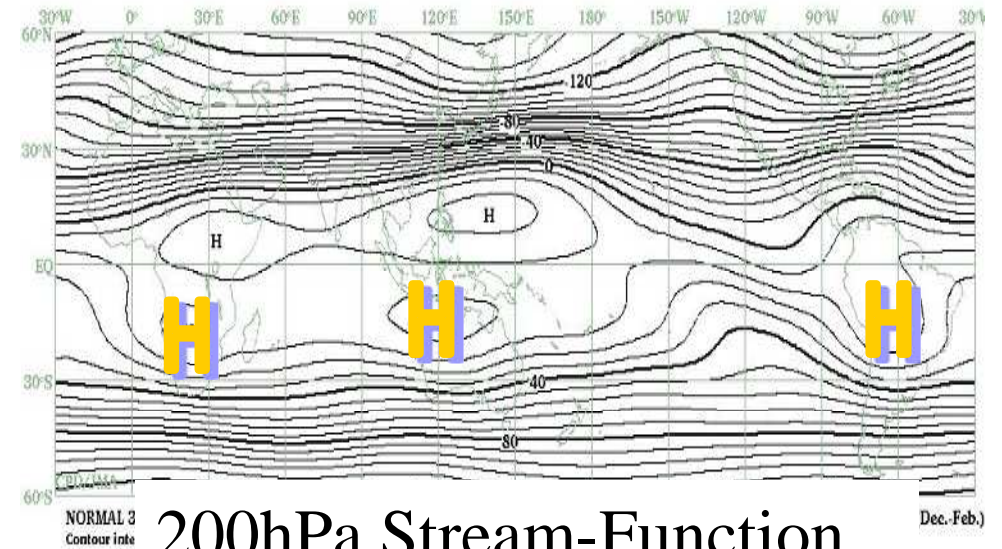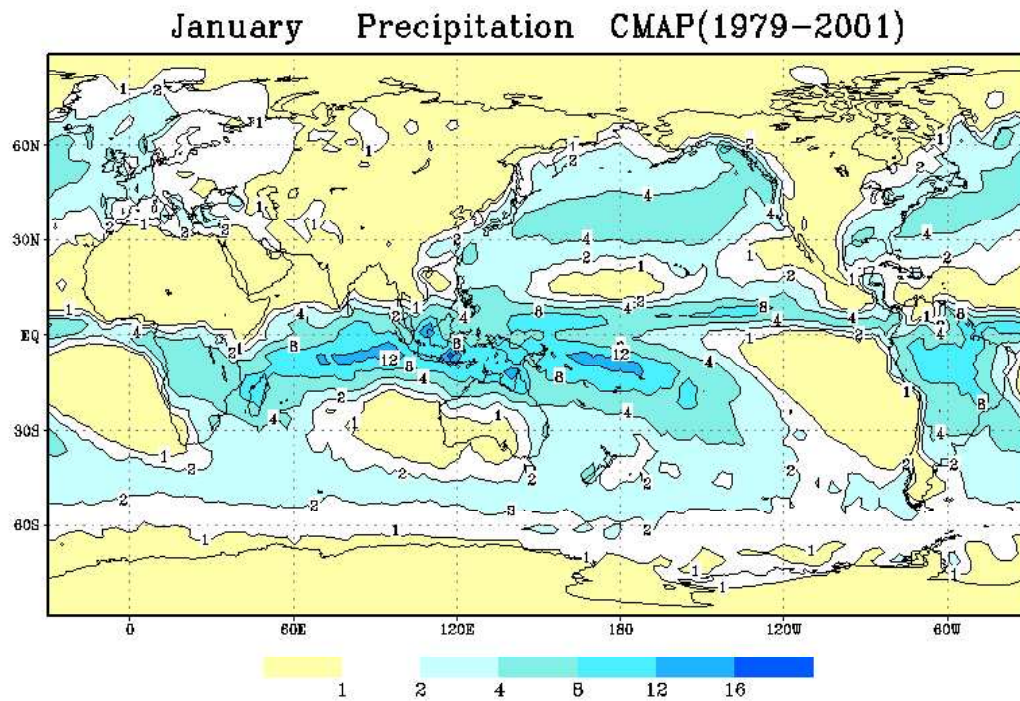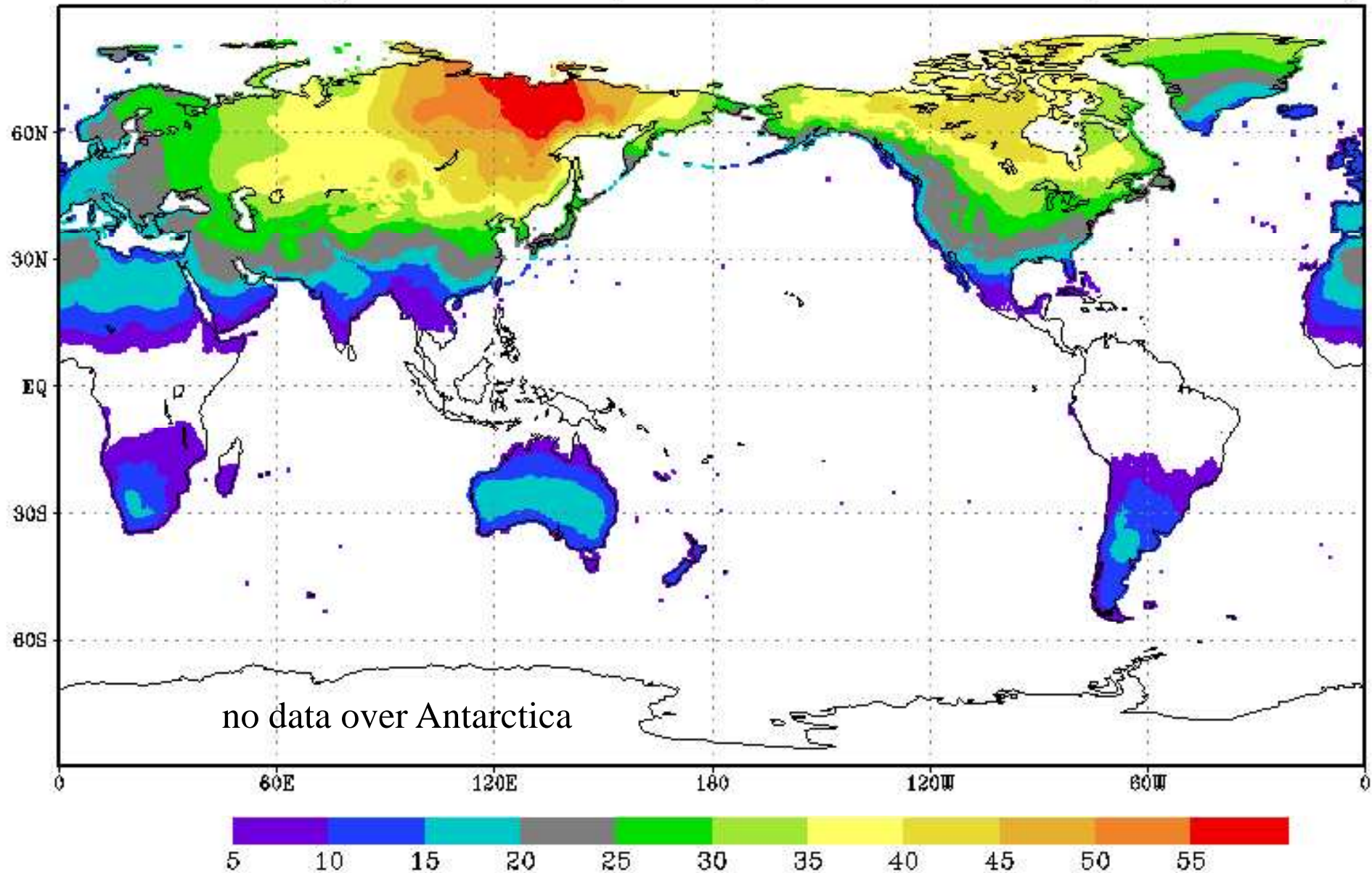Jet Propulsion Laboratory, Pasadena, California
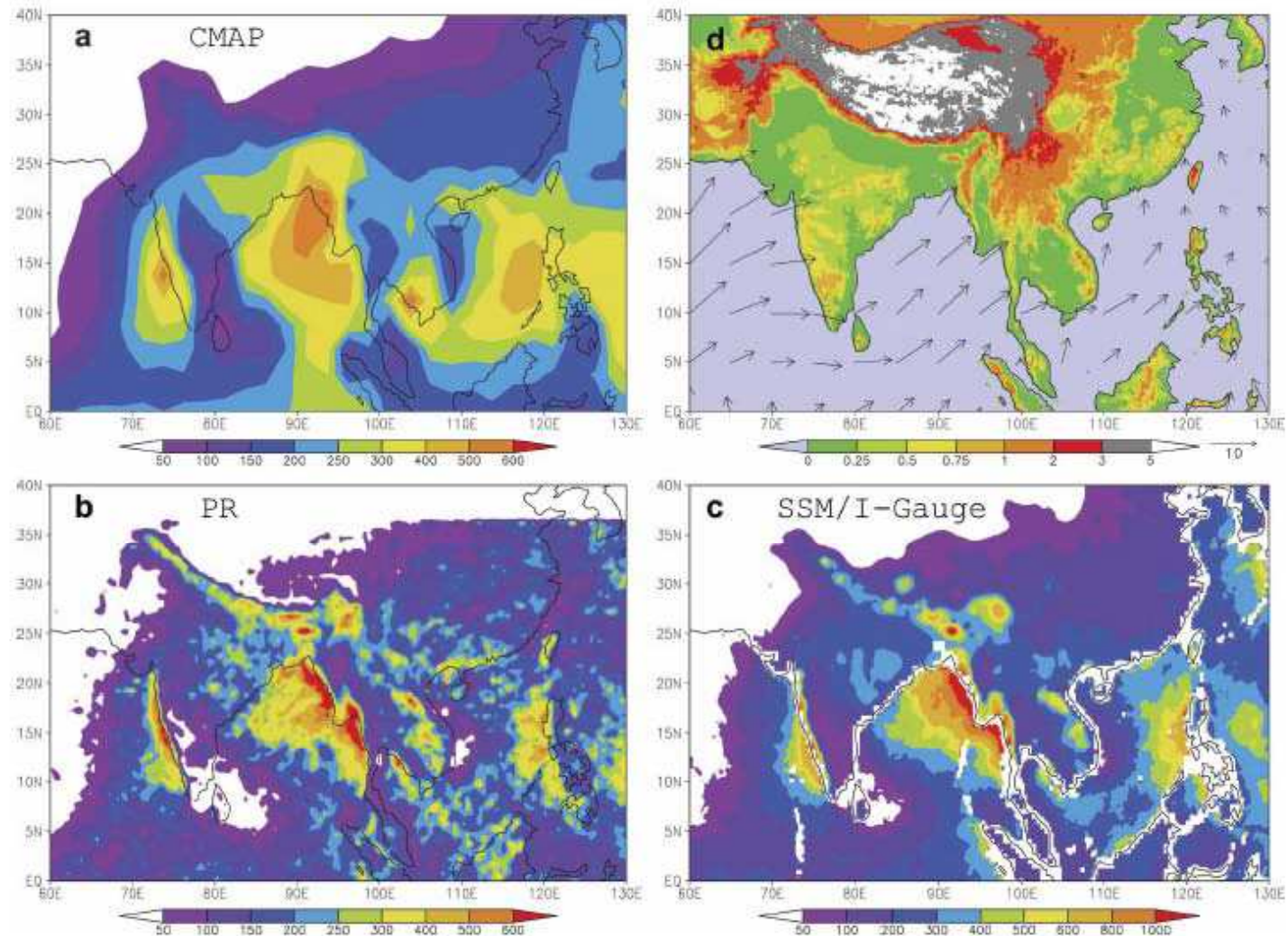
FIG. 1. Jun–Aug climatologies of surface precipitation (mm month^{-1}) based on (a) CMAP, (b) TRMM PR, and (c) SSM/I-gauge merged products. (d) Land orography (km) and QuikSCAT surface wind velocity (m s^{-1}).

Southern Summer Monsoon

January Precipitation CMAP(1979-2001)

200hPa Stream-Function

850hPa Stream-Function

Annual temperature range

Annual Range of Monthly Temperature   CRU(1901–1998)
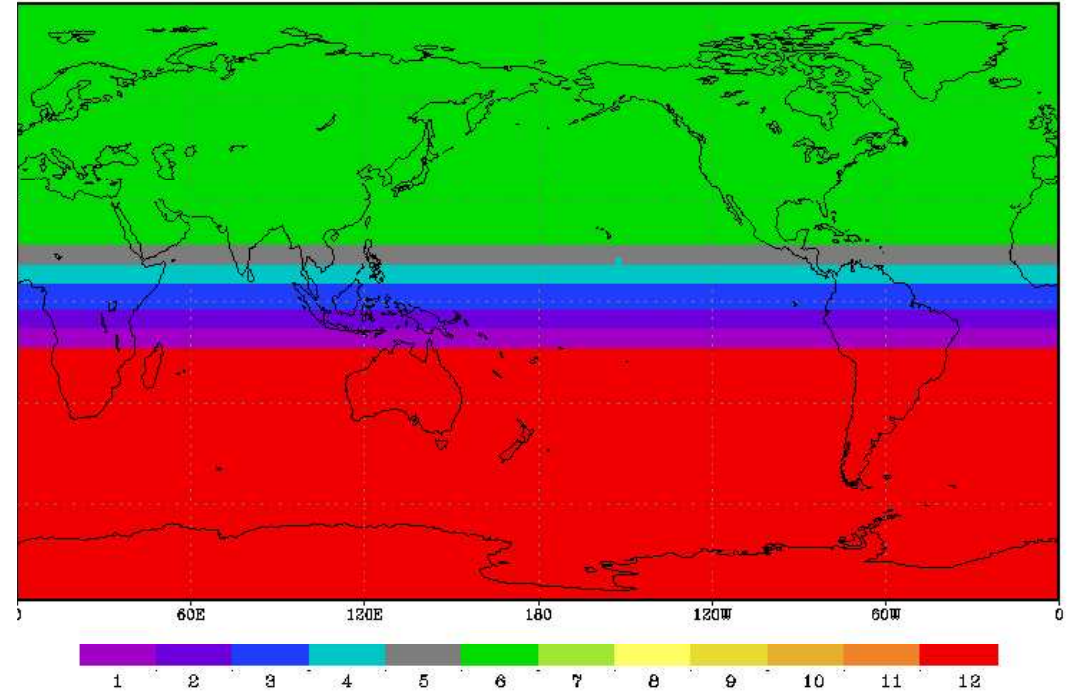
no data over Antarctica

