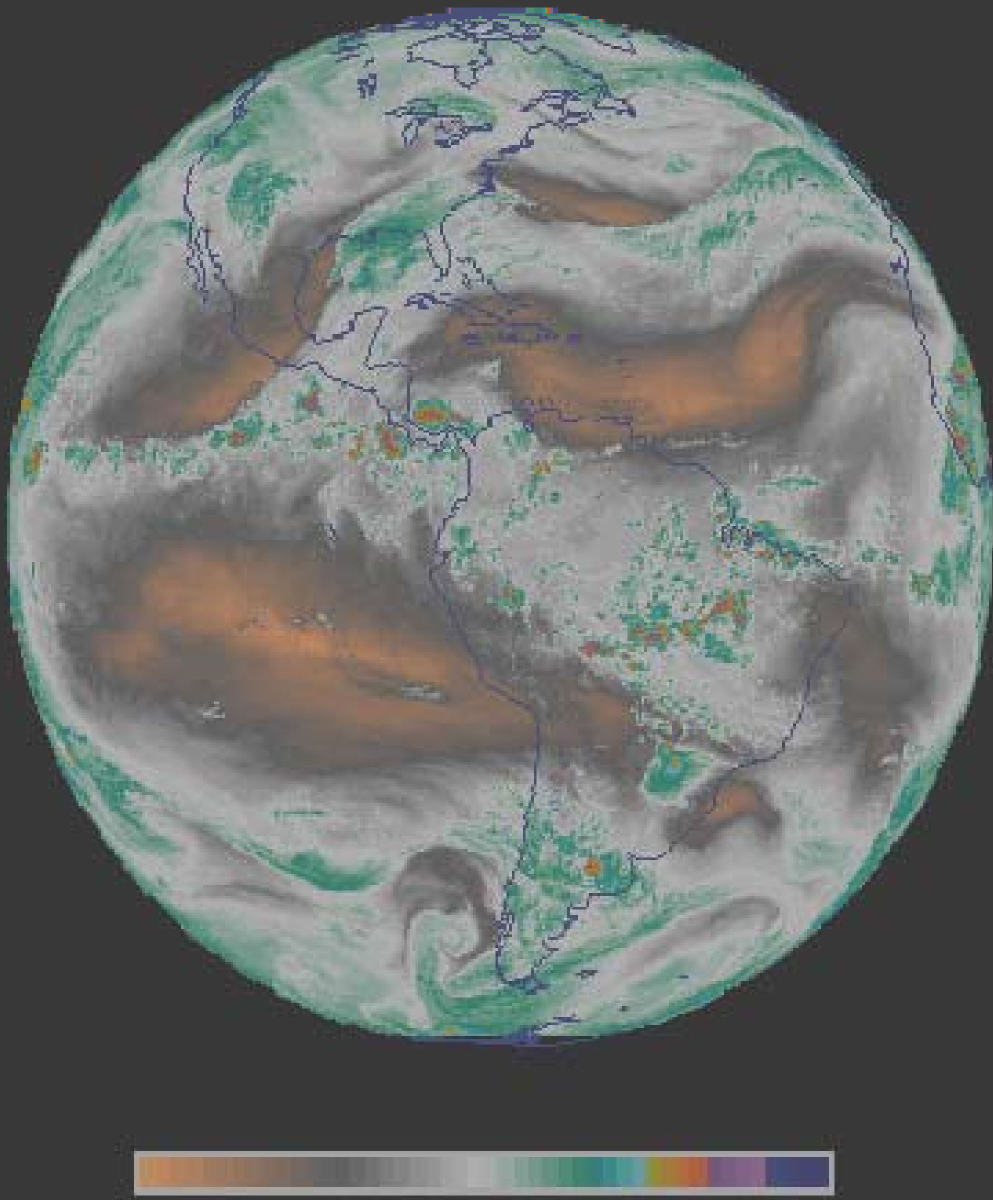
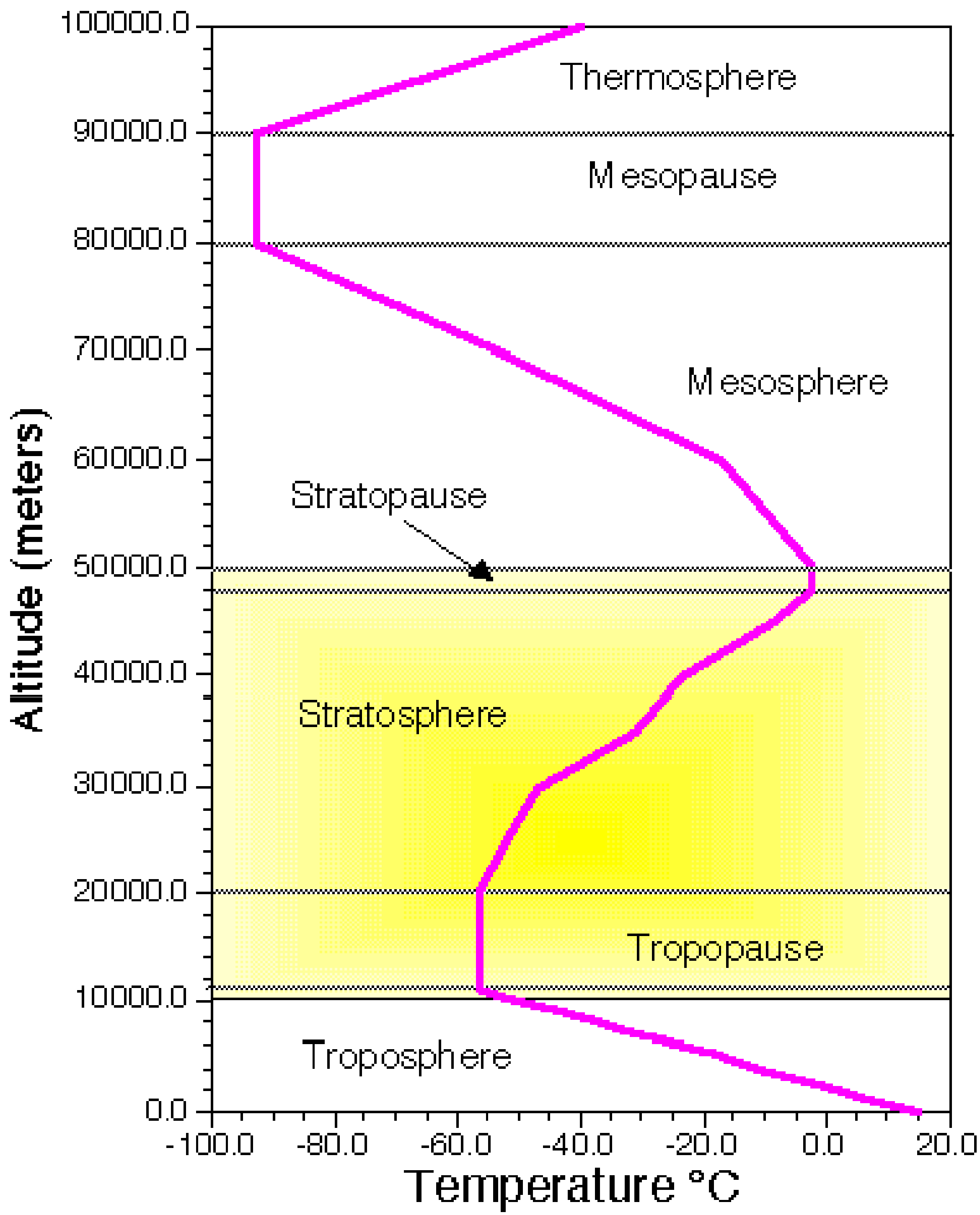
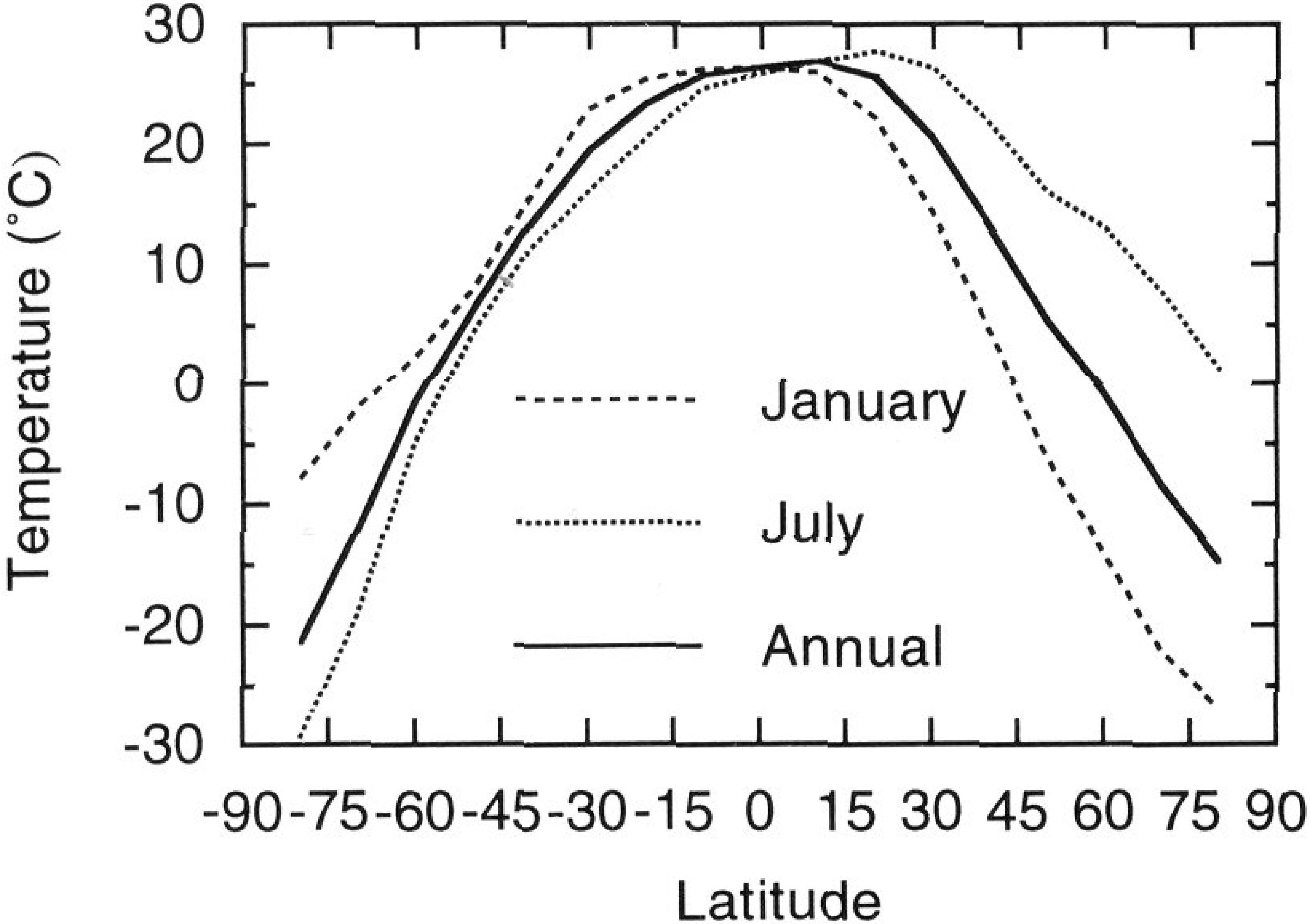
The Circulation of the Atmosphere and Oceans