Month of maximum monthly mean temperature

Month of Maximum T2m   NCEP(1949-2000)
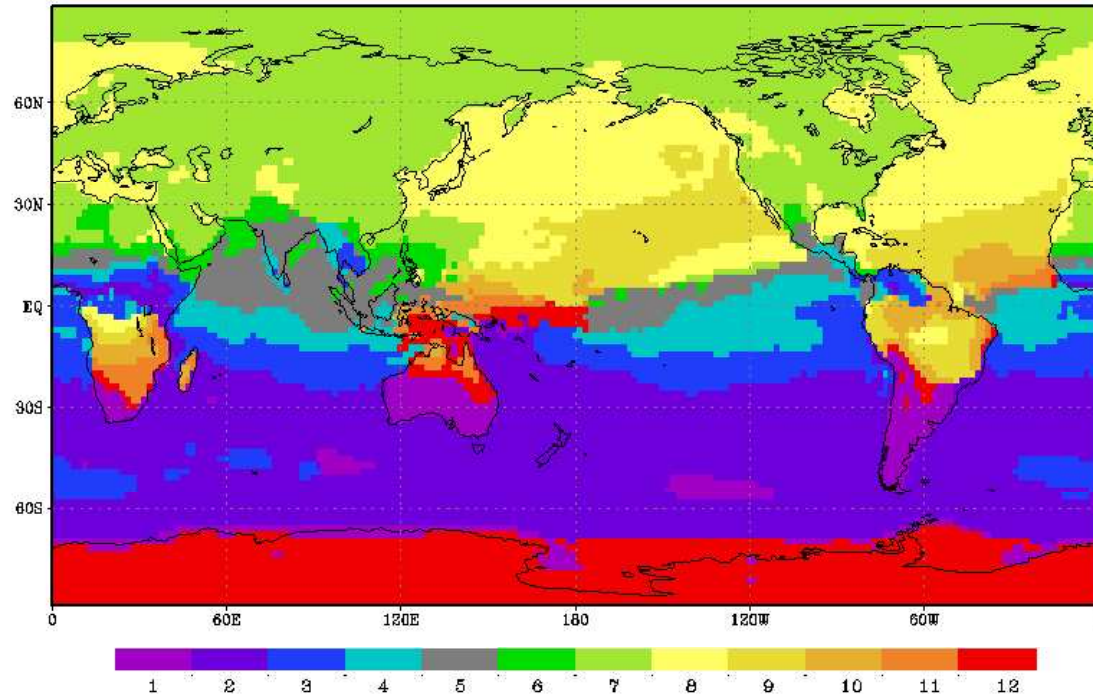
Month of Maximum Downward Solar Radiation Flux at top

(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.

Month of maximum monthly mean temperature

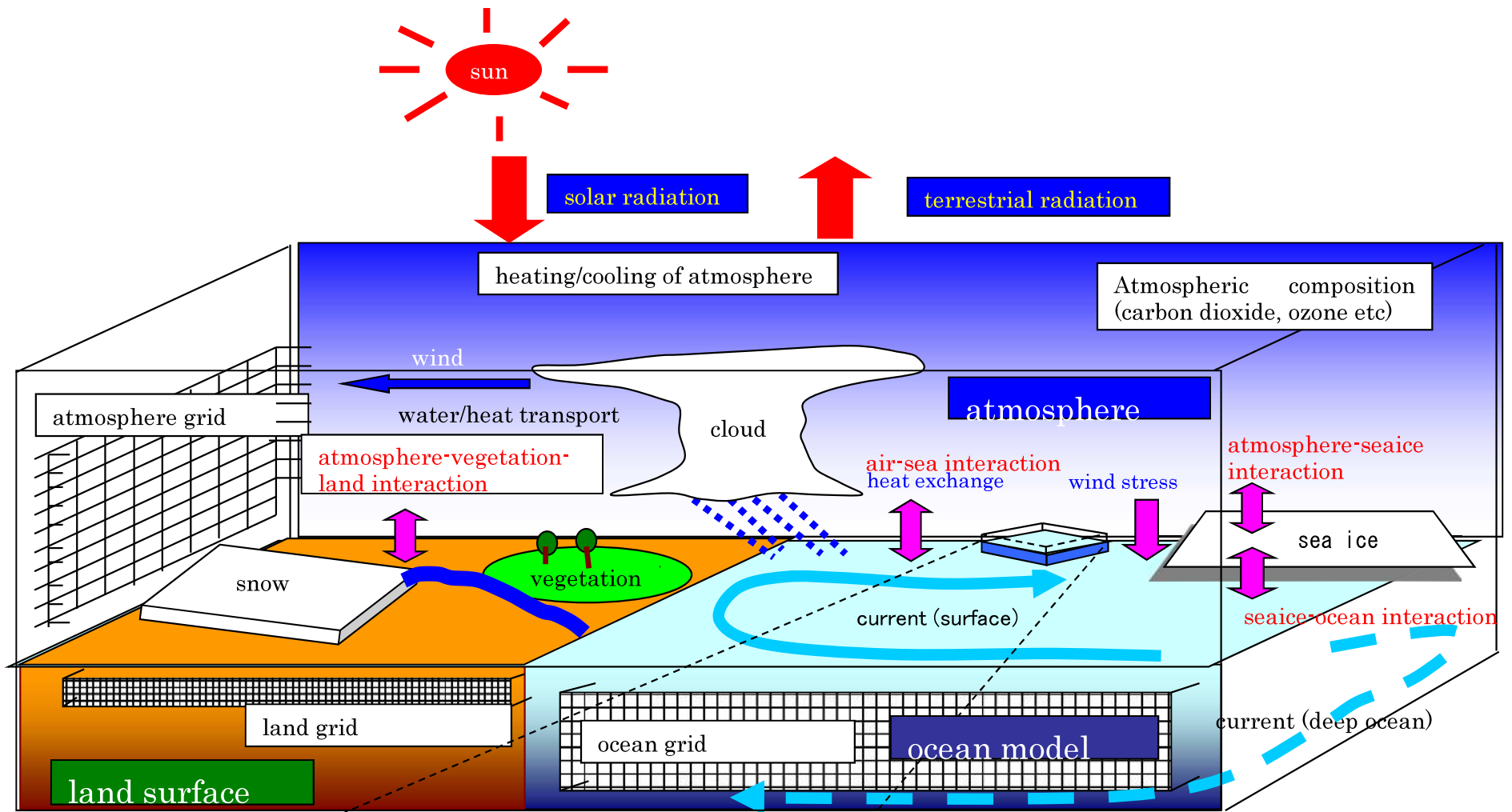Month of Maximum Temperature   CRU(1901–1998)