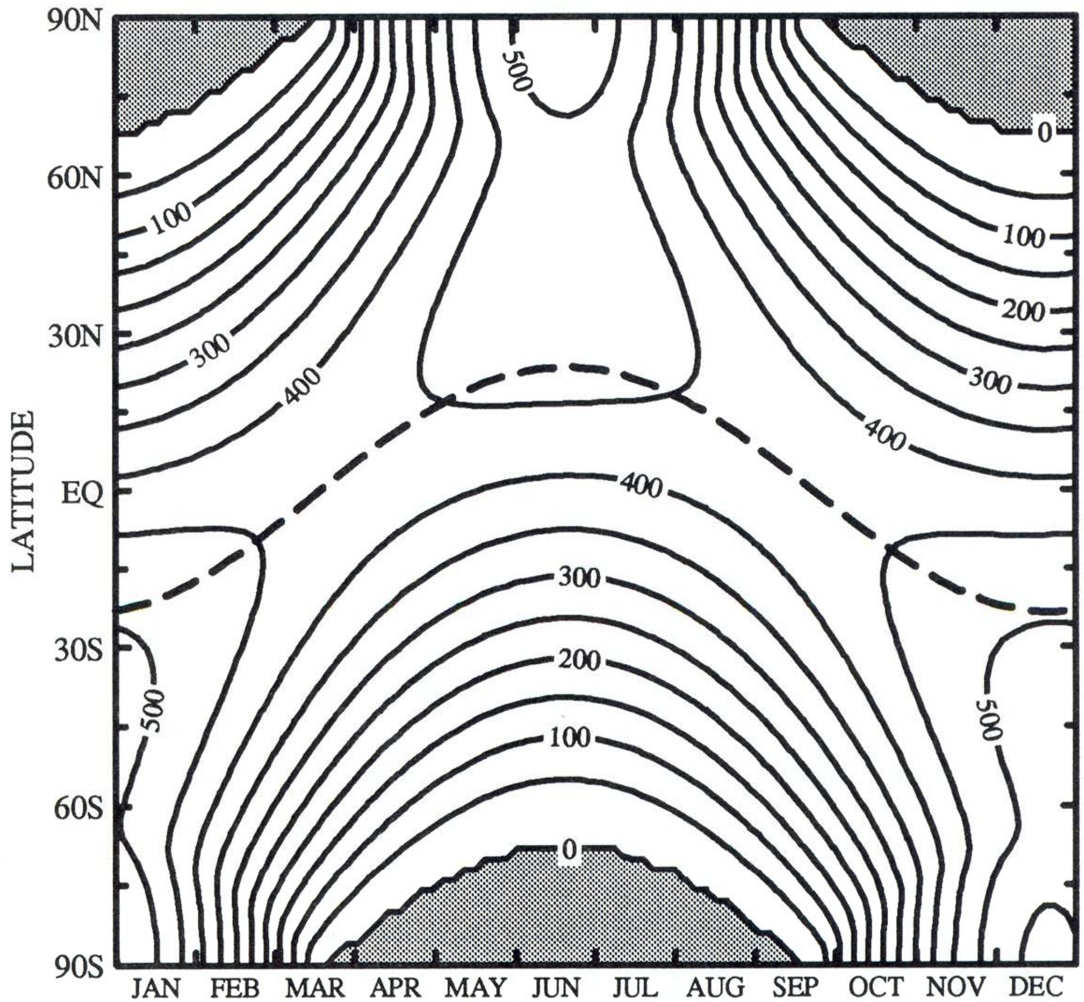


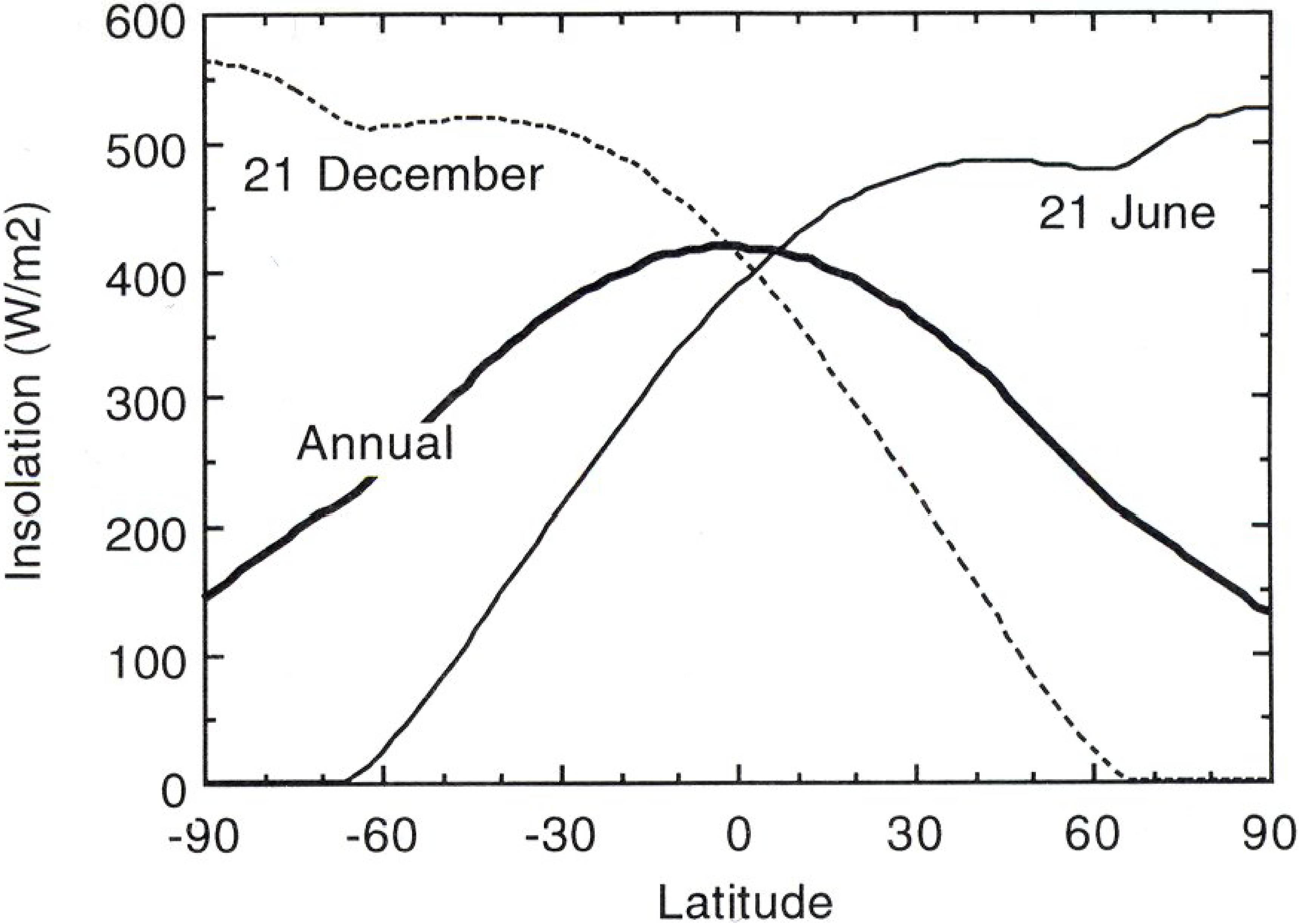

Thermosphere
Mesopause
Mesosphere
Stratopause
Stratosphere
Tropopause
Troposphere
100000.0
90000.0
80000.0
70000.0
60000.0
50000.0
40000.0
30000.0
20000.0
10000.0
0.0
-100.0
-80.0
-60.0
-40.0
-20.0
0.0
20.0
Altitude (meters)
Temperature °C

The Real World Environment Varies in All Three Spatial Dimensions and Time

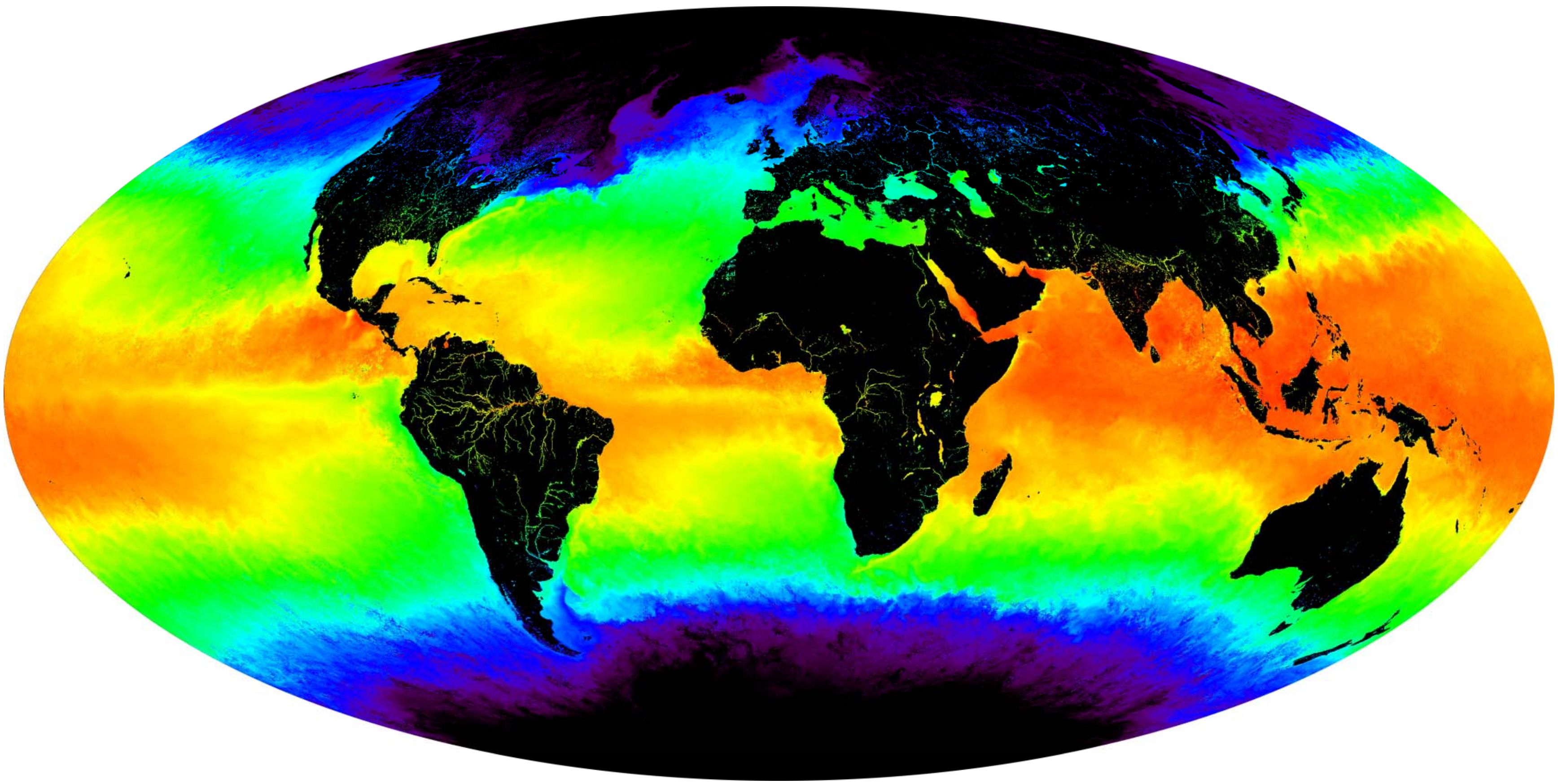

30
20
10
0
-10
-20
-30
Temperature (°C)
January
July
Annual
-90 -75 -60 -45 -30 -15 0 15 30 45 60 75 90
Latitude

Seasonal variation of solar radiation




600
500
400
300
200
100
0
Insolation (W/m2)
21 December
21 June
Annual
-90
-60
-30
0
30
60
90
Latitude