month
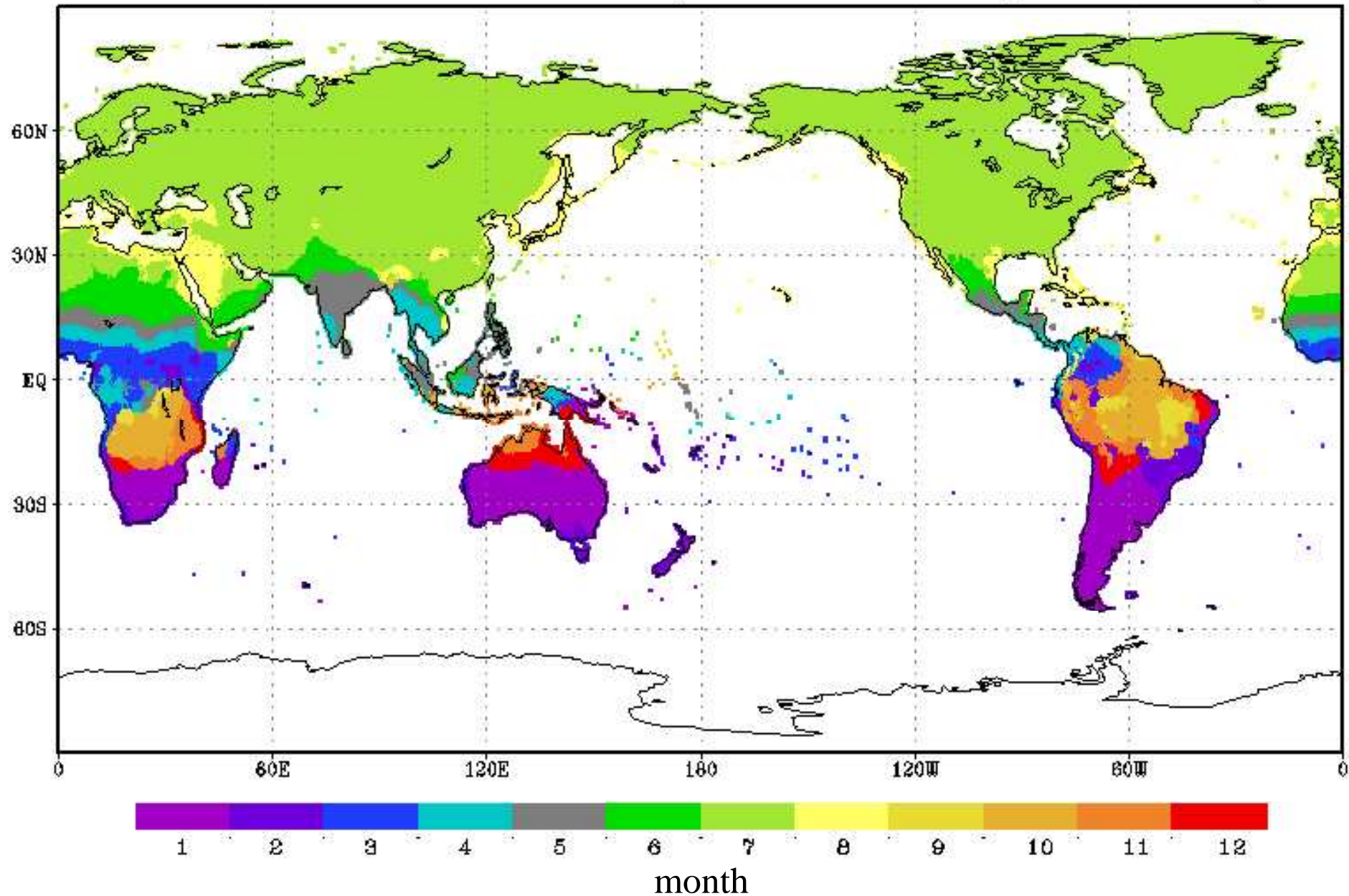
Koppen climate classification

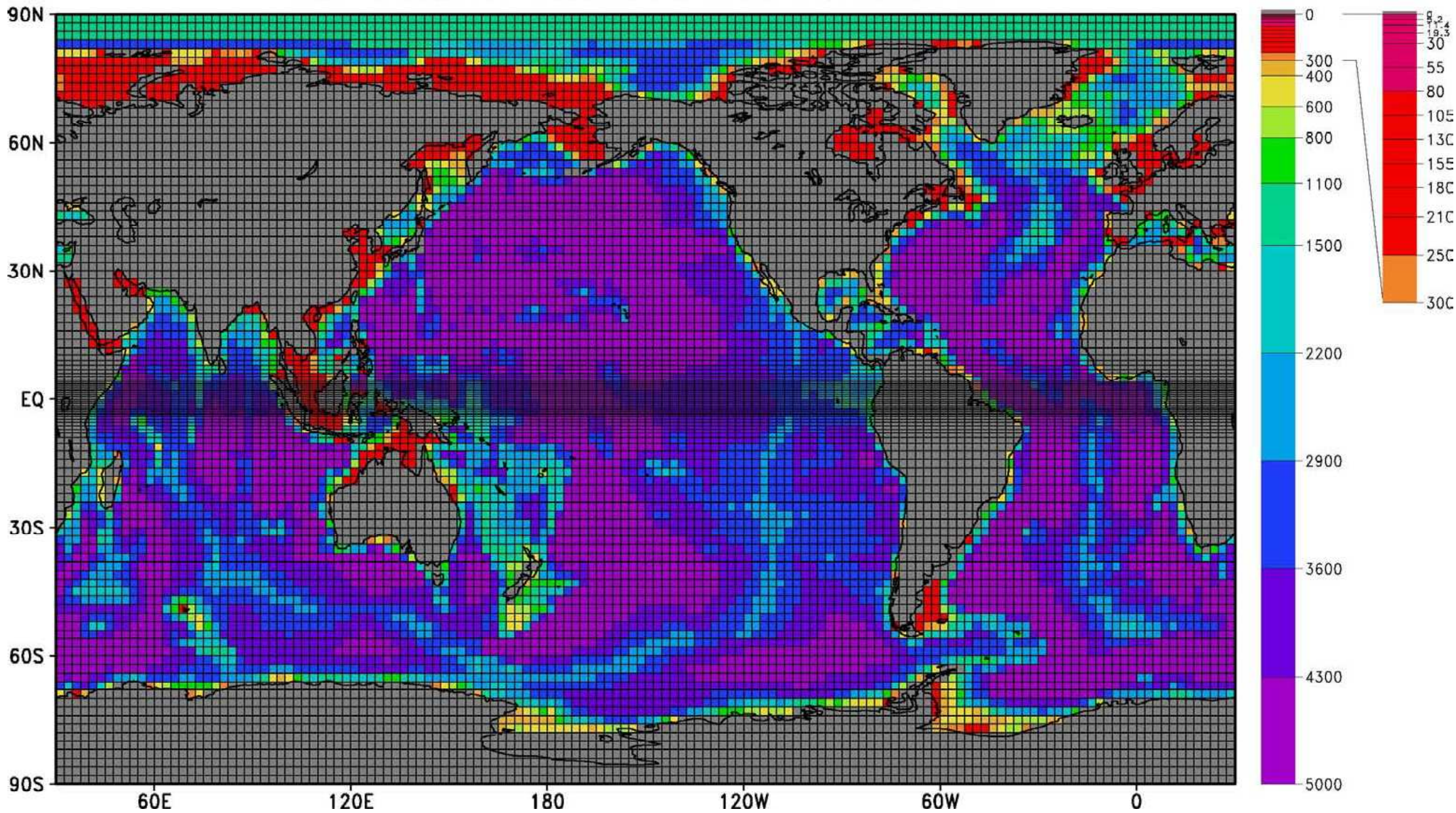Observation
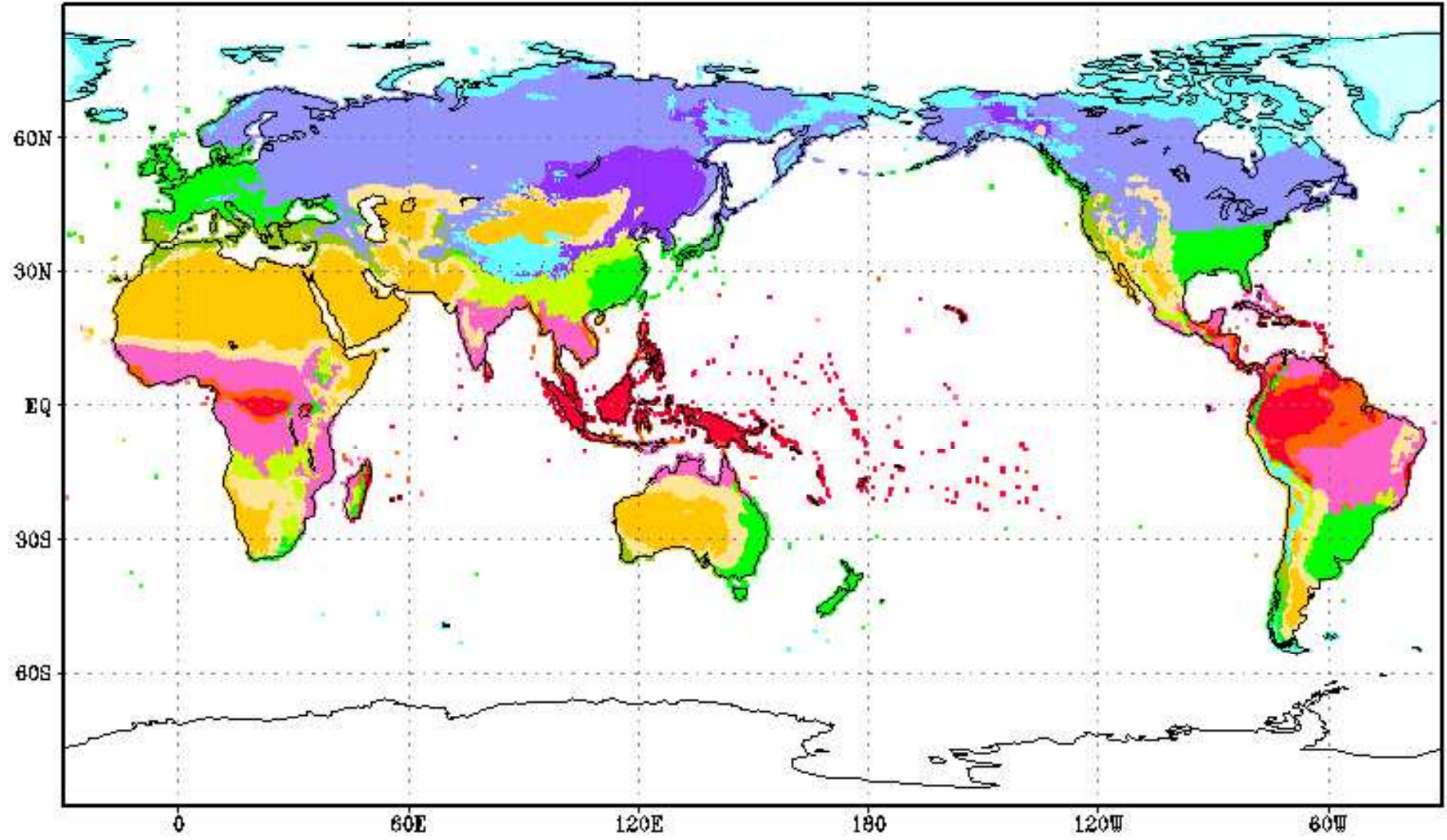
Climate Modeling
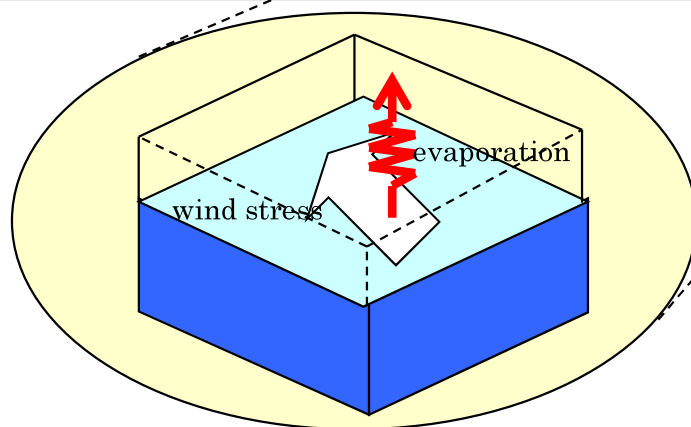
Schematic of climate model

MRI Coupled Atmosphere-Ocean General Circulation Model (MRI-CGCM2)

- AGCM
  - MRI/JMA98
  - T42 (2.8x2.8), L30 (top at 0.4 hPa)
  - Longwave radiation - Shibata and Aoki (1989)
  - Shortwave radiation - Shibata and Uchiyama (1992)
  - Cumulus - Prognostic Arakawa-Schubert type
  - PBL - Mellor and Yamada level 2 (1974)
  - Land Surface - L3SiB or MRI/JMA_SiB
- OGCM
  - Resolution : 2.5x(0.5-2.0), 23layers
  - Eddy mixing : Isopycnal mixing, GM
  - Seaice : Mellor and Kantha (1989)
- Coupling
  - Time interval : 24hours
  - Flux adjustment: used "with" or "without"
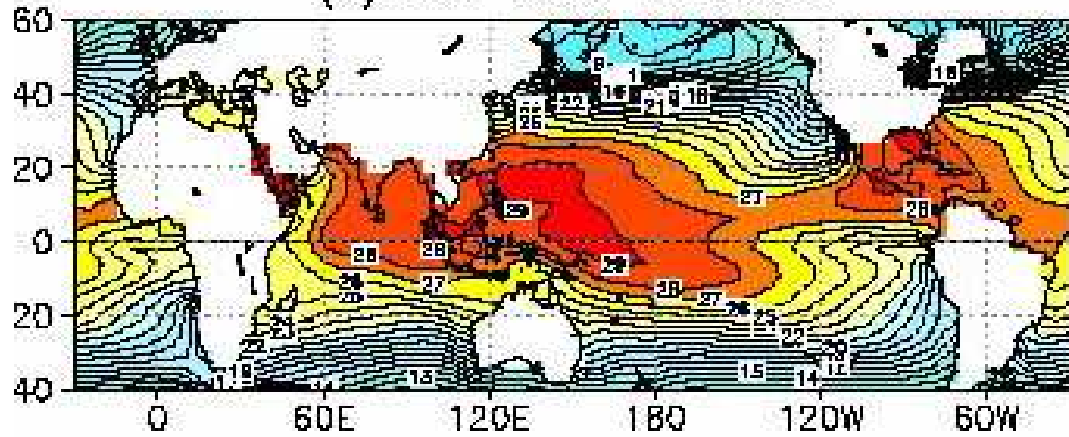
MRI-CGCM2.2 OCEAN BOTTOM TOPOGRAPHY

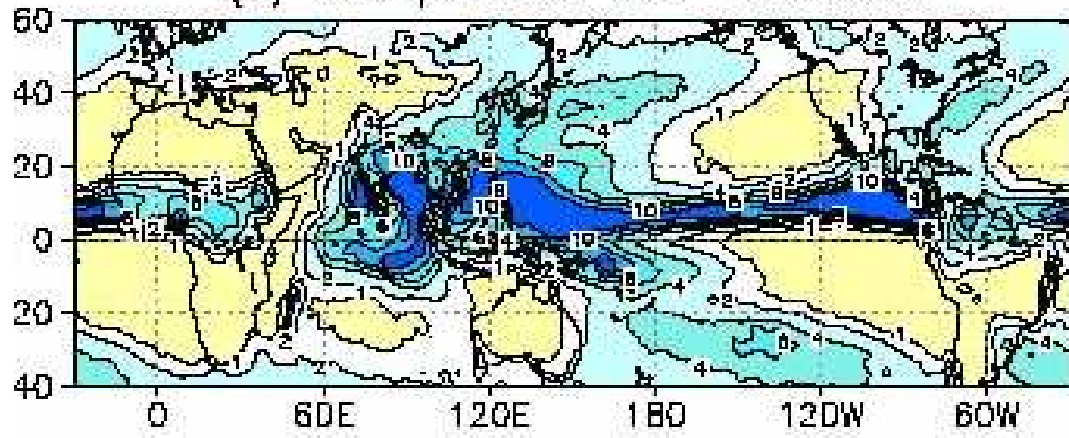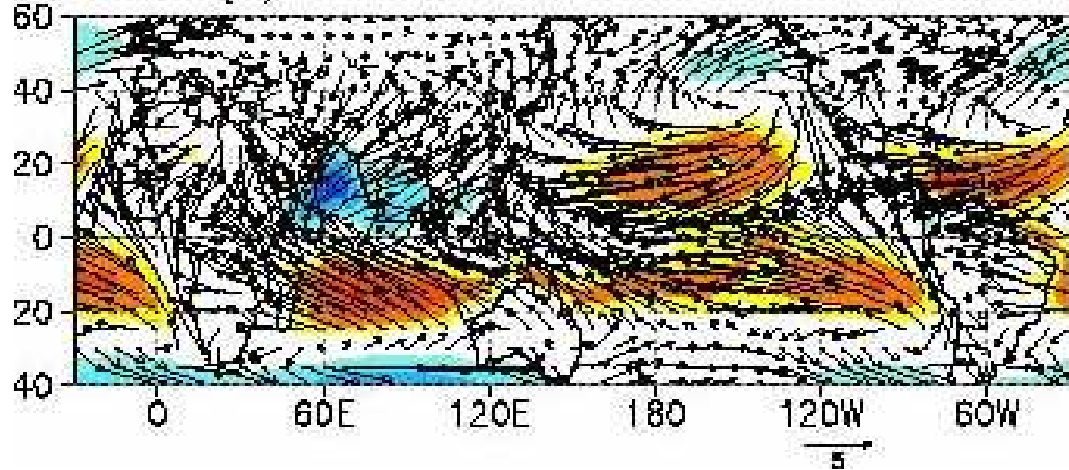# Climate Model Evaluation: Global Energy Balance

Obs
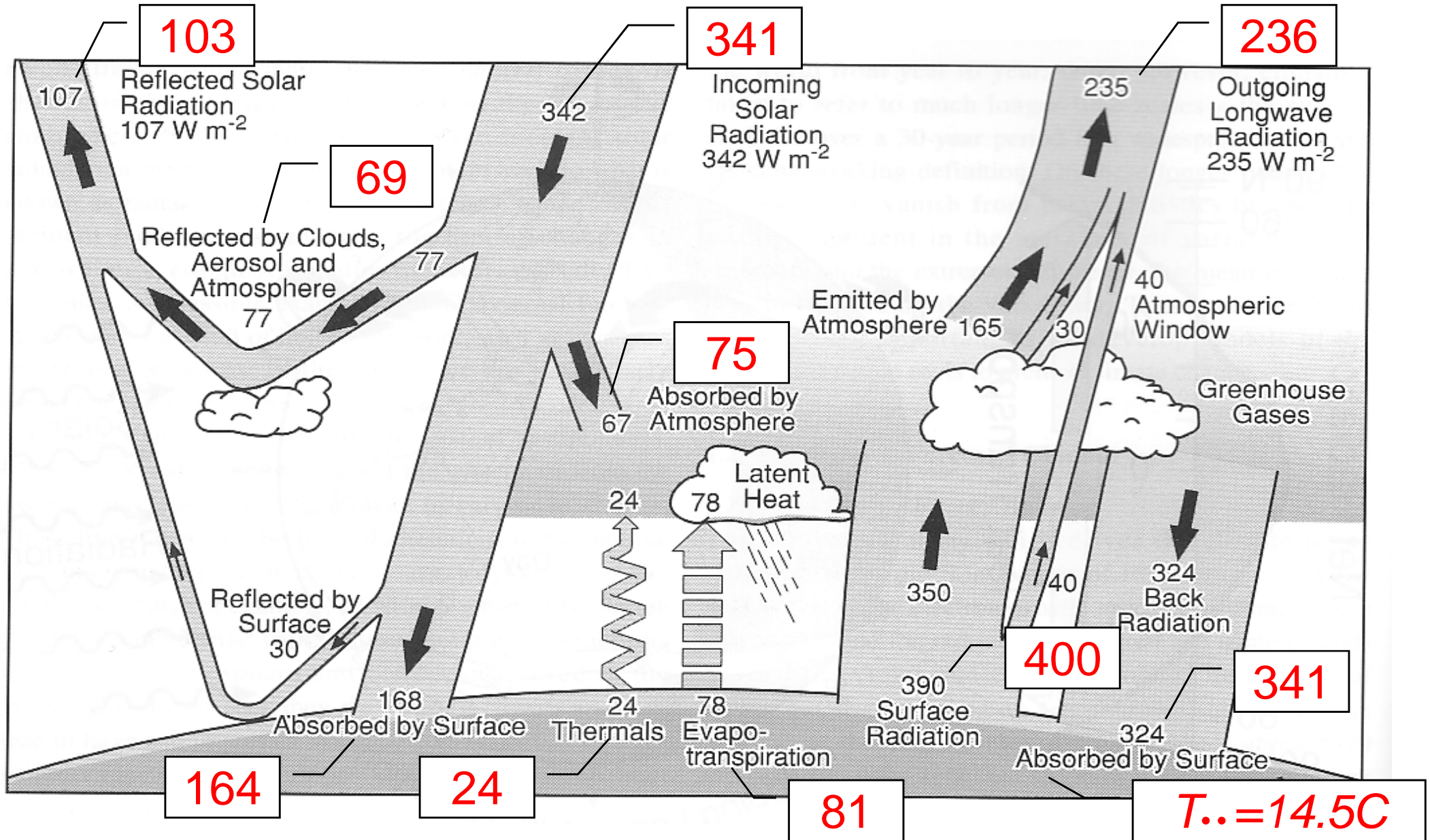
Model

(a) SST JJA Observed

(b) Precipitation JJA Observed
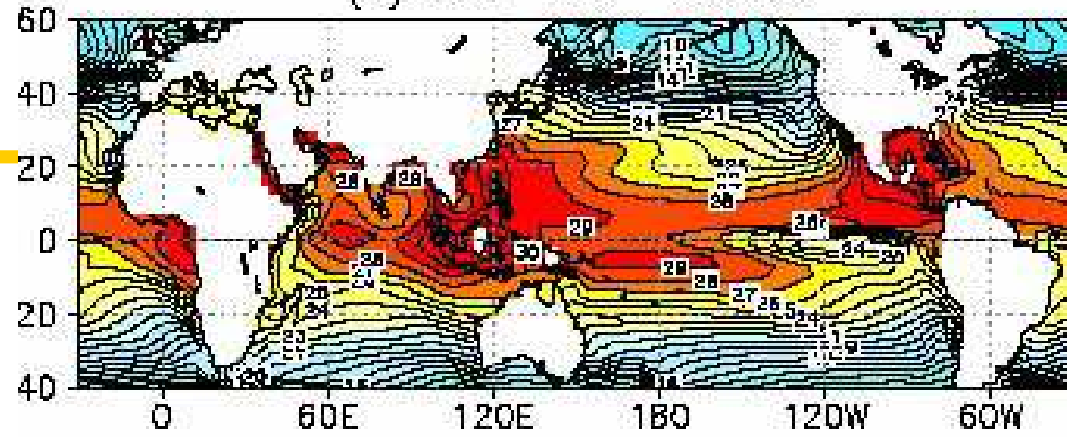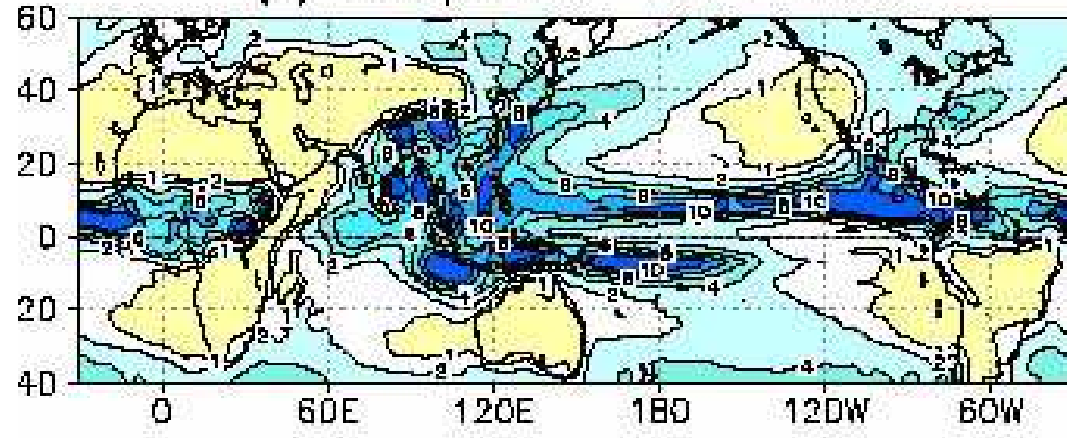
(c) Surface Wind JJA Observed

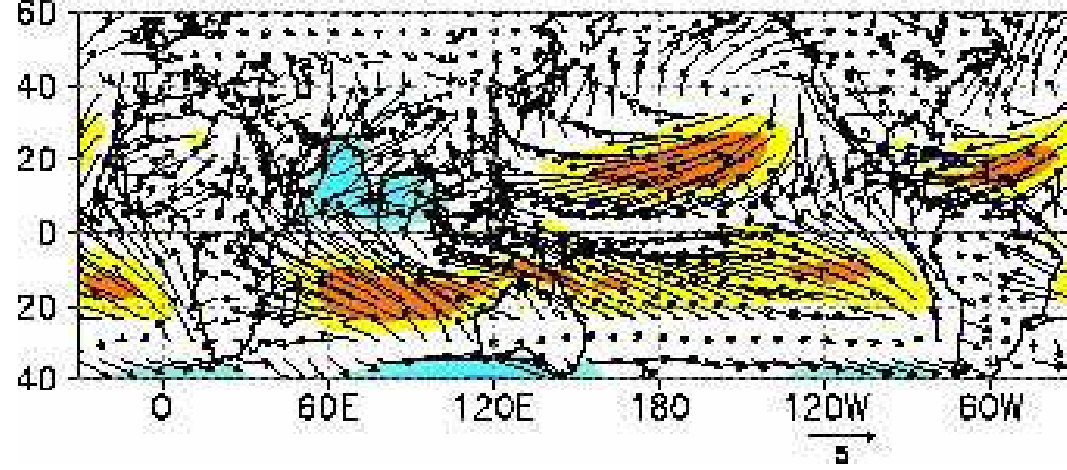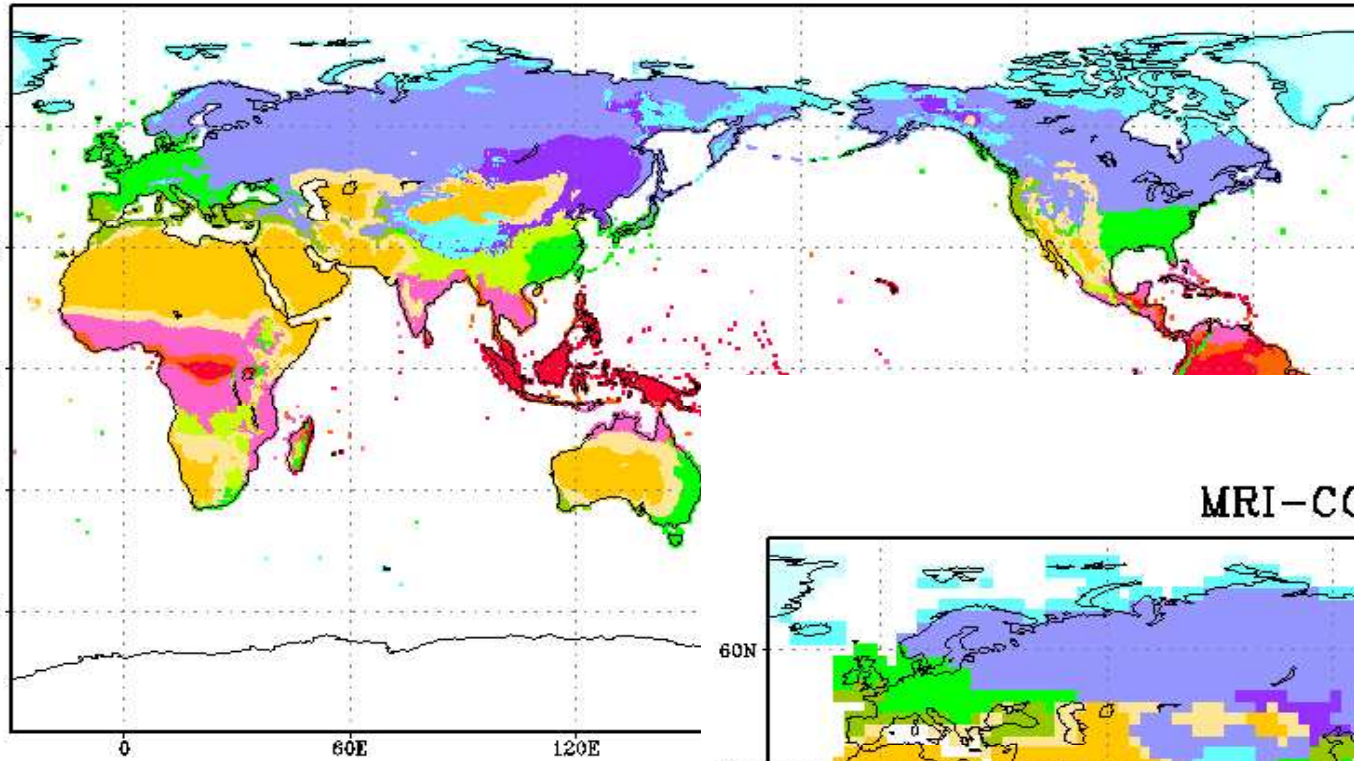(a) SST JJA CGCM2
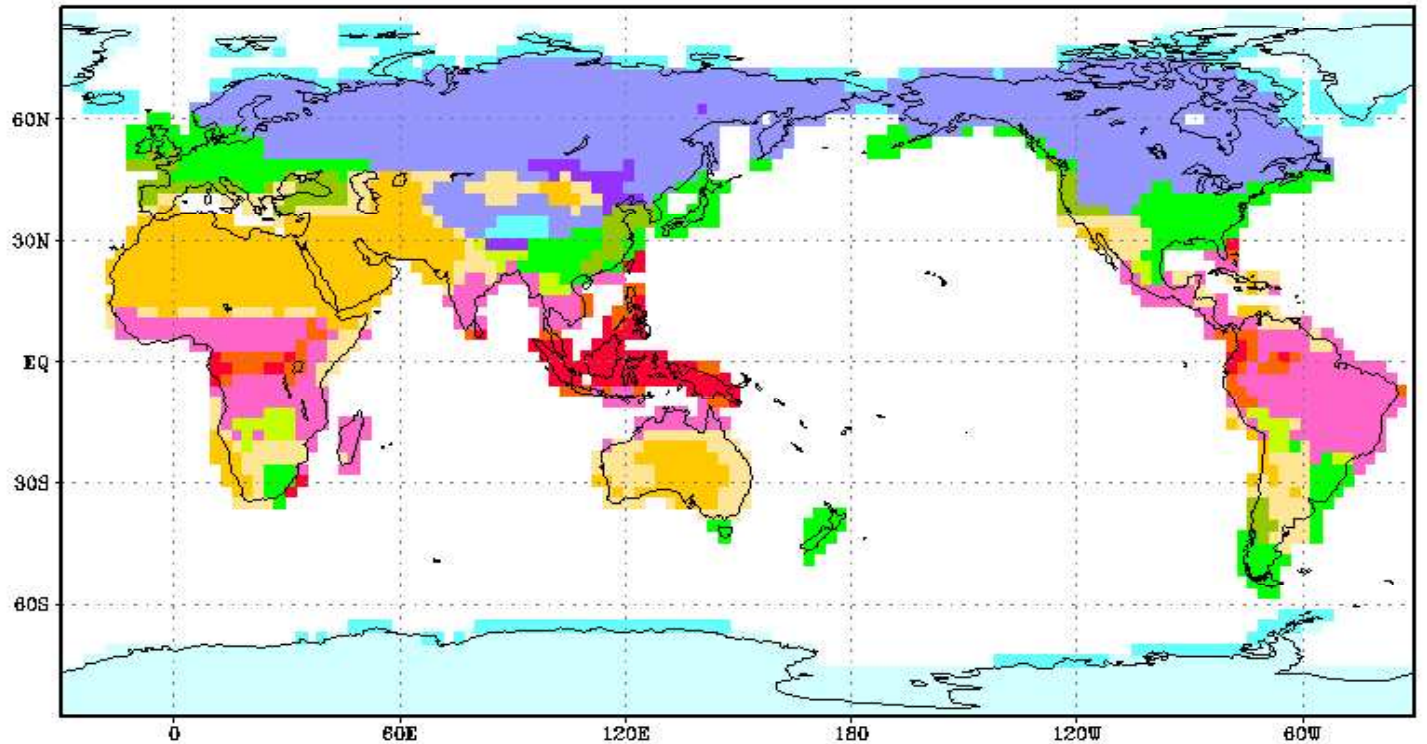
(b) Precipitation JJA CGCM2

(c) Surface Wind JJA CGCM2

Koppen climate

Observation

MRI-CGCM2.2 control

Af Am Aw BS BW Cs

Af Am Aw BS BW Cs Cw Cf Df Dw ET EF

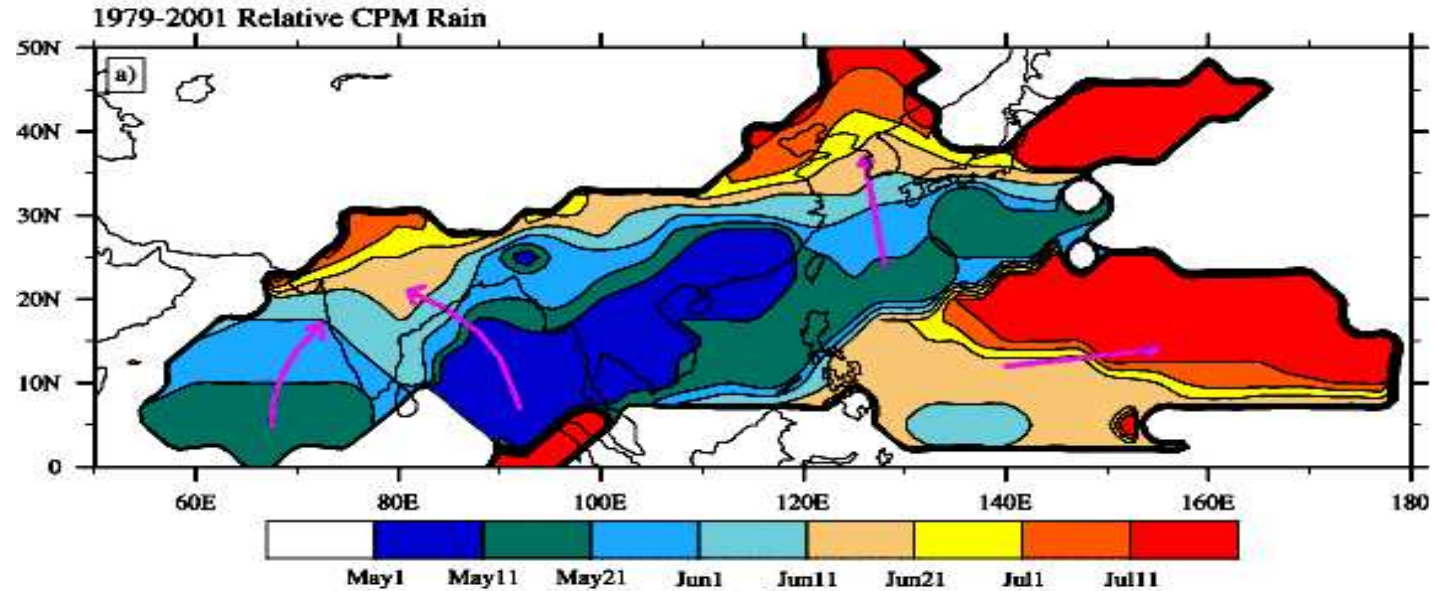# Onset Pentad: the Julian pentad in which the relative climatological pentad mean rainfall rate exceeds 4 mm/day.