Sea Surface Temperature



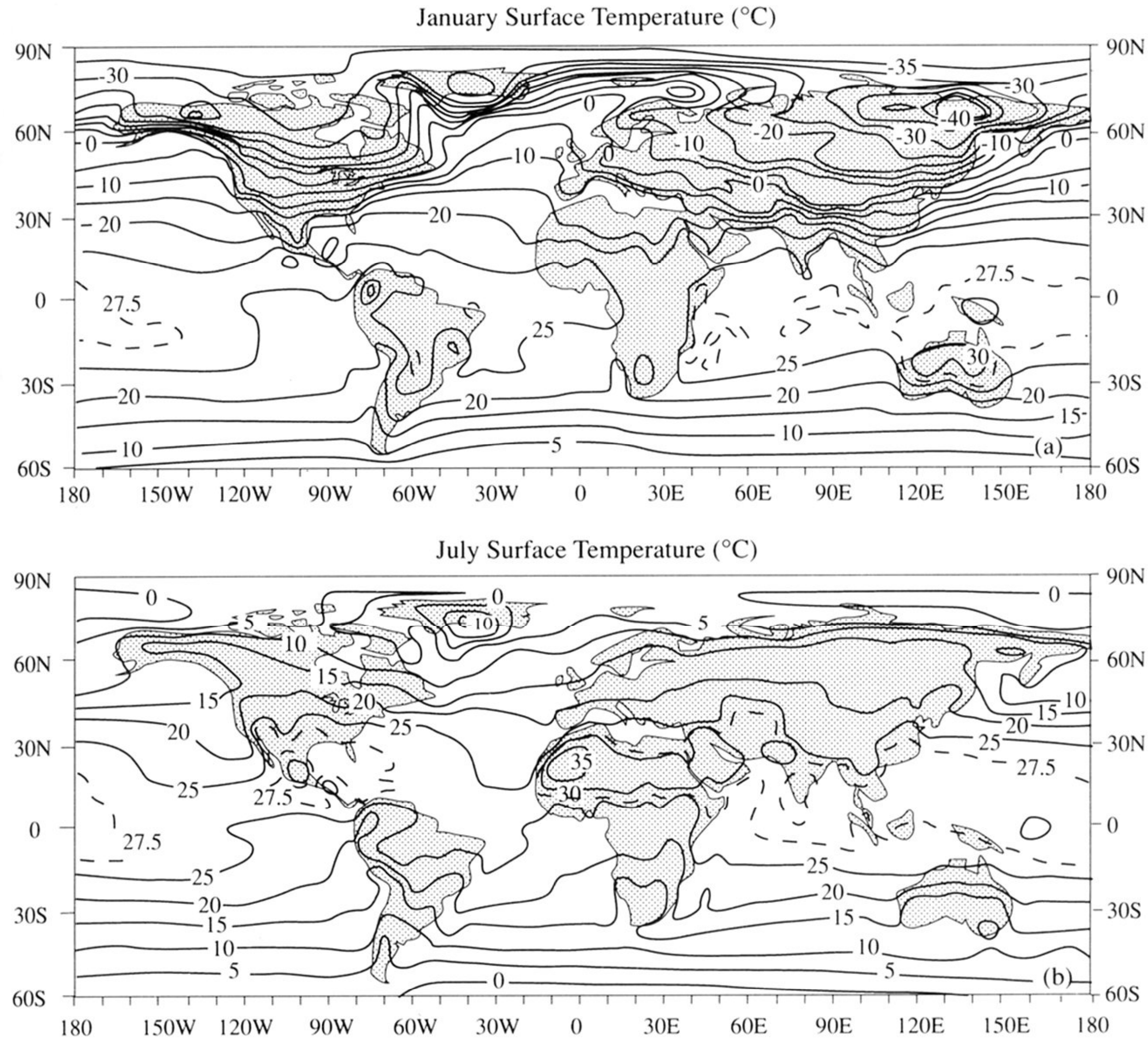


Fig. 1.6 Global map of the (a) January and (b) July surface temperature. [From Shea (1986). Reproduced with permission from the National Center for Atmospheric Research.]

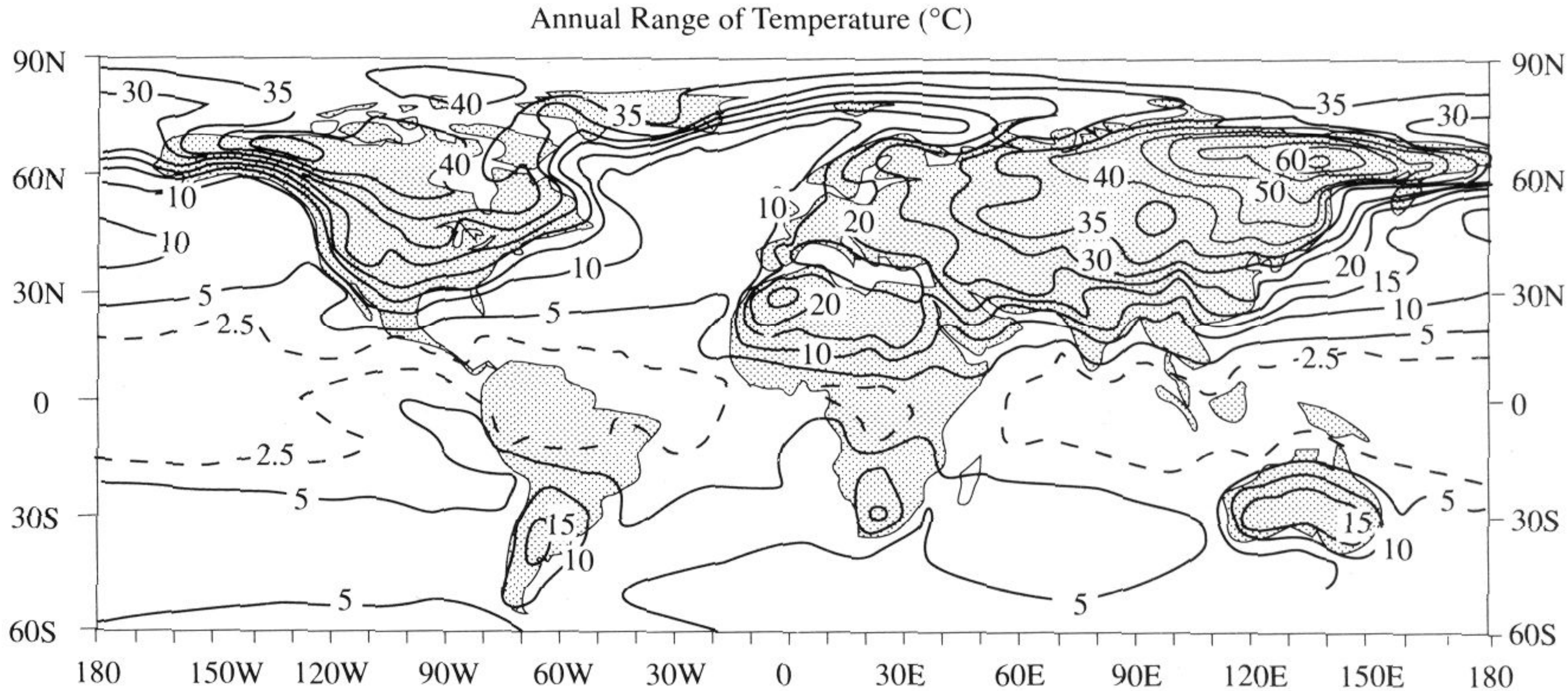
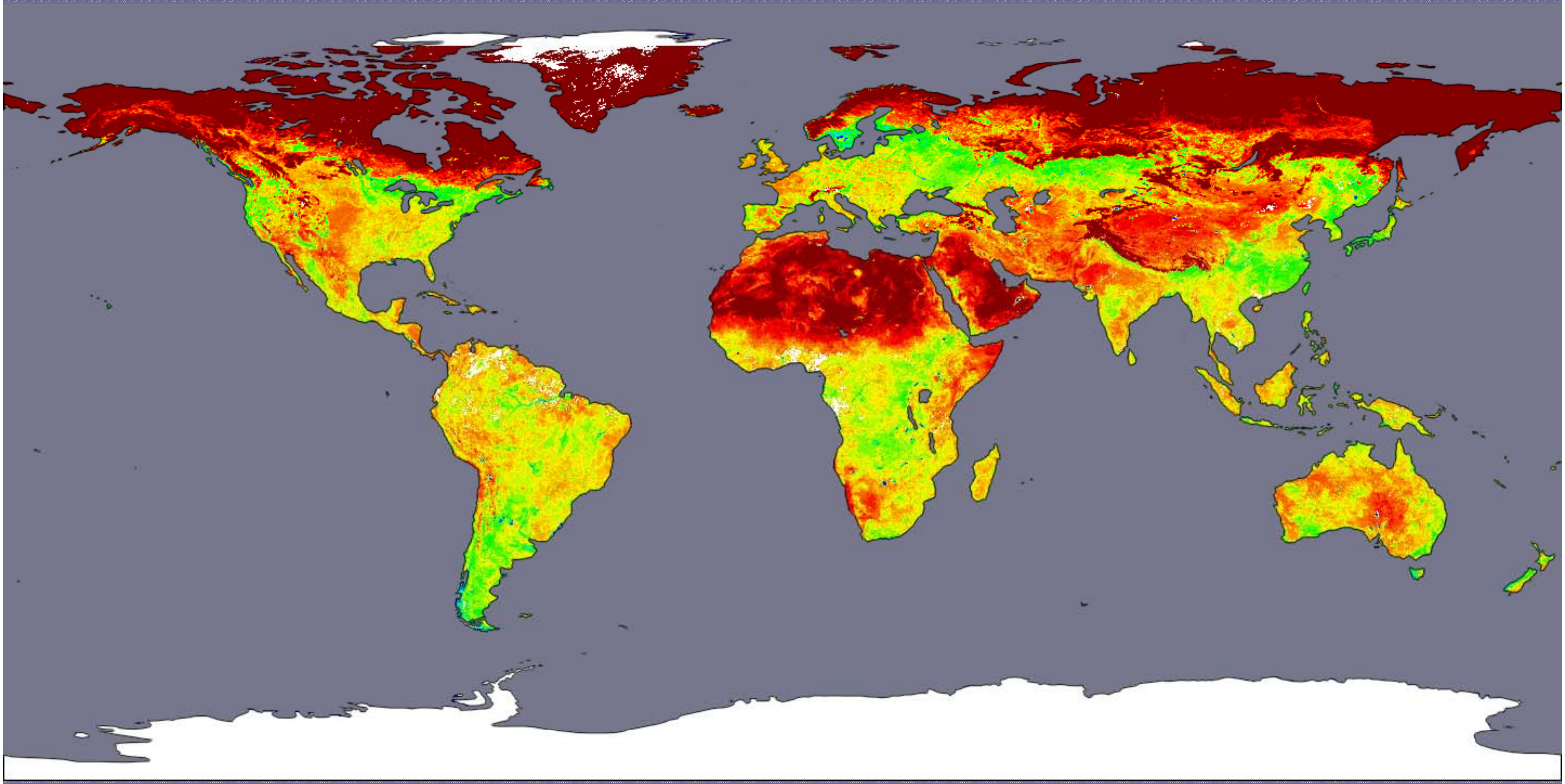
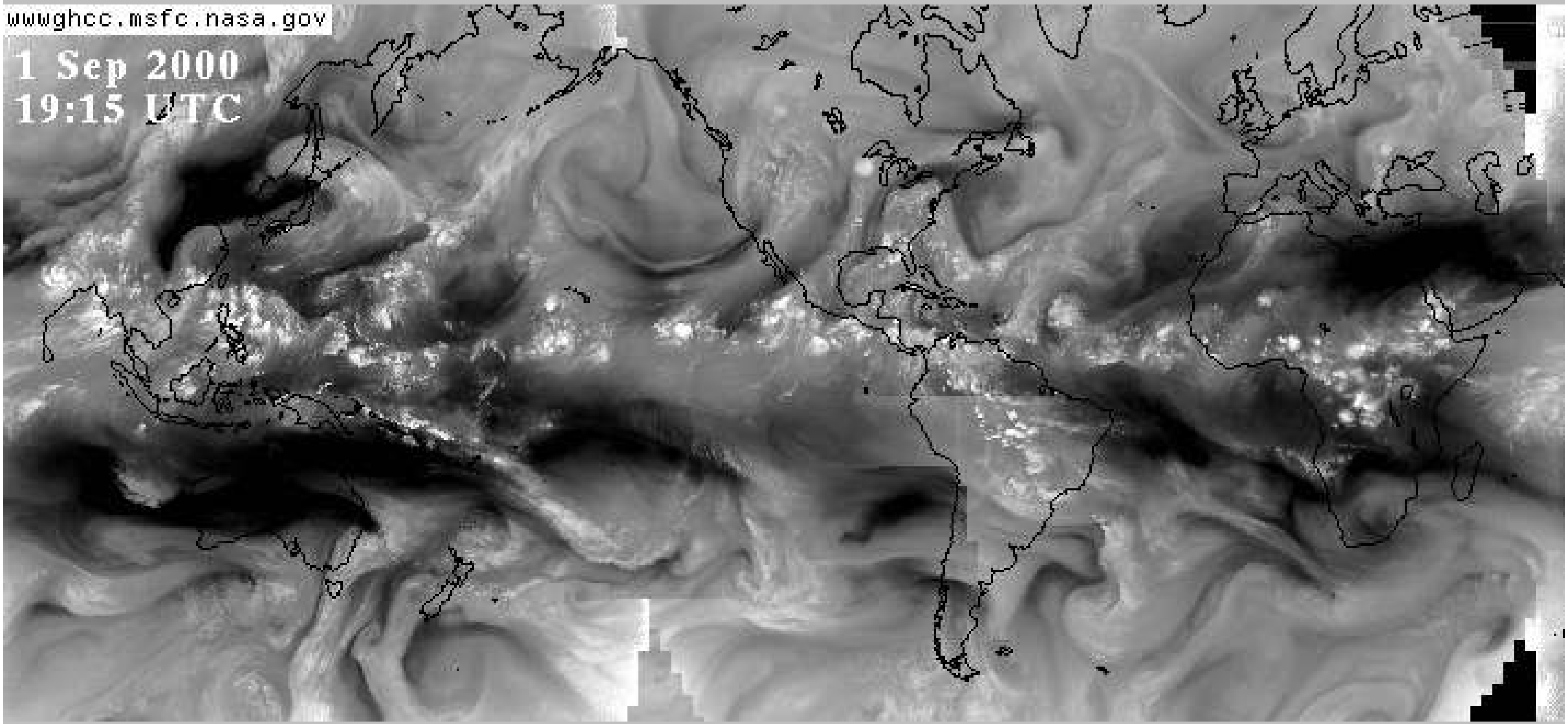


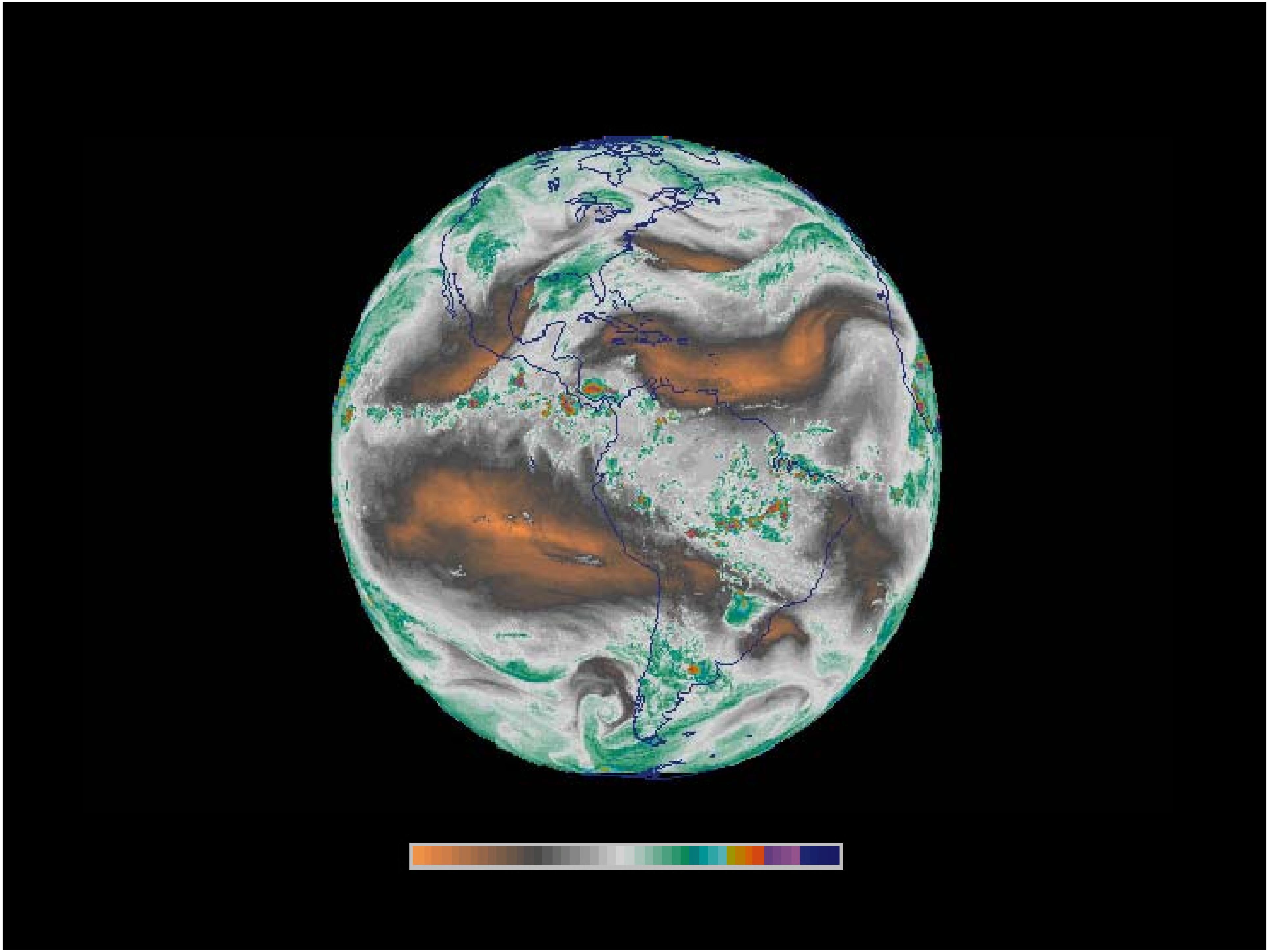
Fig. 1.7 Map of the amplitude of the annual cycle of surface temperature. [From Shea (1986). Reproduced with permission from the National Center for Atmospheric Research.]

Global average surface albedo



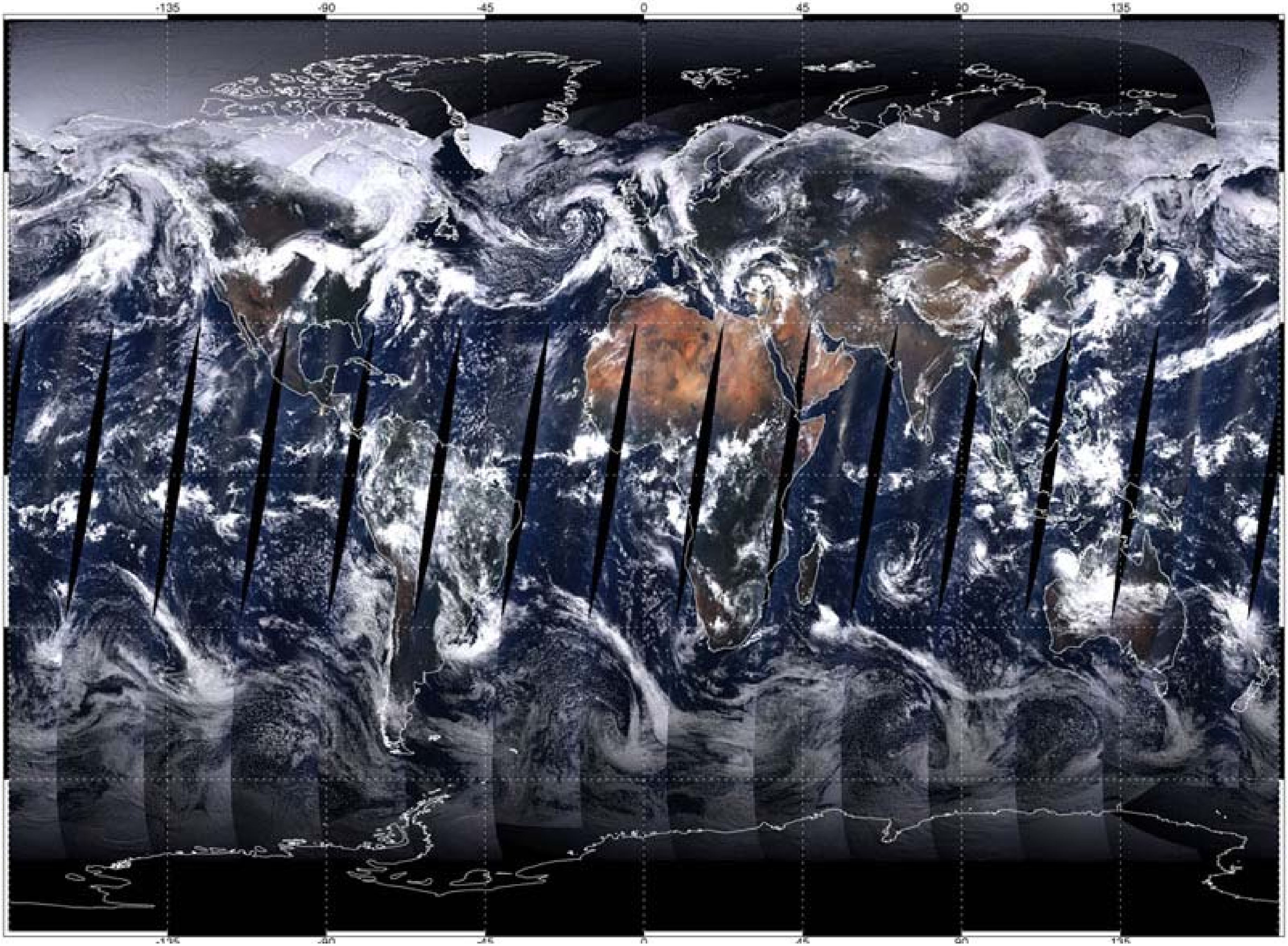
Inhomogeneity of Water Vapor





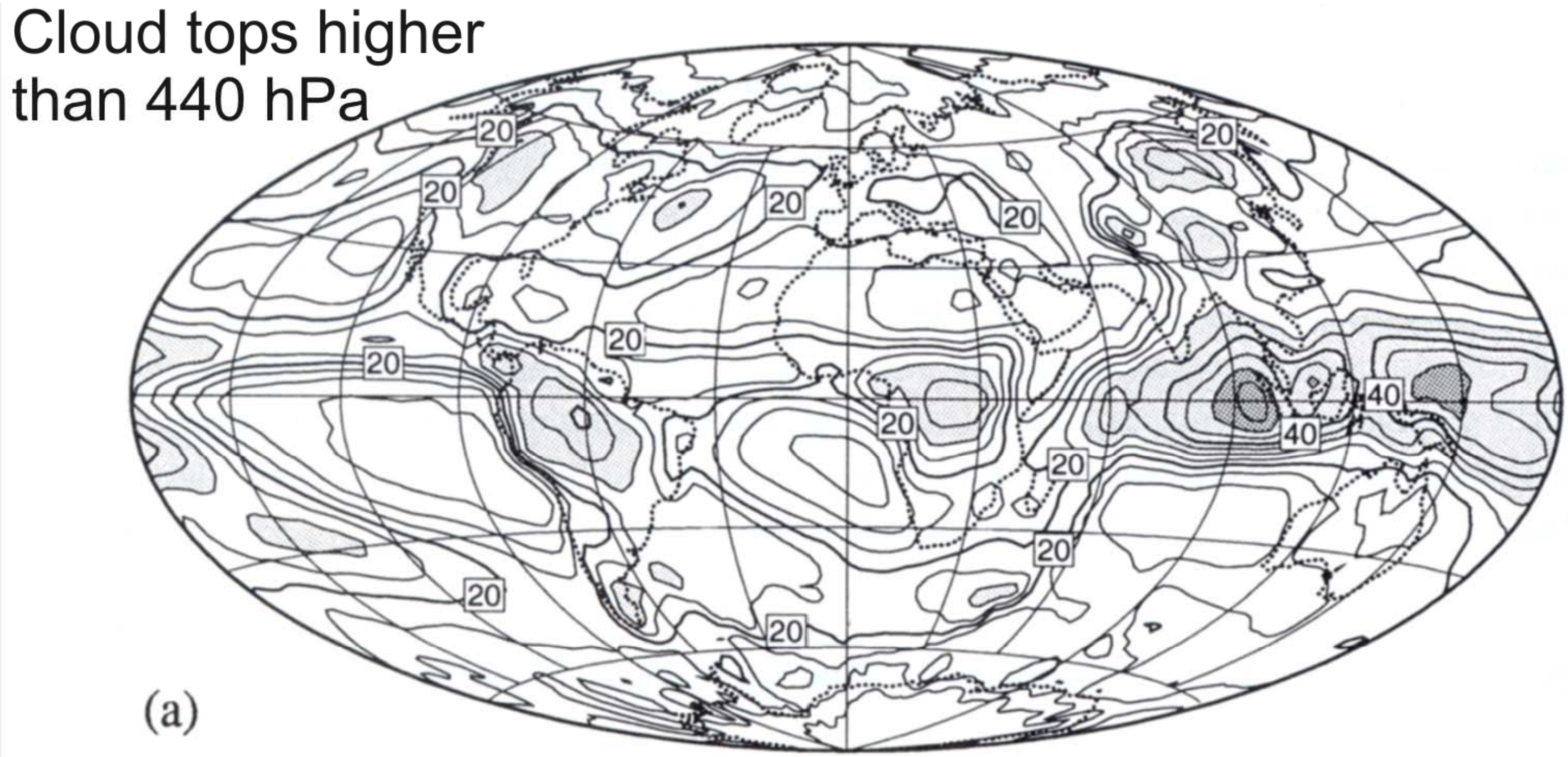
Effects of Clouds on Radiative Transfer

- Responsible for much of Earth's albedo
- Important greenhouse effect from longwave absorption and re-emission