Indian region

• **Northeastward progression over AS and the northwestward progression over the Bay of Bengal are well reproduced.**
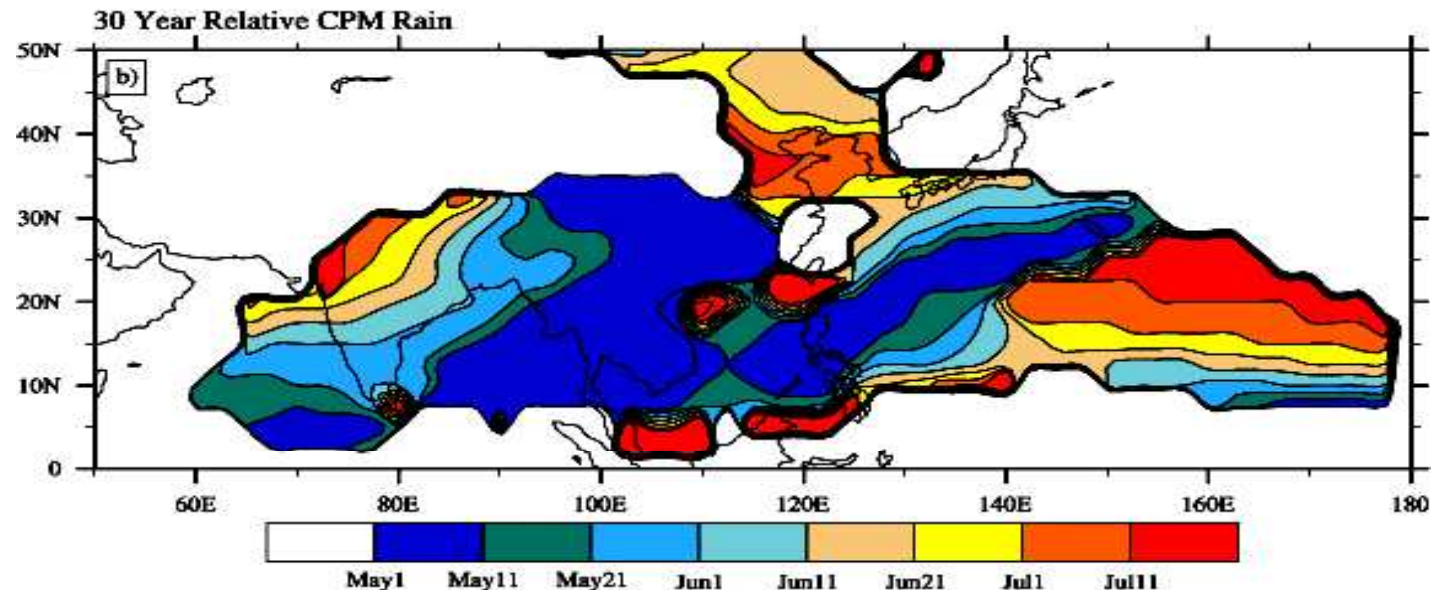
East Asia

• **The model simulates earlier monsoon onset over southeast Asia.**

• **Onset over Indochina in early May, the mid-May onset over the SCS & later northward progression due to Meiyu/Baiu rainband are all simulated, although the precise timings differ slightly.**

• **In northern China, onset is earlier and precipitation is heavier.**

Monsoon Onset Date
Xie-Arkin Observation
1979-2001 Relative CPM Rain

MRI CGCM2.2.2 Simulation
30 Year Relative CPM Rain

May1 May11 May21 Jun1 Jun11 Jun21 Jul1 Jul11

Mean Evolution of Monsoon: Withdrawal

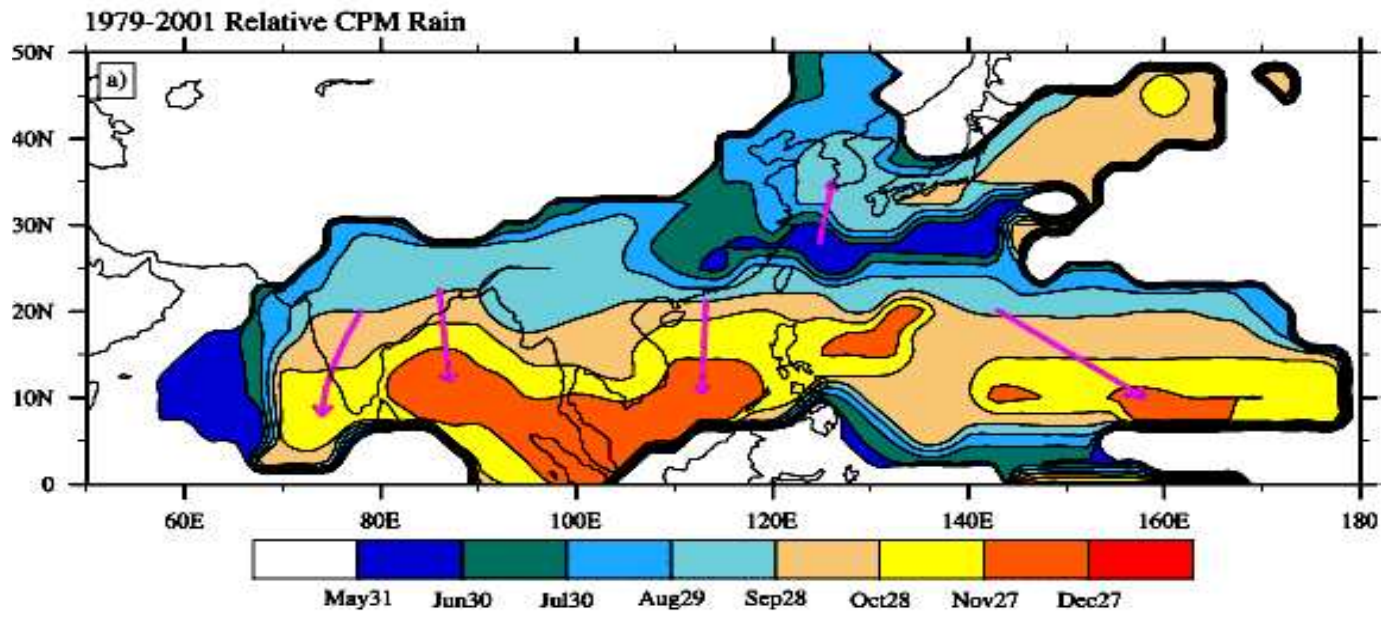Withdrawal Pentad: the transitional pentad in which rainfall drops below 4 mm/day.

Observation shows:

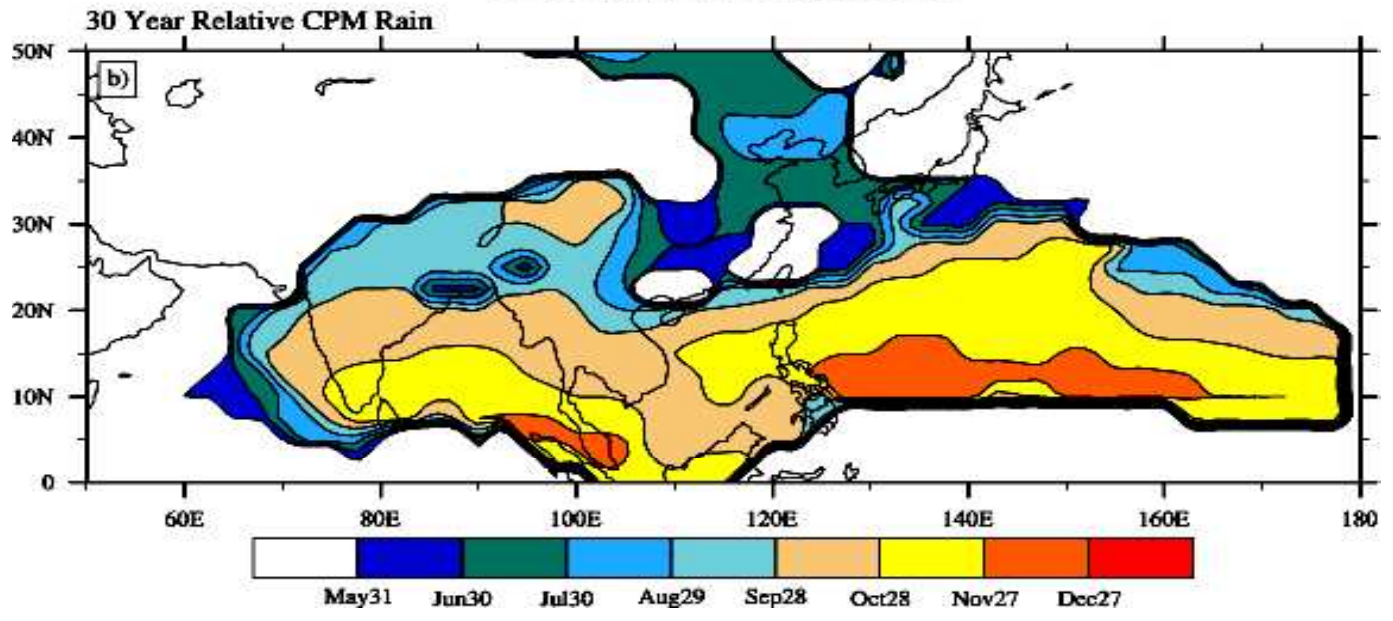➤ southward retreat of monsoon over India, southeast Asia and Western north Pacific

➤ northward retreat over East Asia.

Simulation close to observation.

Withdrawal Date of Rainy Season
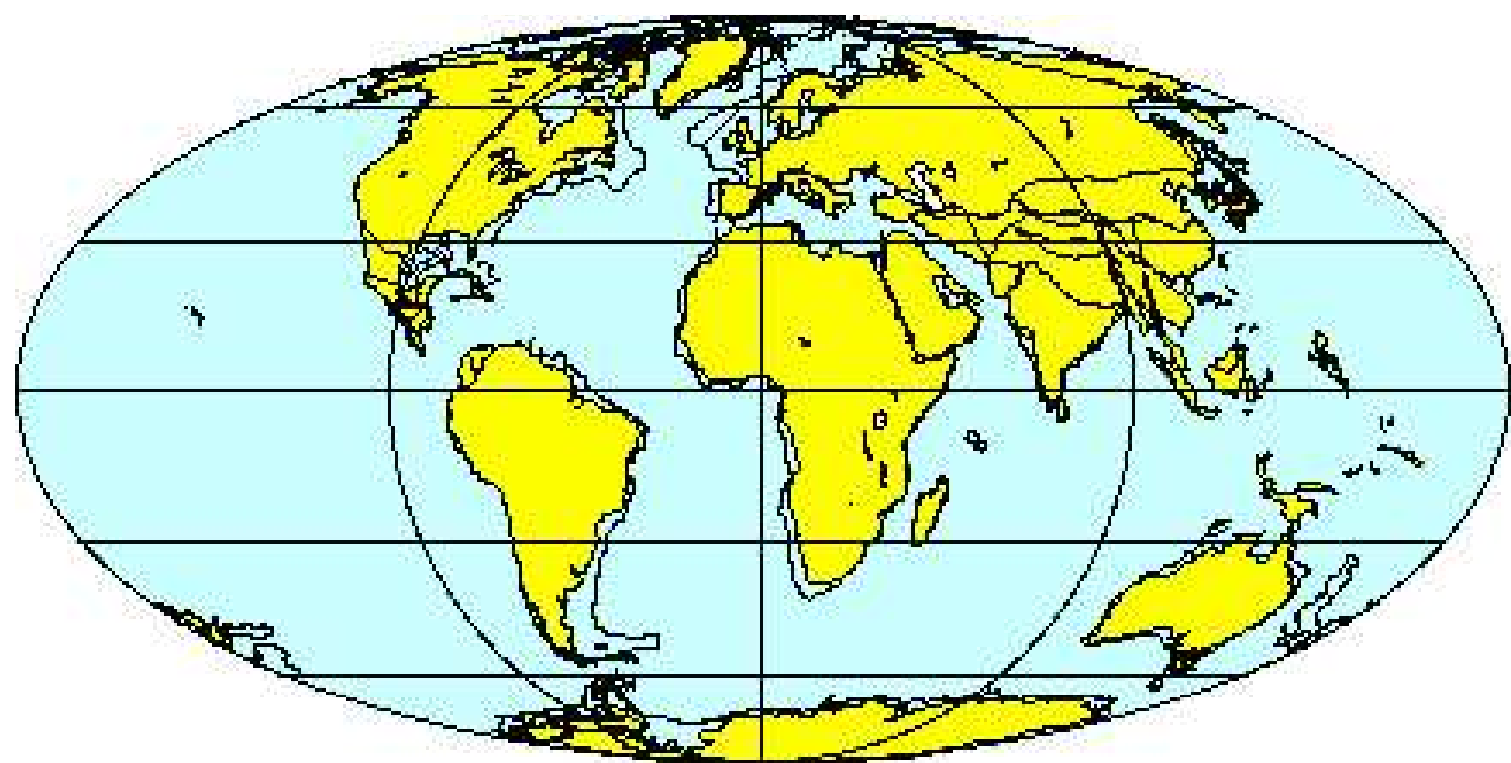
Xie-Arkin Observation