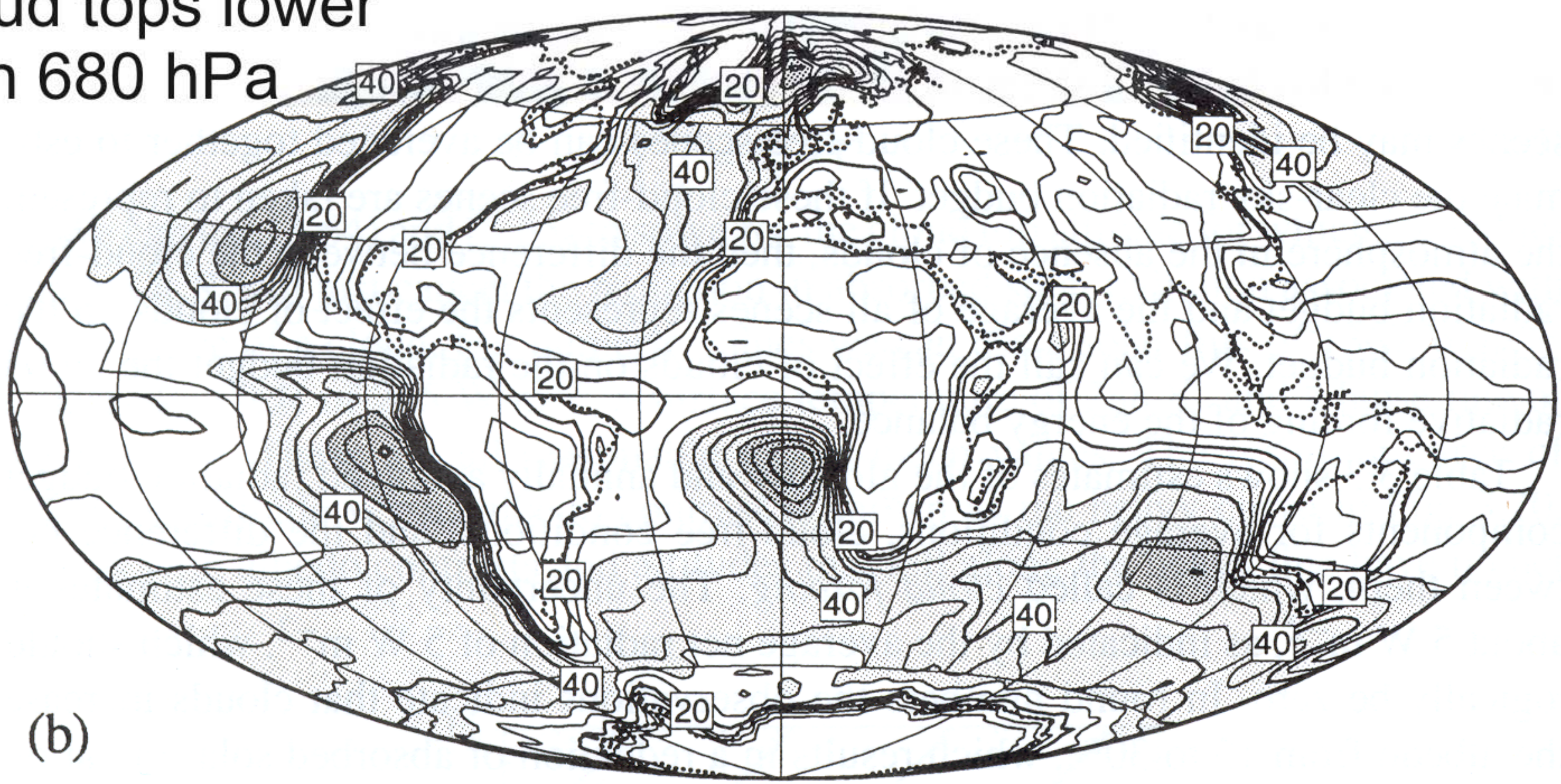


-135
-90
-45
0
45
90
135

Annual Average Cloud Fractional Area

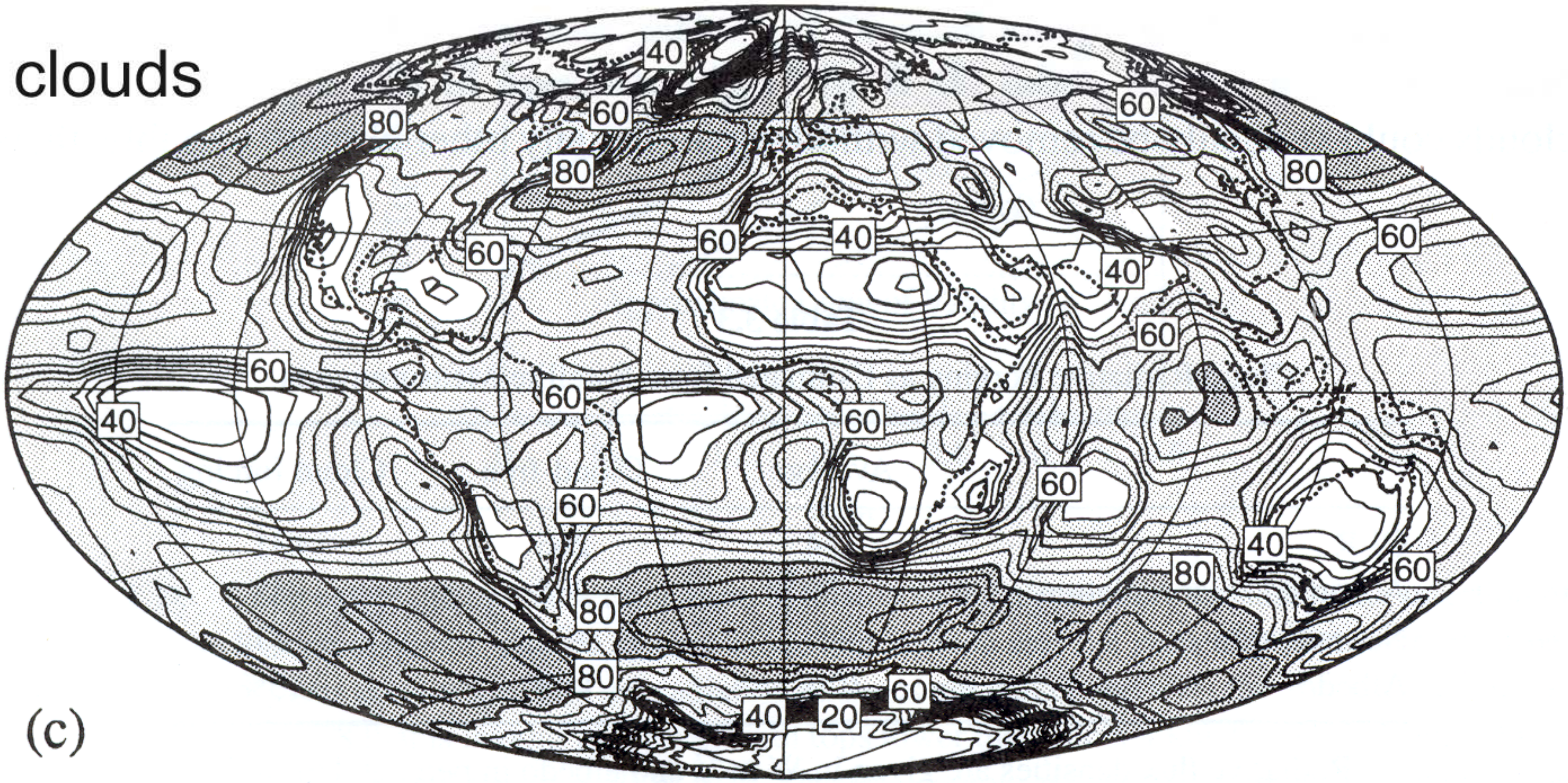


Cloud tops lower than 680 hPa



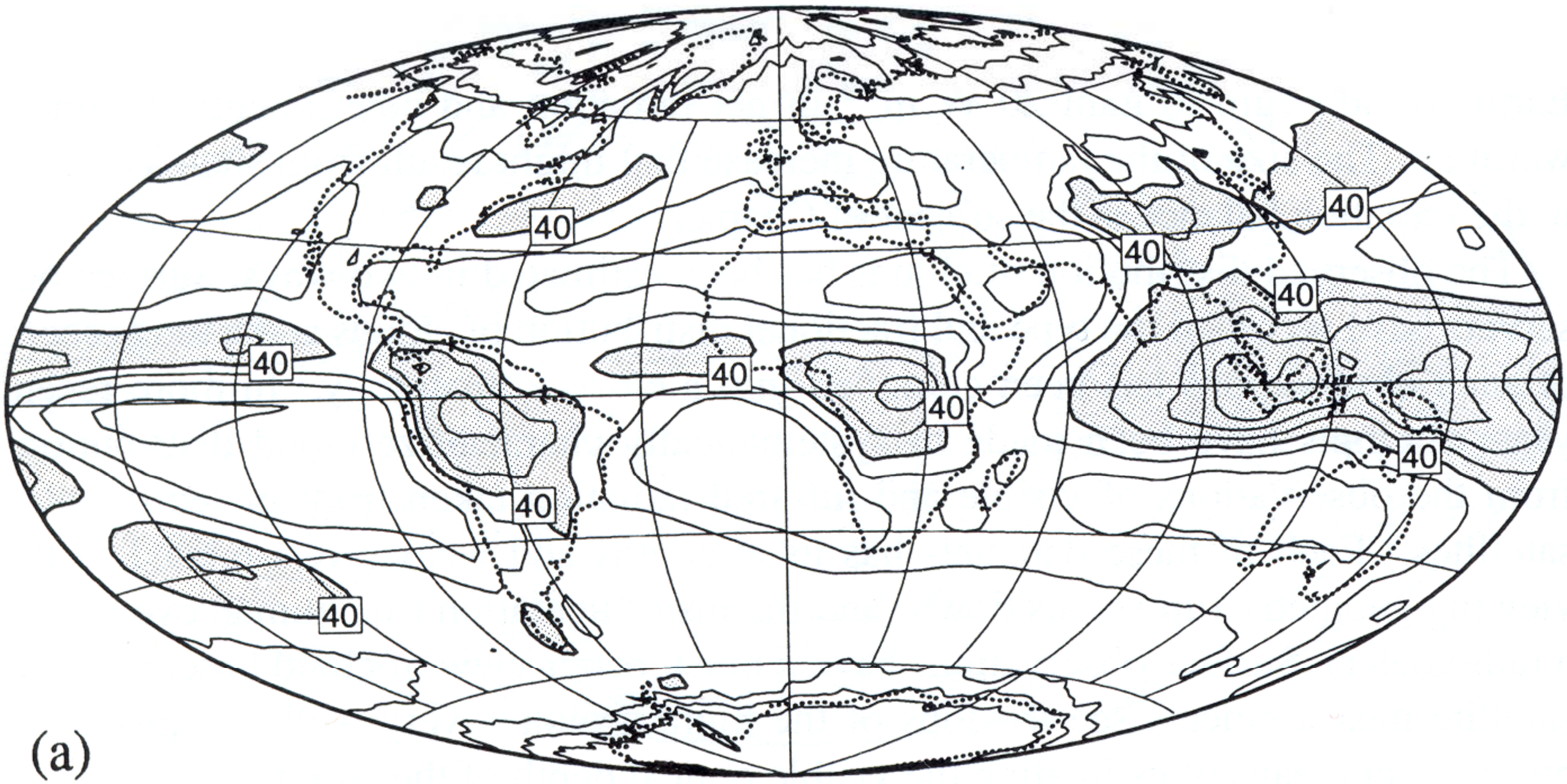
(b)

All clouds


40
60
80
80
60
80
60
60
40
60
40
60
60
60
40
60
60
60
60
40
80
60
80
80
60
40
20

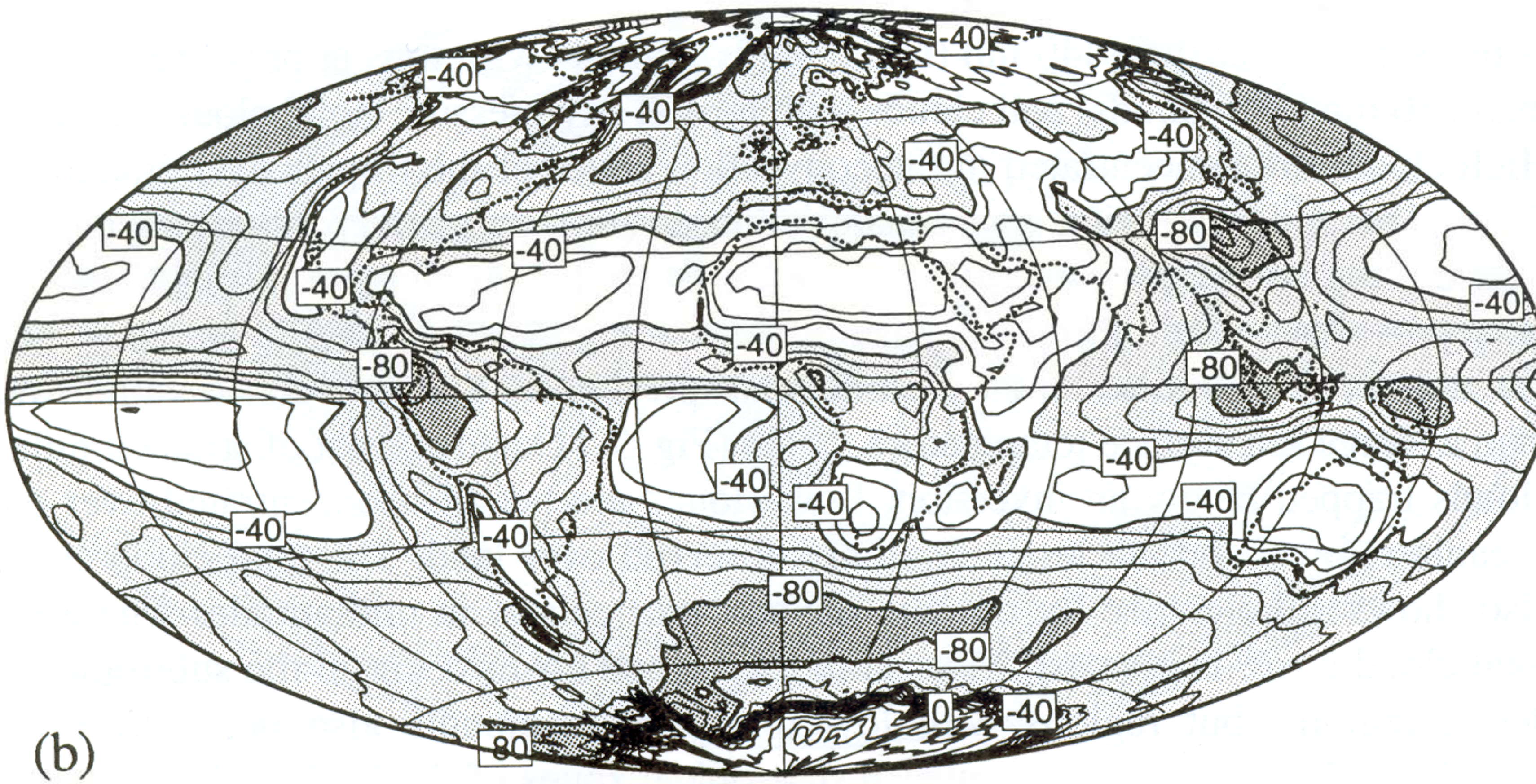
(c)

Observed cloud effects on radiation

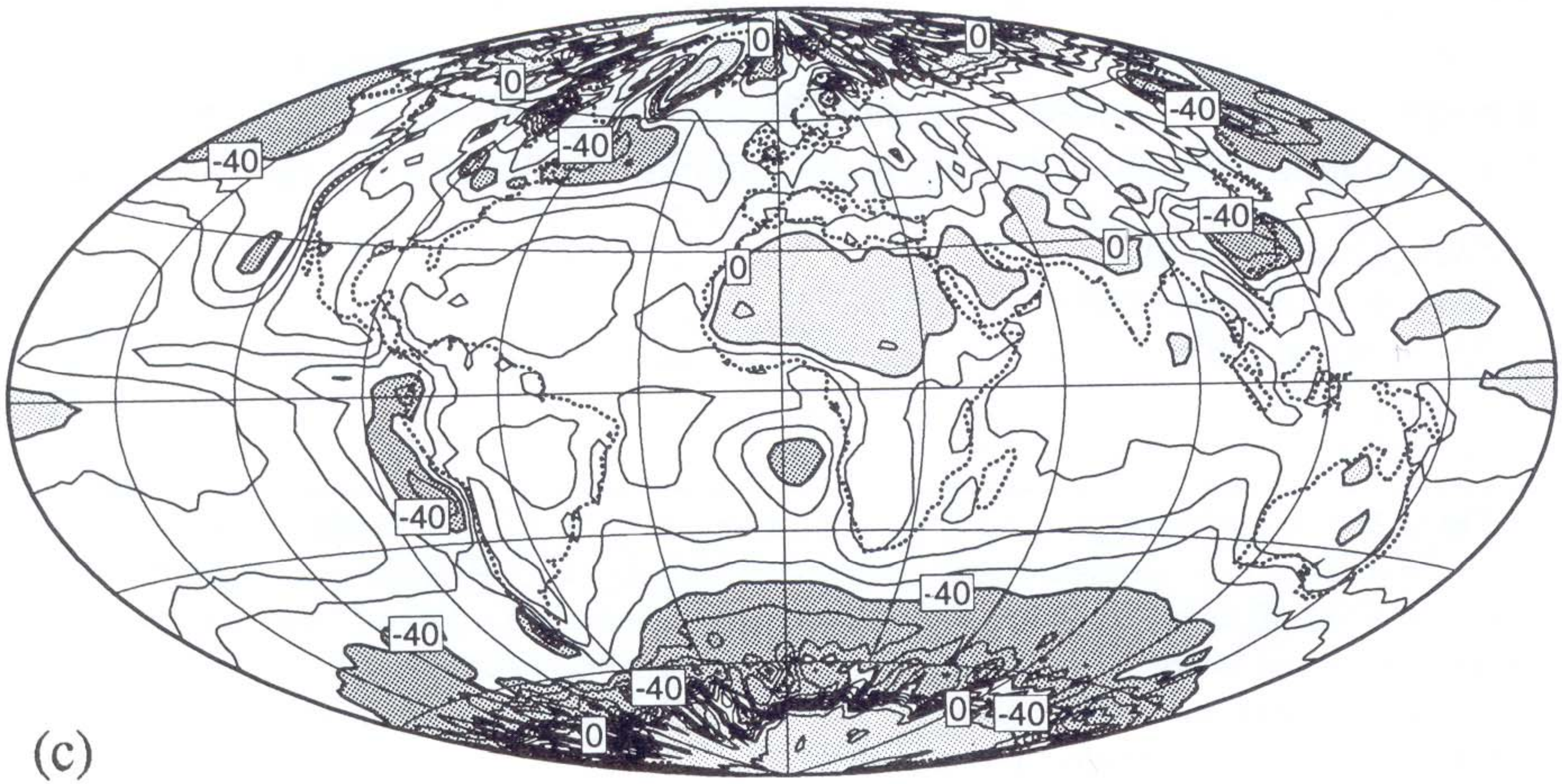


(a)

OLR reduction (W/m²)

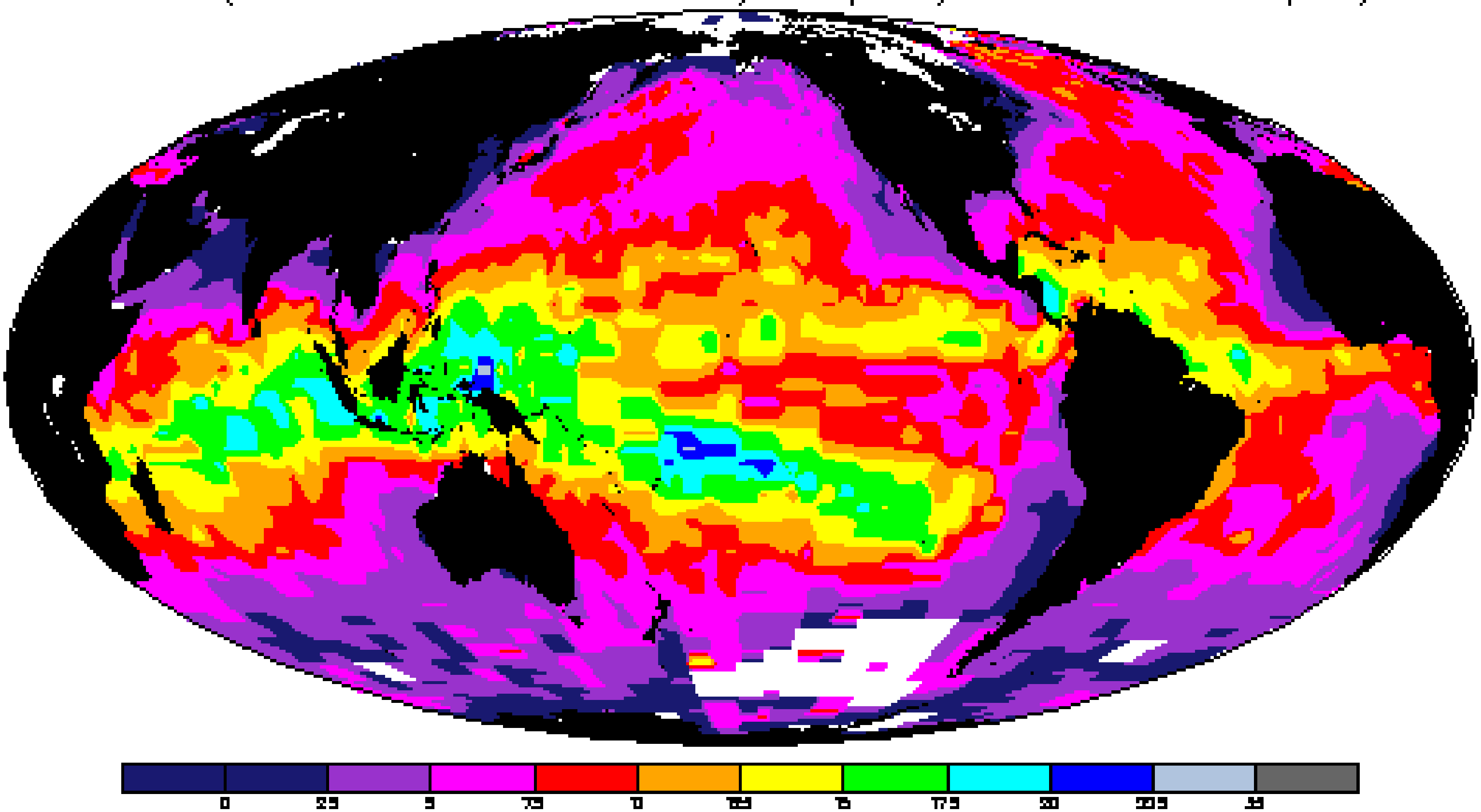


Change in shortwave radiation (W/m^2)