1979-2001 Relative CPM Rain

MRI CGCM2.2.2 Simulation

30 Year Relative CPM Rain

Role of orography on climate

Plate Reconstruction

30 Ma

65 Ma

http://www.ig.utexas.edu/research/projects/
plates/plates.htm

a 30 Myr ago

b 10 Myr ago

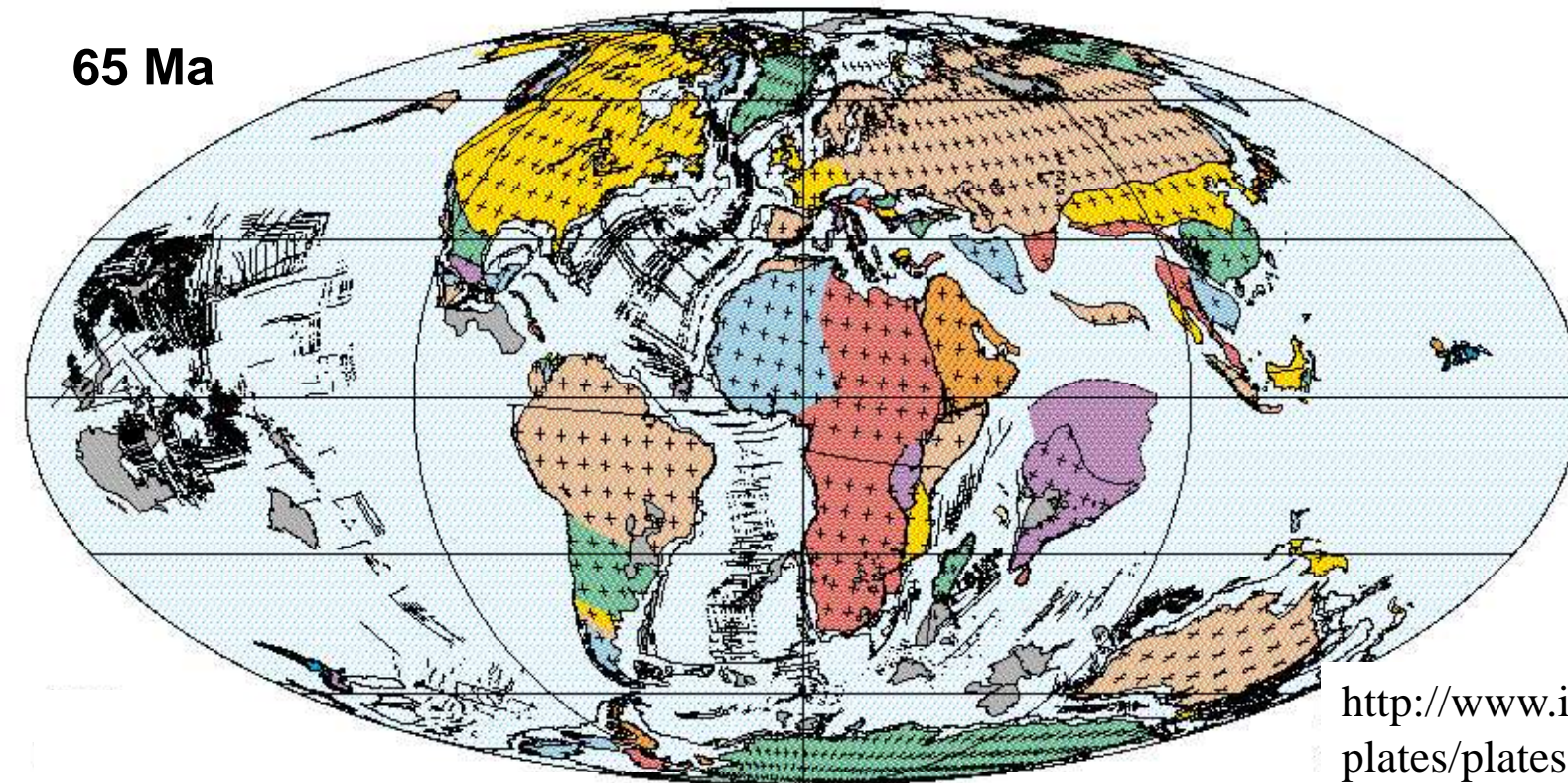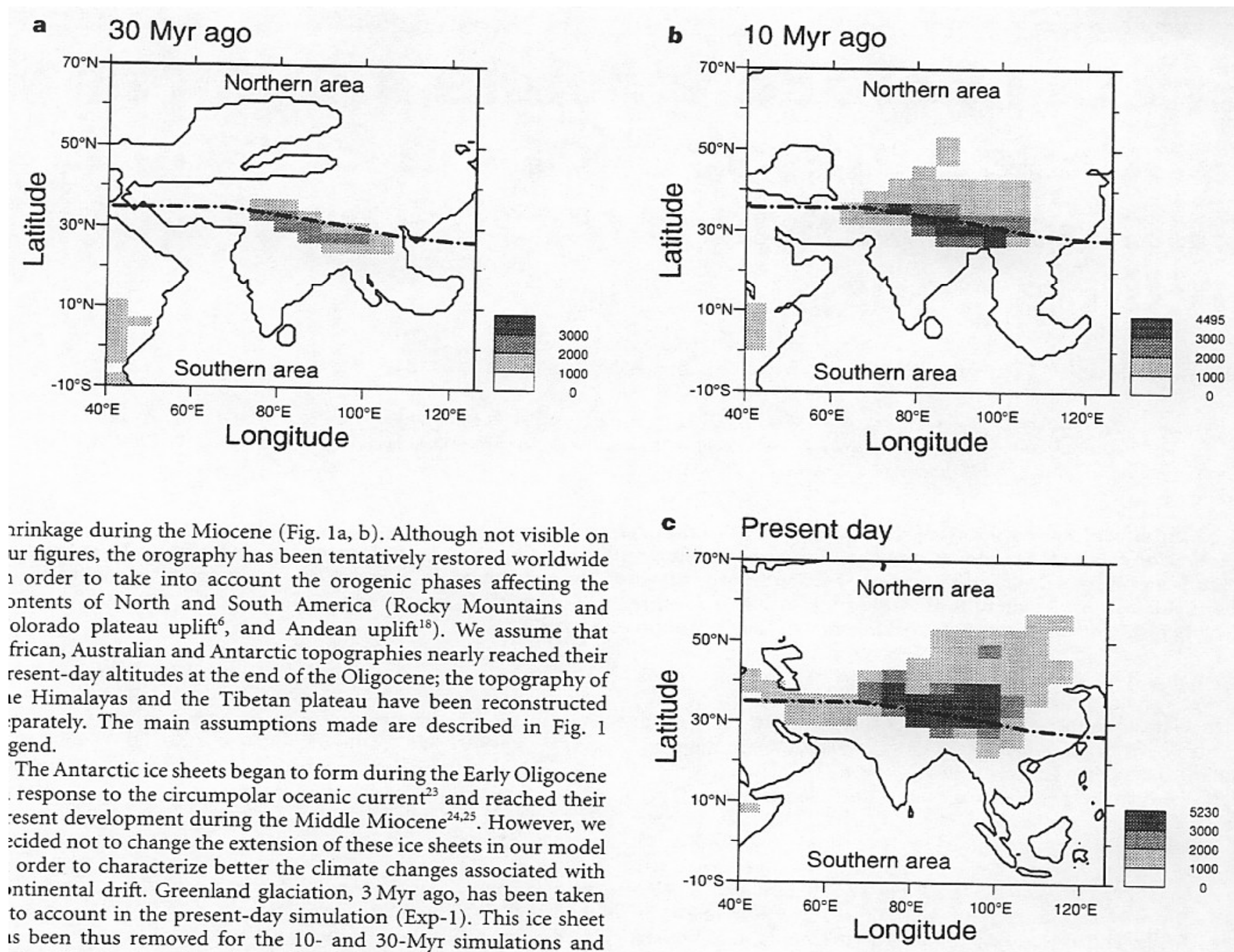**c** Present day

nrinkage during the Miocene (Fig. 1a, b). Although not visible on ur figures, the orography has been tentatively restored worldwide n order to take into account the orogenic phases affecting the ontents of North and South America (Rocky Mountains and olorado plateau uplift[6], and Andean uplift[18]). We assume that frican, Australian and Antarctic topographies nearly reached their resent-day altitudes at the end of the Oligocene; the topography of e Himalayas and the Tibetan plateau have been reconstructed eparately. The main assumptions made are described in Fig. 1 gend.

The Antarctic ice sheets began to form during the Early Oligocene . response to the circumpolar oceanic current[23] and reached their resent development during the Middle Miocene[24,25]. However, we ecided not to change the extension of these ice sheets in our model order to characterize better the climate changes associated with ntinental drift. Greenland glaciation, 3 Myr ago, has been taken to account in the present-day simulation (Exp-1). This ice sheet us been thus removed for the 10- and 30-Myr simulations and

Ramstein et al. (1997) Nature

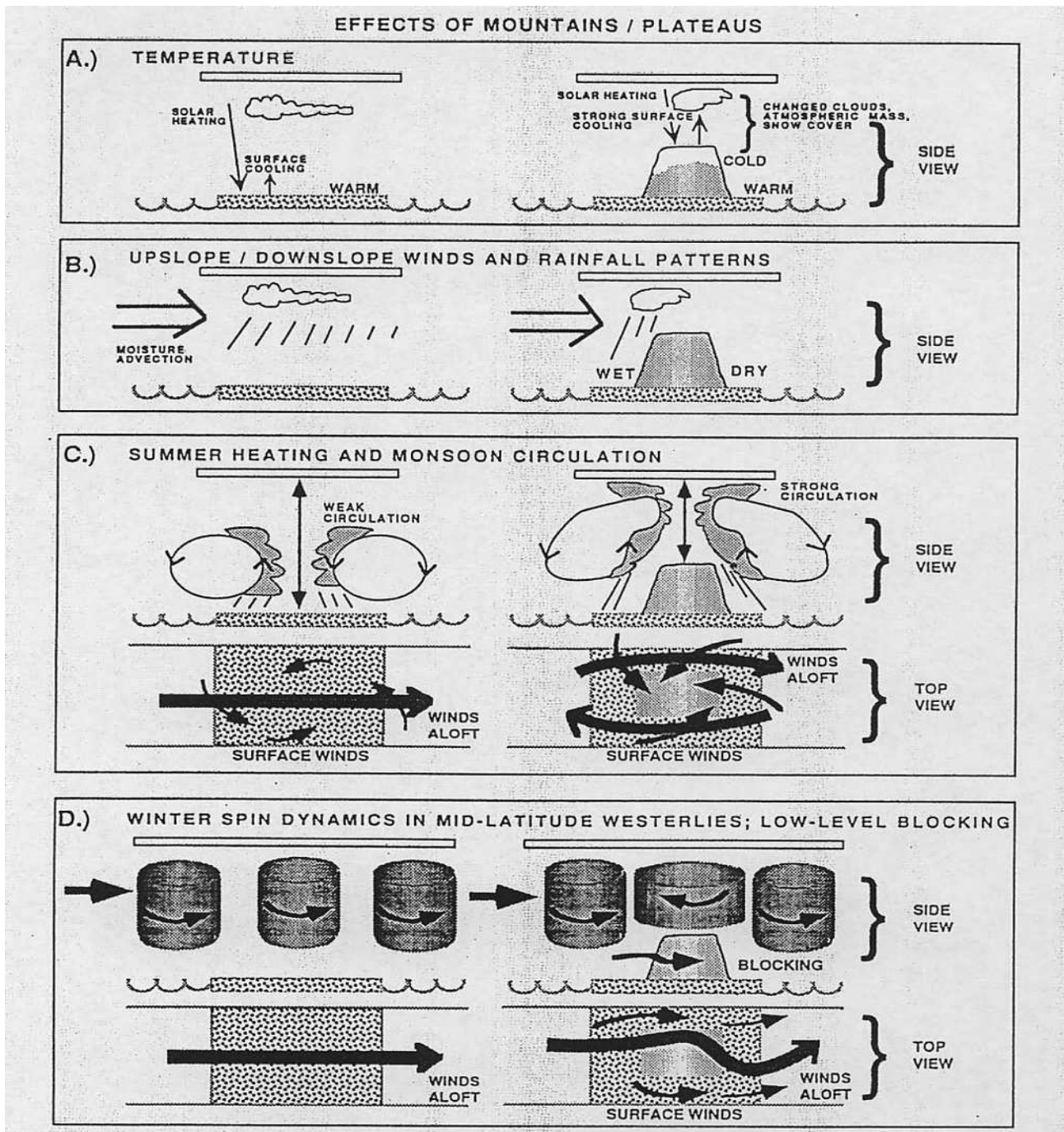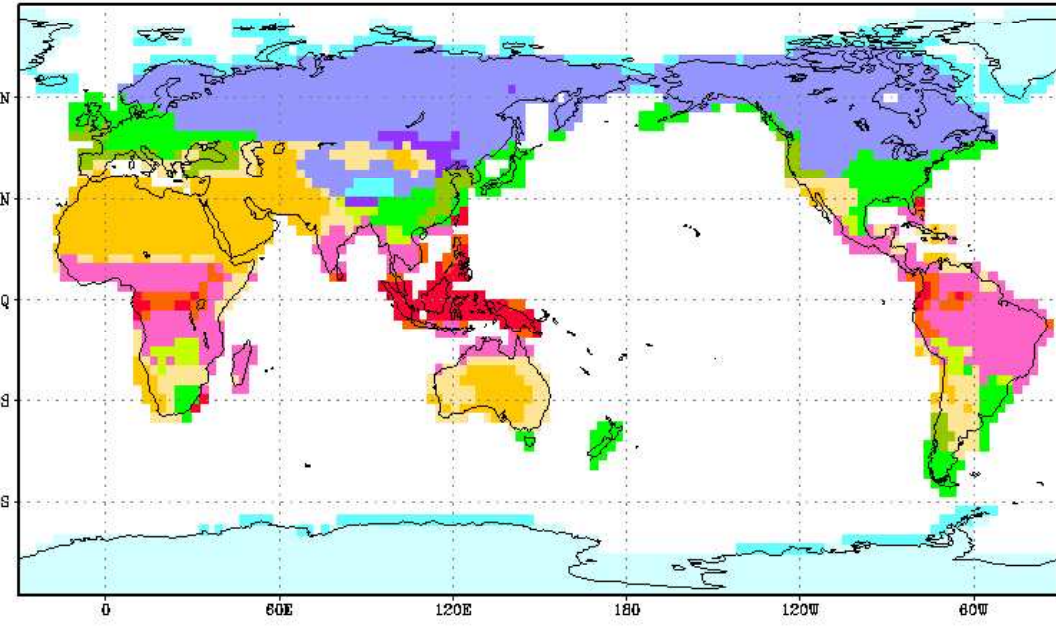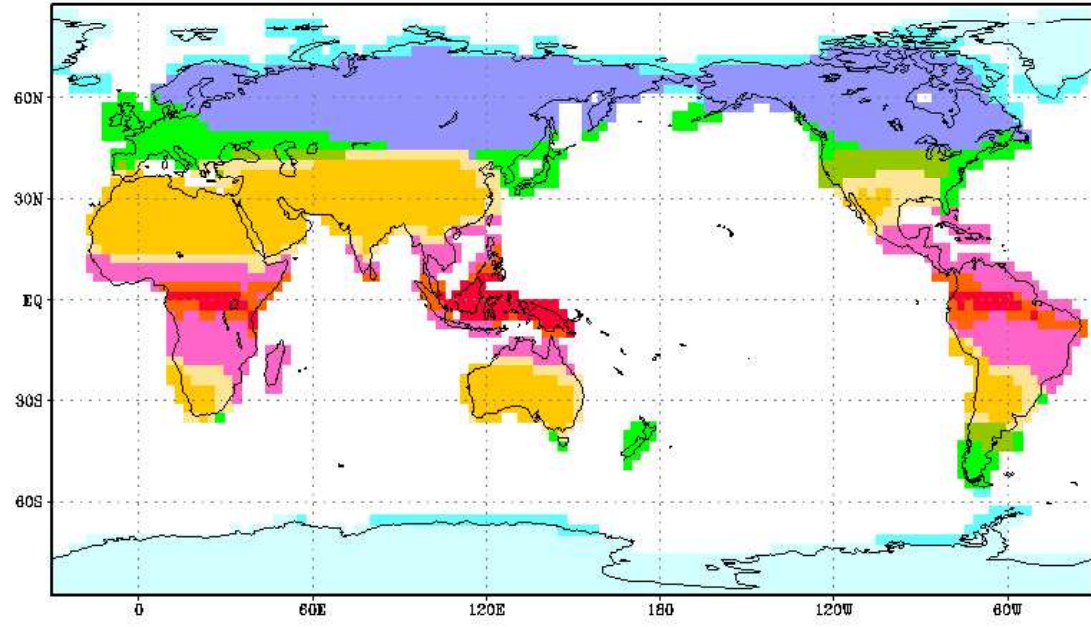# Effect of mountains on climate