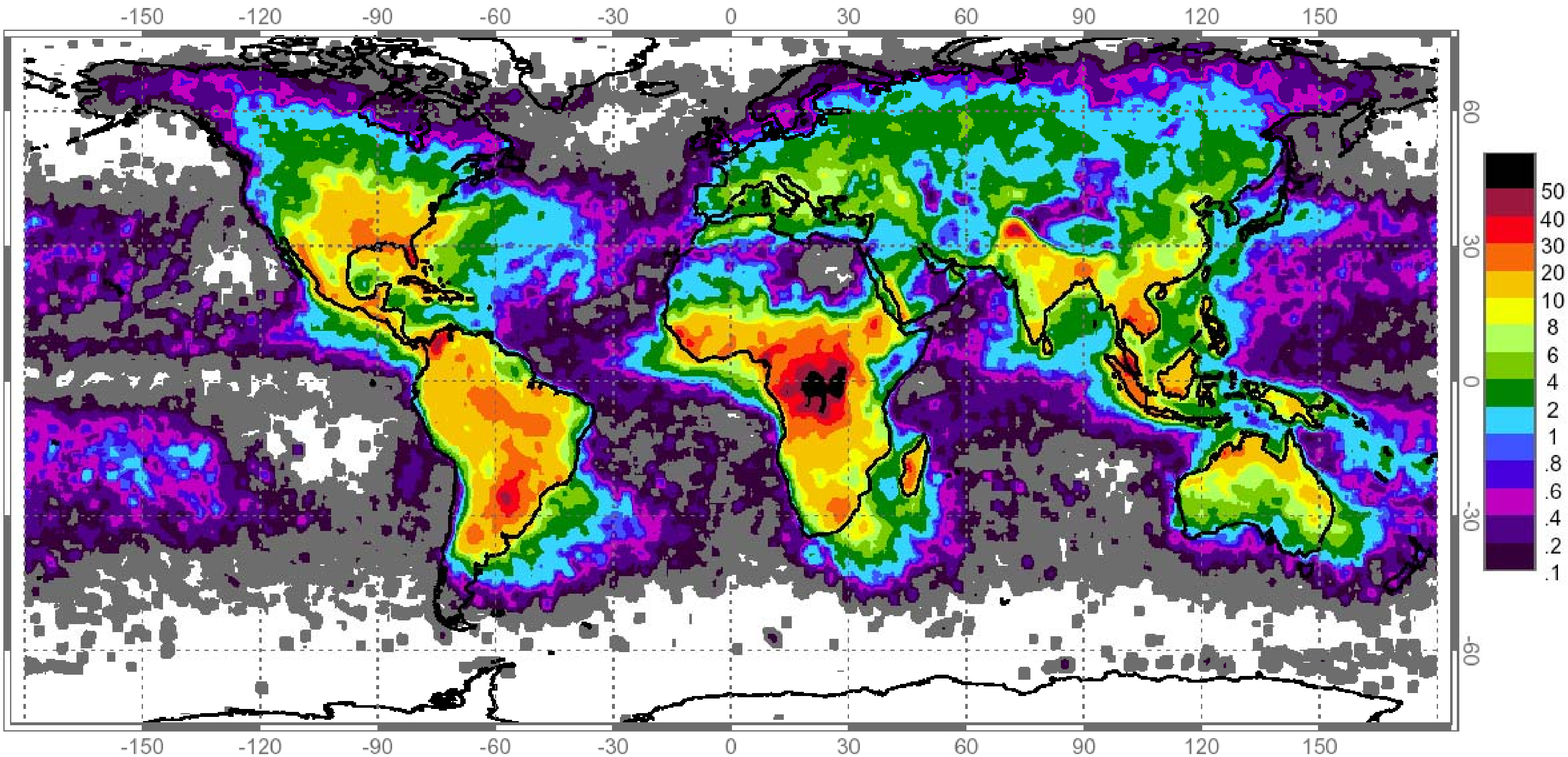


Increase in net radiation (W/m^2)

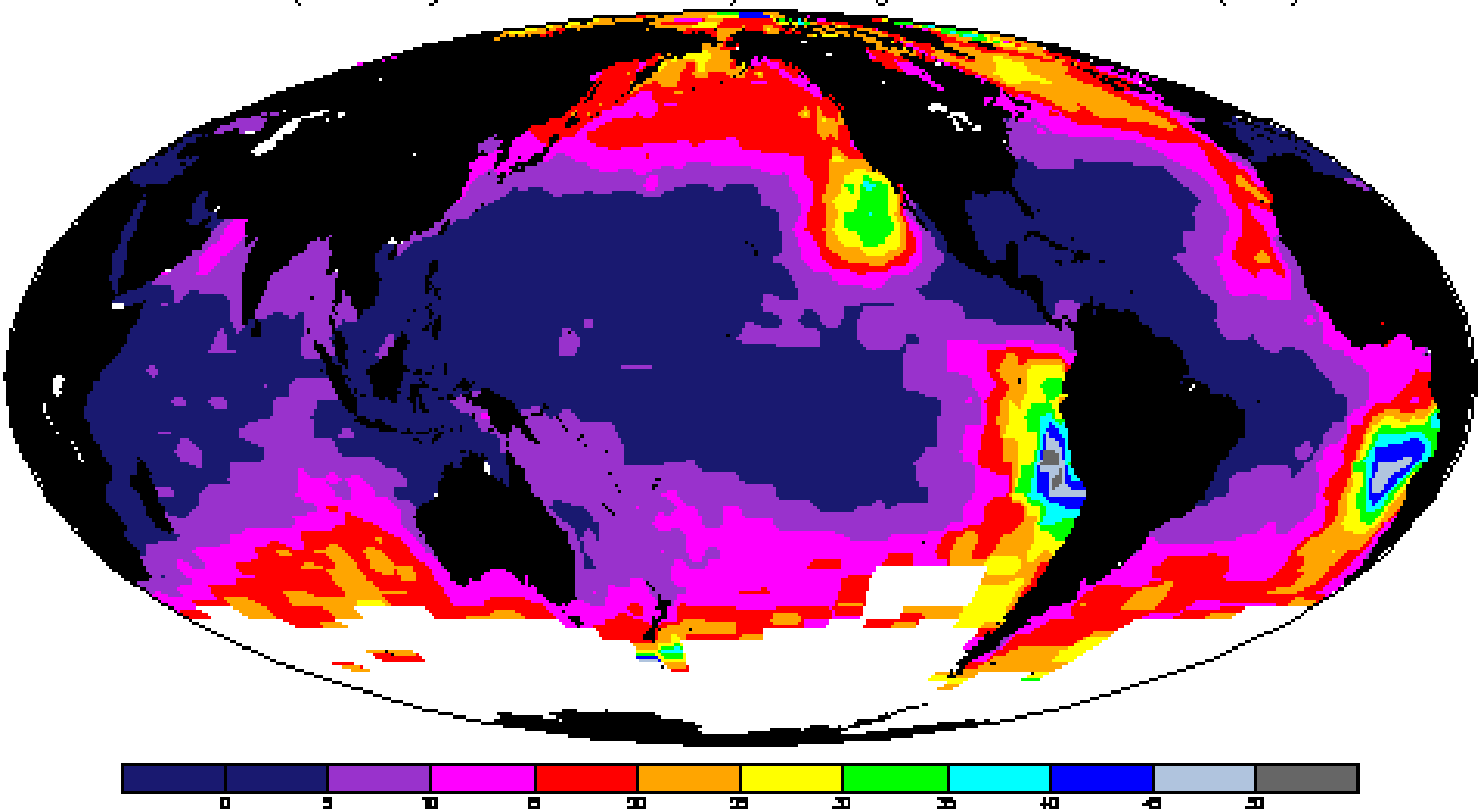
CL 3 (Cumulonimbus without Anvil) Frequency-of-Occurrence (DJF)

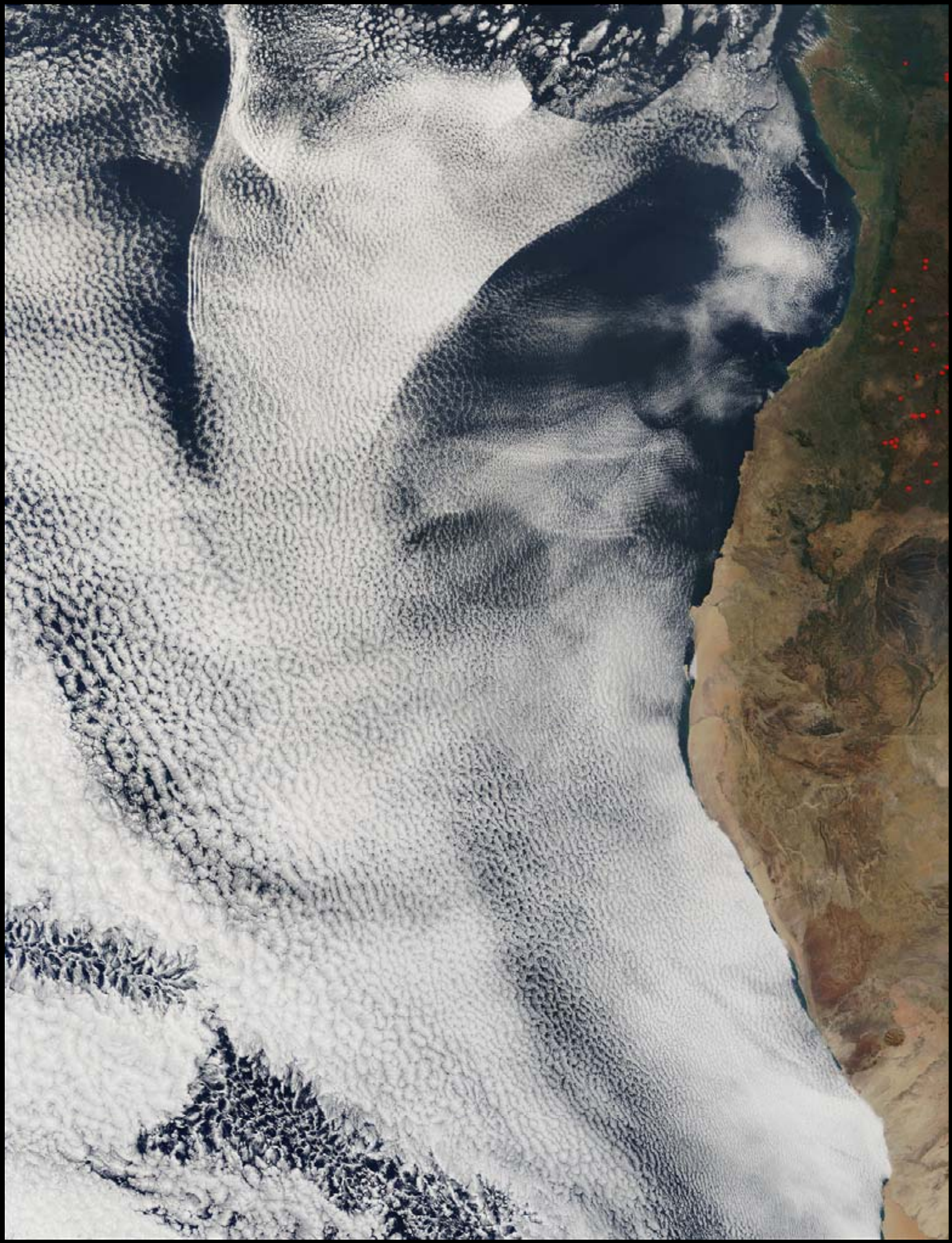


Annual Average Lightning Frequency (from Space)



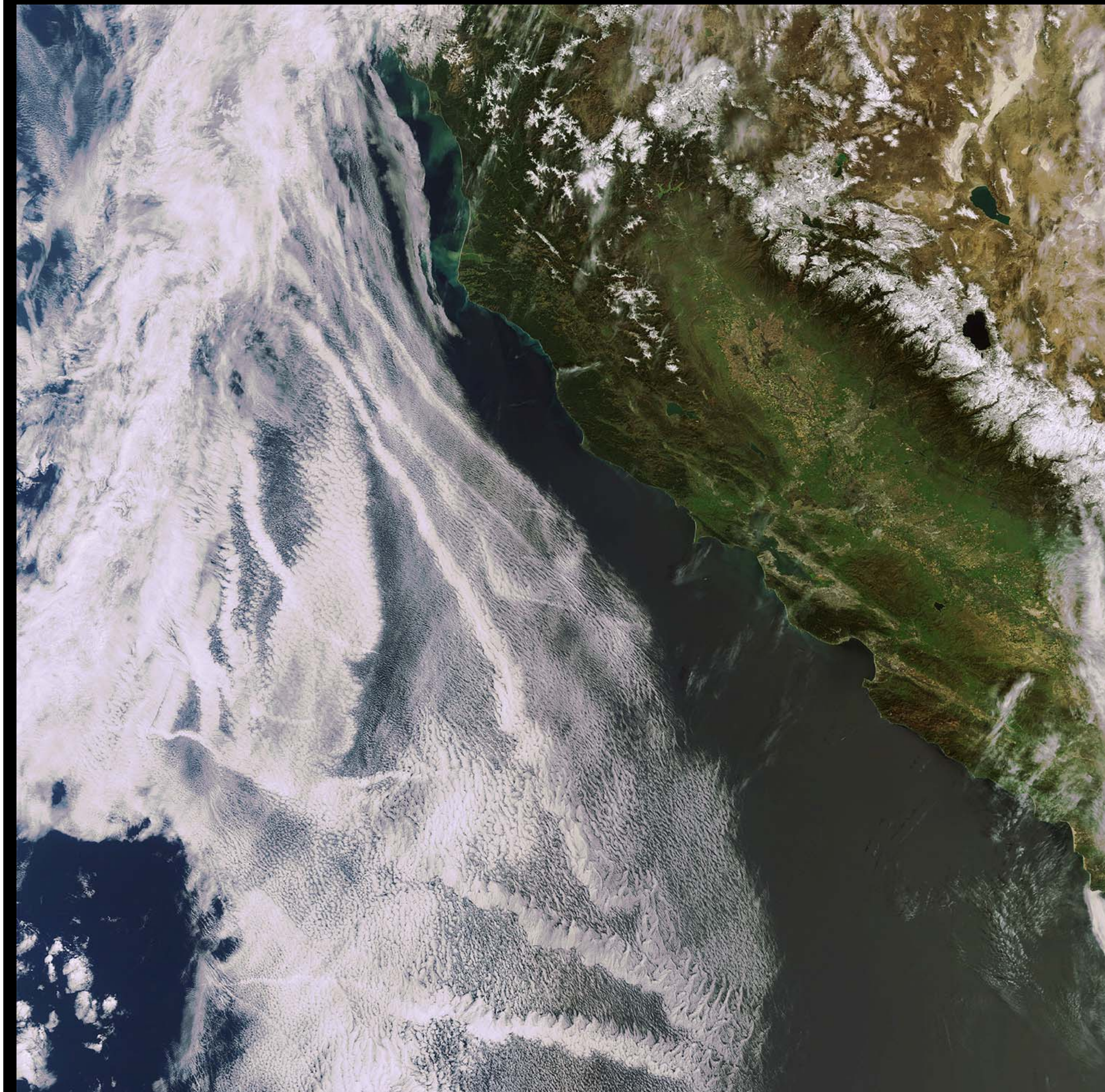
CL 5 (Ordinary Stratocumulus) Average Cloud Amount (JJA)