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

Paleo climate

Orbital parameters
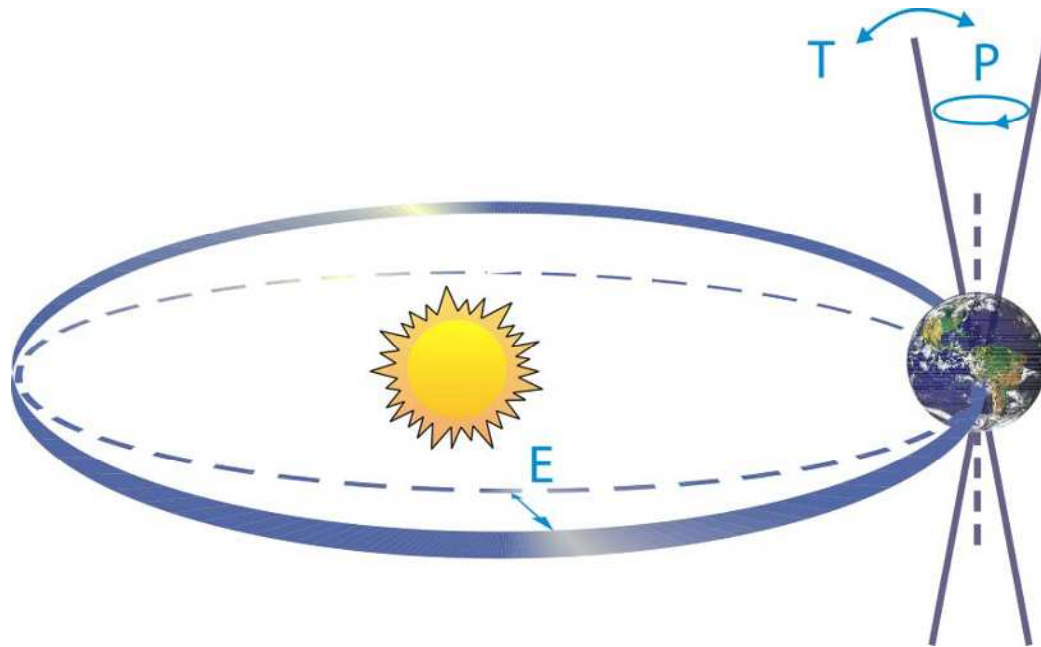
There are three fundamental ways the Earth's radiation balance can change, thereby causing a climate change:

(1) changing the incoming solar radiation (e.g., by changes in the Earth's orbit or in the Sun itself),

(2) changing the fraction of solar radiation that is reflected (this fraction is called the albedo – it can be changed, for example, by changes in cloud cover, small particles called aerosols or land cover), and

(3) altering the longwave energy radiated back to space (e.g., by changes in greenhouse gas concentrations).

(4) local climate also depends on how heat is distributed by winds and ocean currents.

Schematic of the Earth's orbital changes (Milankovitch cycles) that drive the ice age cycles. 'T' denotes changes in the tilt (or obliquity) of the Earth's axis, 'E' denotes changes in the eccentricity of the orbit (due to variations in the minor axis of the ellipse), and 'P' denotes precession, that is, changes in the direction of the axis tilt at a given point of the orbit. Source: Rahmstorf and Schellnhuber (2006).

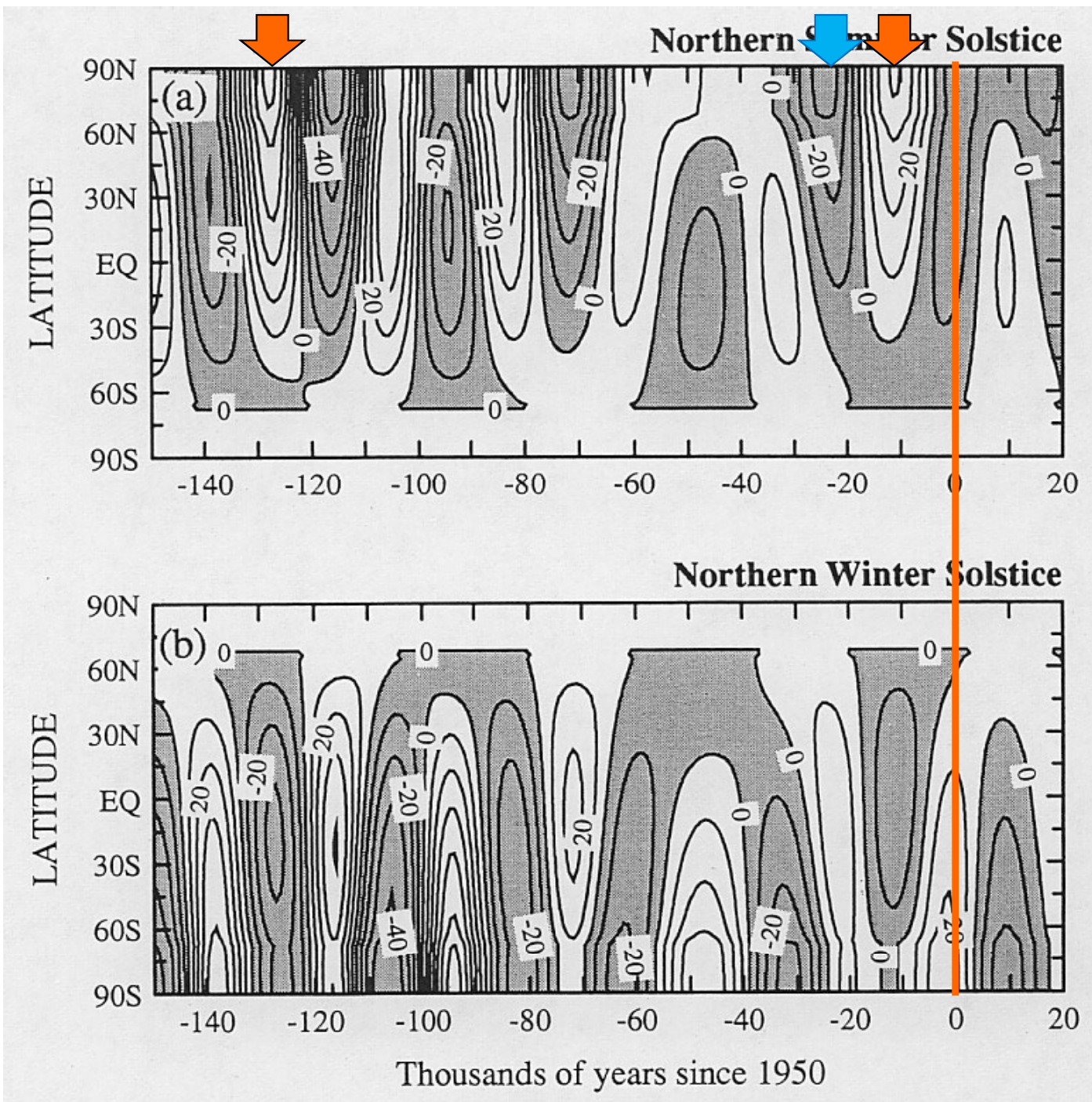# Eccentricity, Obliquity and Precession
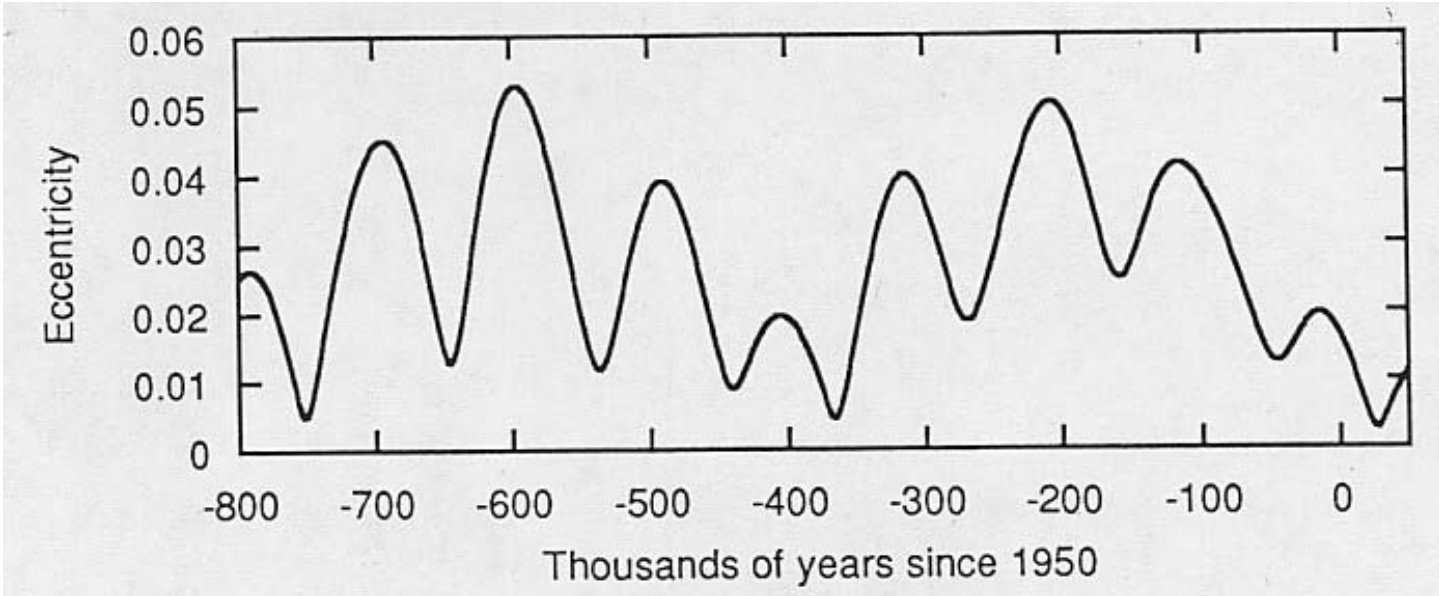
Eccentricity has a dominant period of 100,000 years
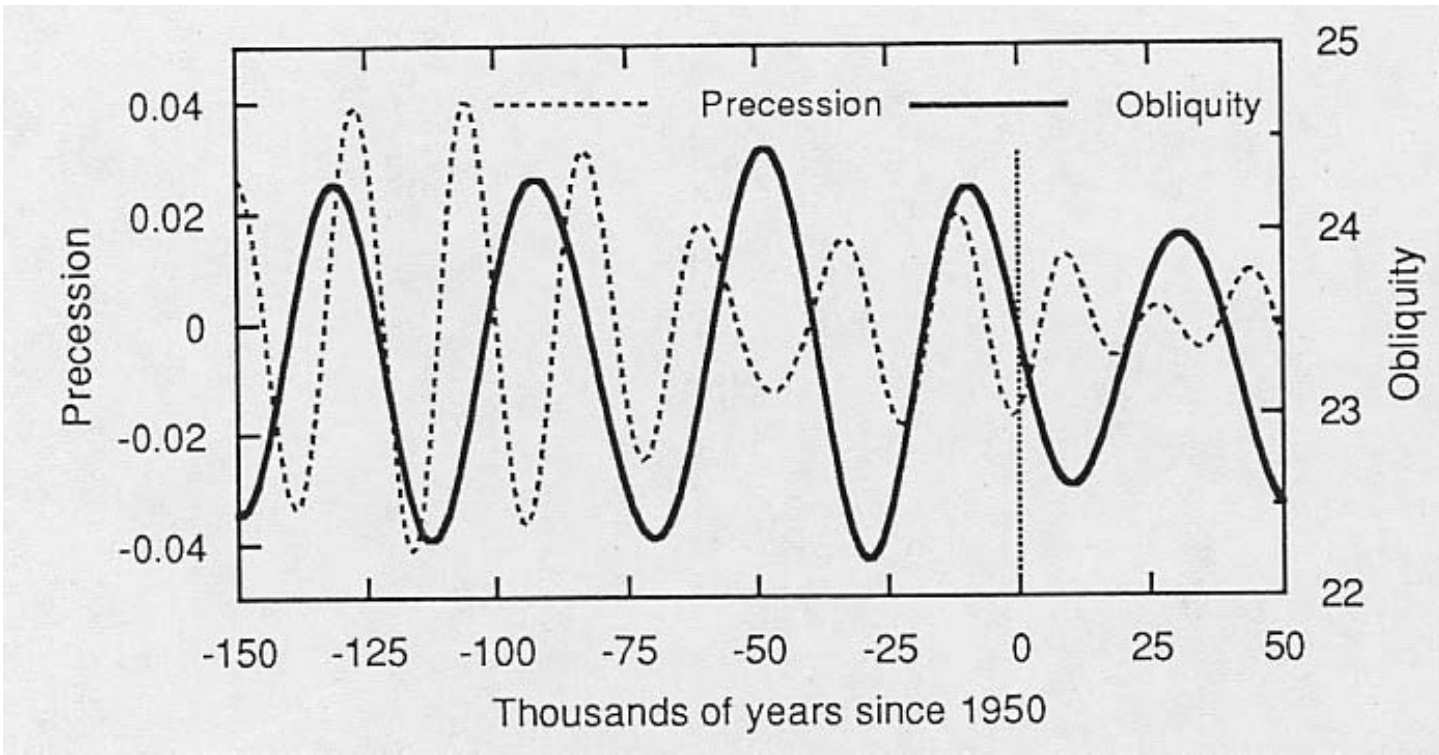
The obliquity and the precession have dominant periodicity of 40,000 years and 20,000 years, respectively

Hartmann (1994)

Insolation anomalies from 150,000 years ago until 20,000 years in the future

Insolation variations at NH summer sostice show amplitudes as large as 60 W m⁻² near the pole.

Large positive anomalies at 125,000 ya and 10,000 ya correspond fairly well with the times of the last two interglacial periods. The last glacial maximum about 20,000 ya was preceded by a relative minimum in NH summertime insolation.
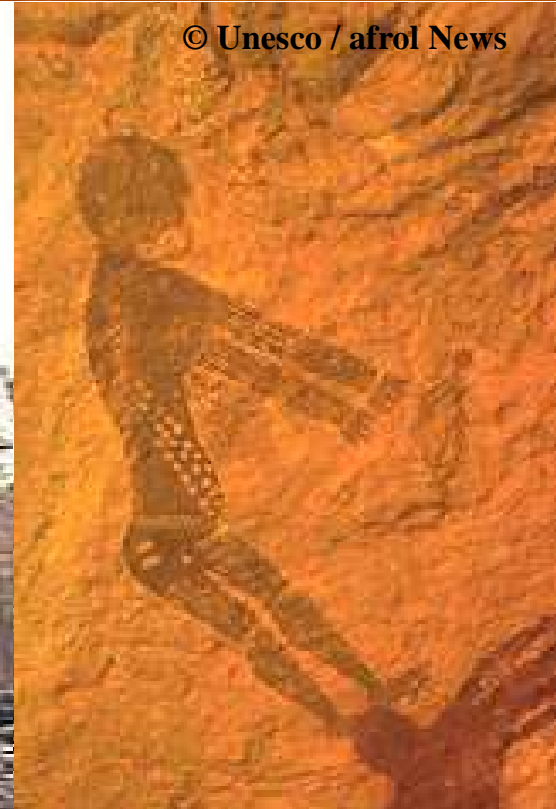
Hartmann (1994)

Dome Concordia (Antarctica): 740,000 years of climate change
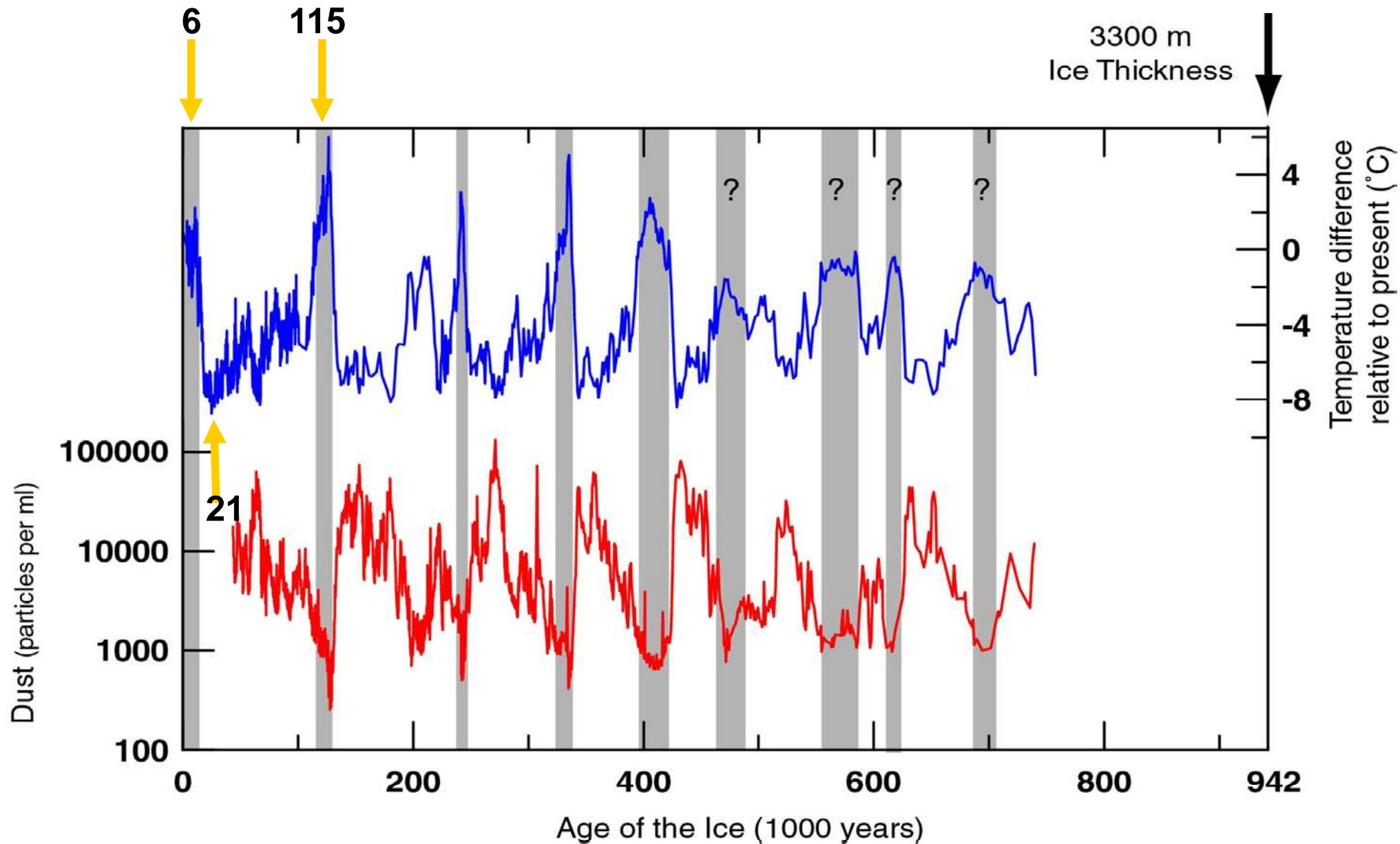
(EPICA Members, 2004)

Mid-Holocene: 6ka

Tassili n'Ajjer, Algeria
- Sahara was greener

© Unesco / afrol News

Last Glacial Maximum: 21ka
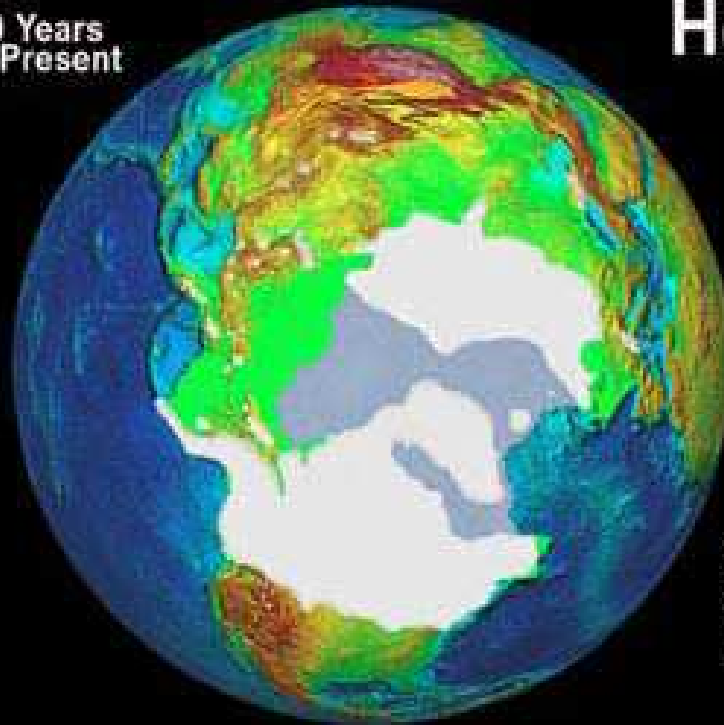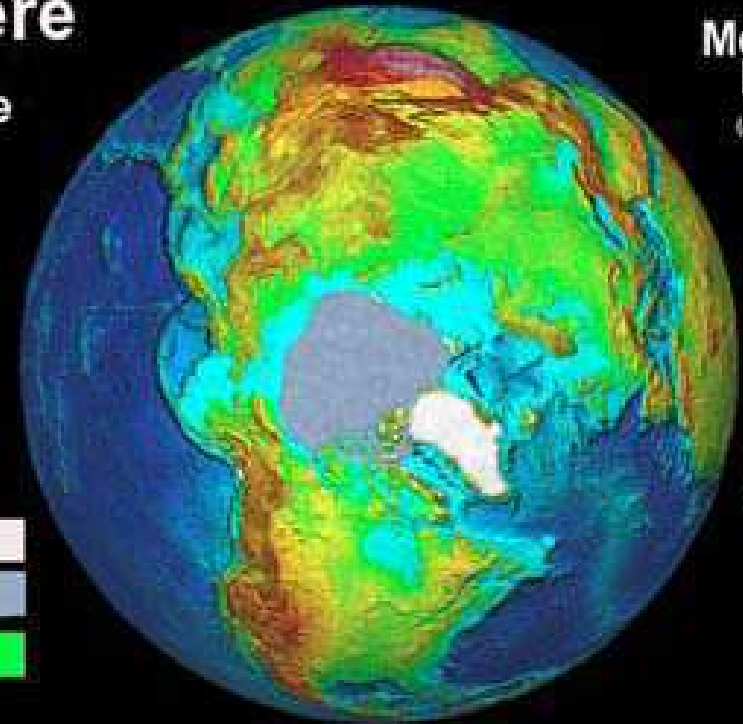
PMIP data: global lake status
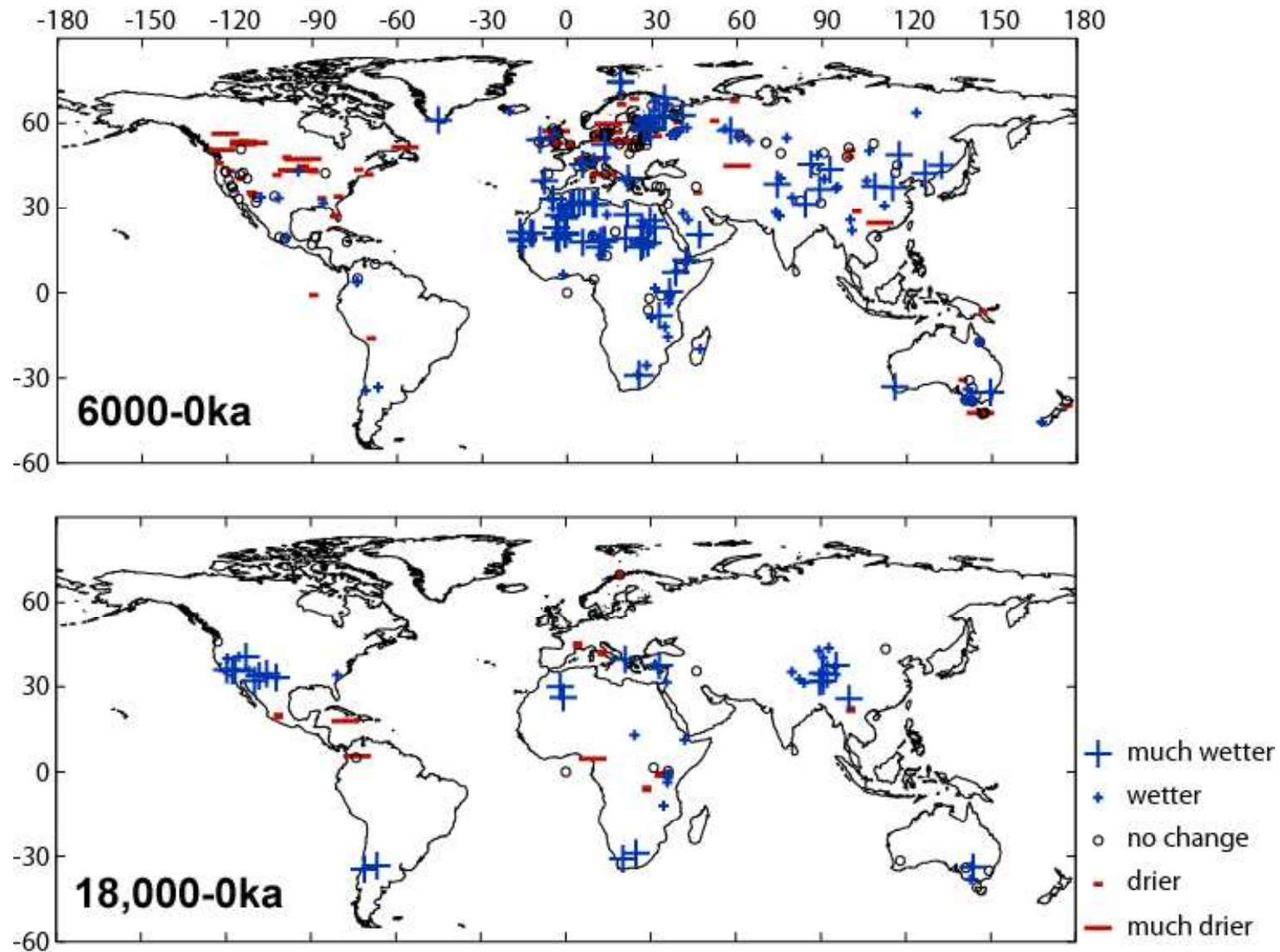
Jul-Aug-Sep Precipitation difference (6ka−0ka) with 4 CGCMs

Zhao et al. 2004