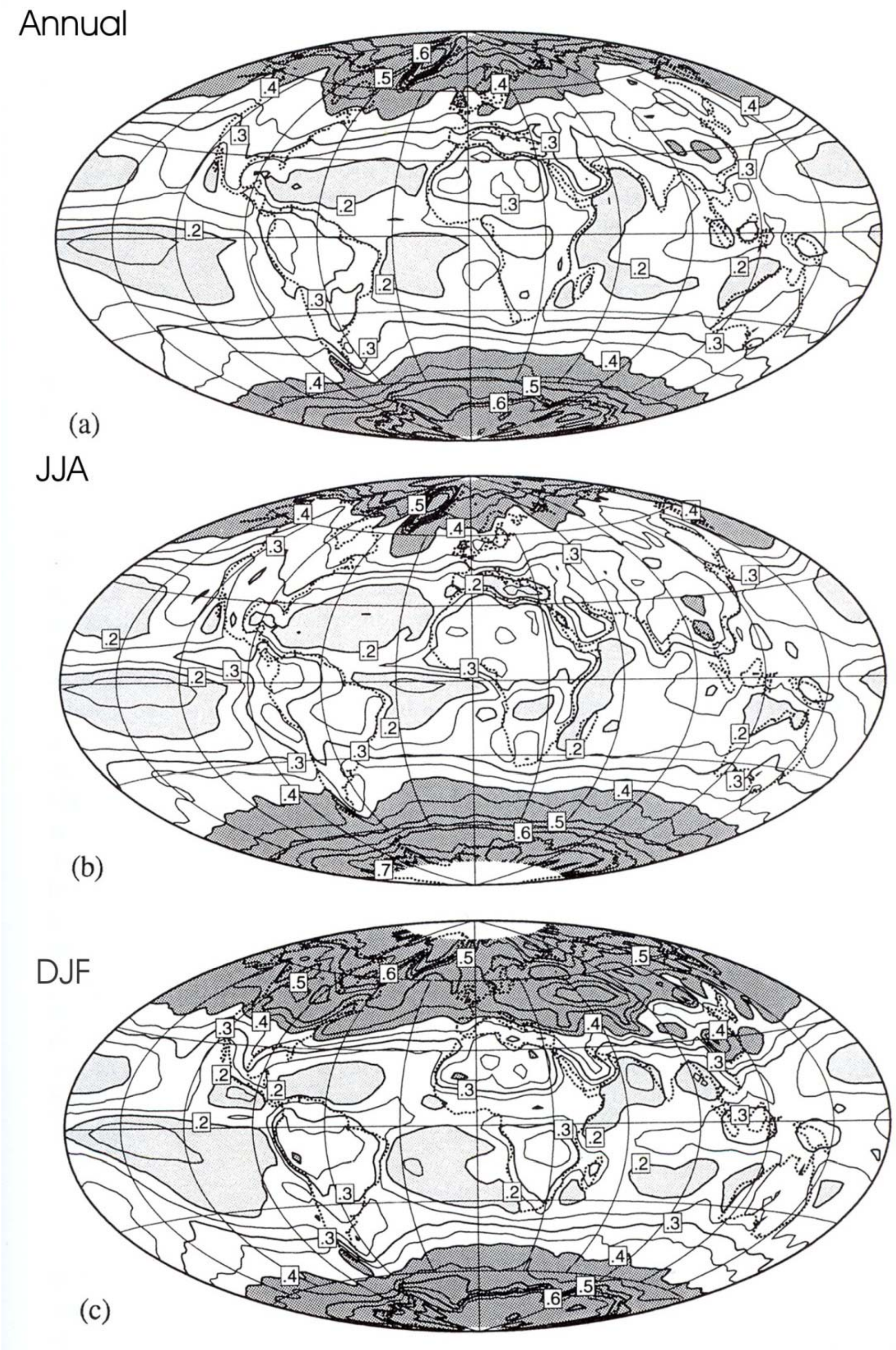
Stratocumuli off Angola and Namibia

Stratocumuli off Baja California

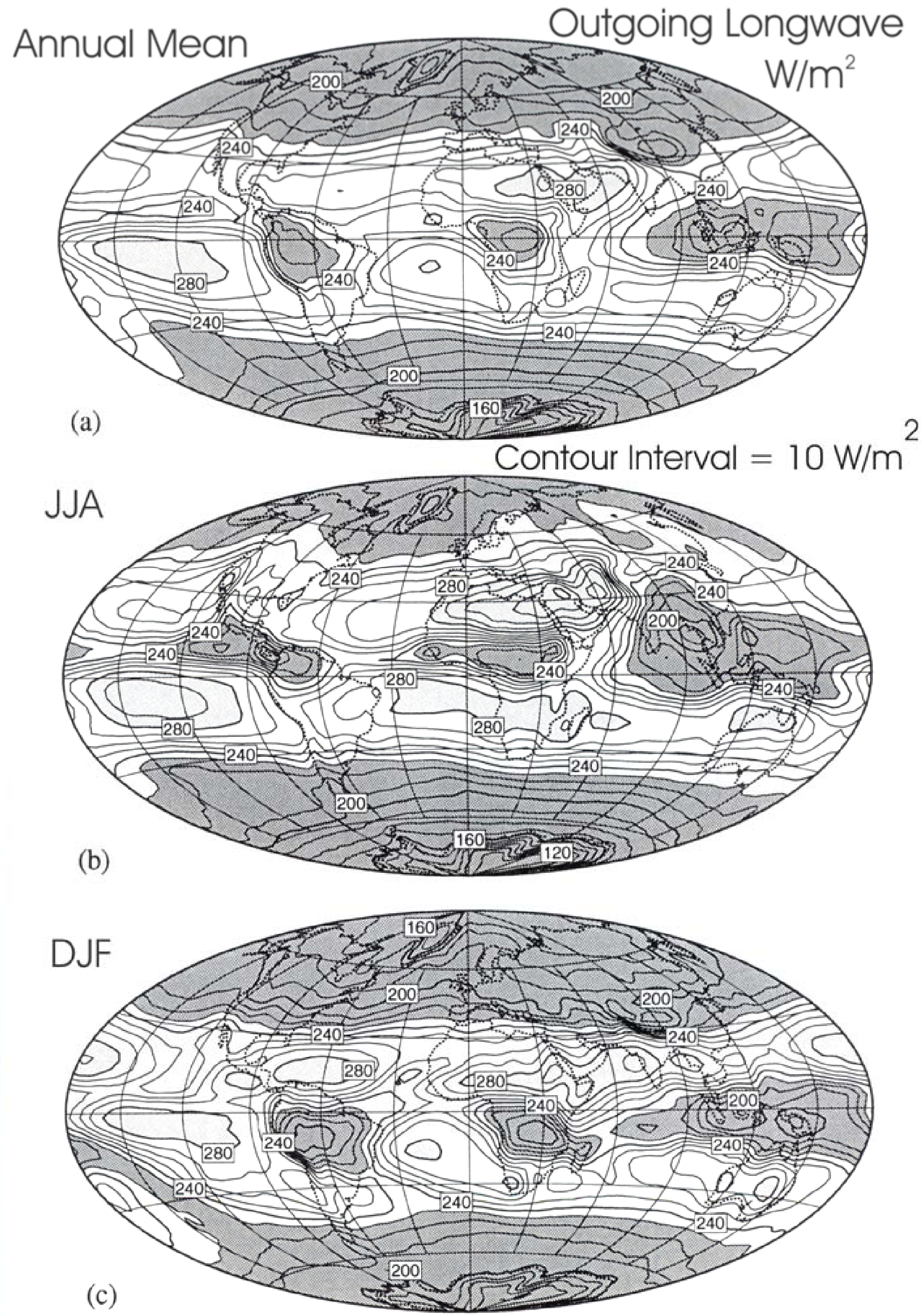


Ship tracks off California

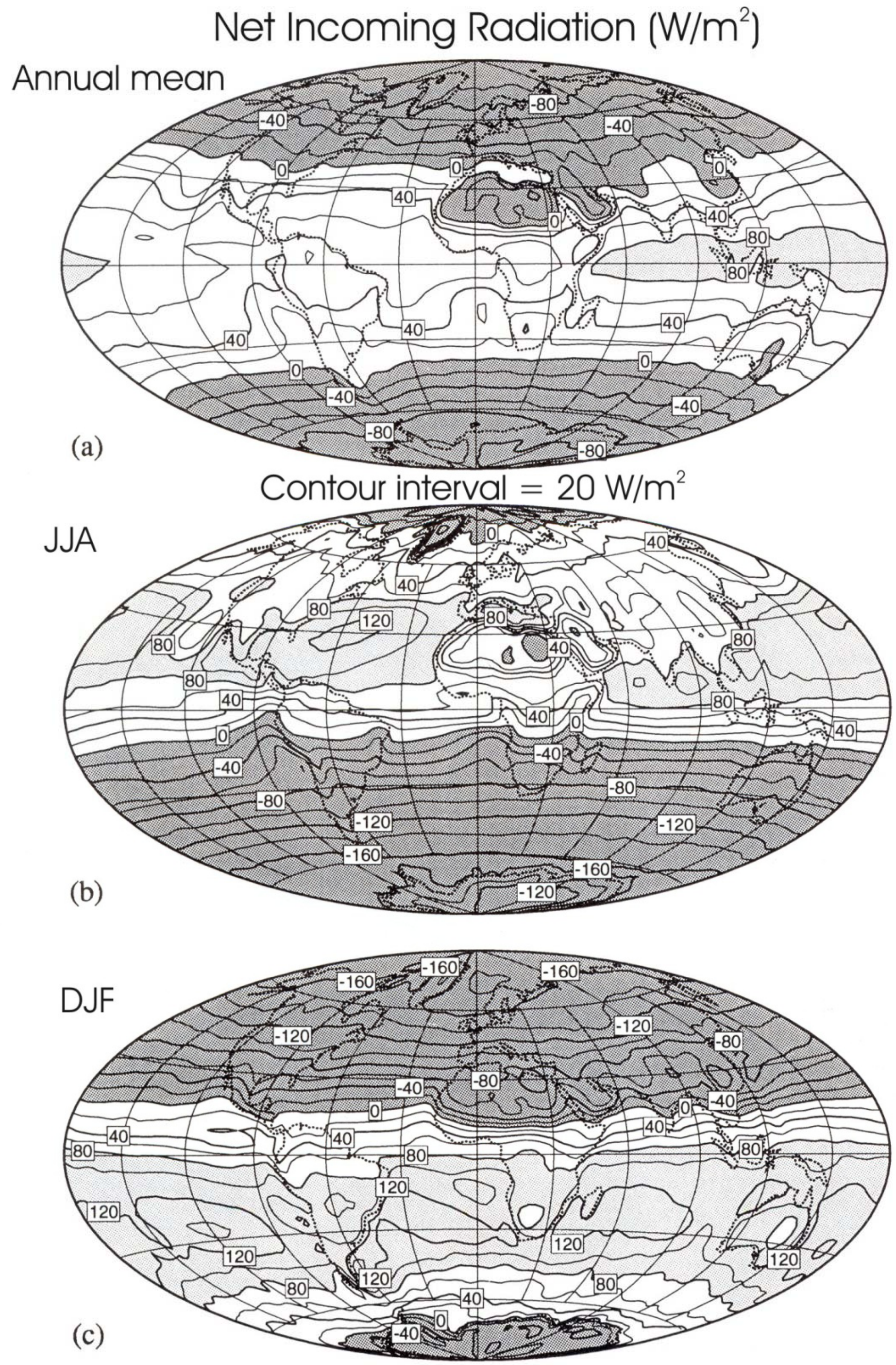
Total Albedo



Outgoing Longwave Radiation

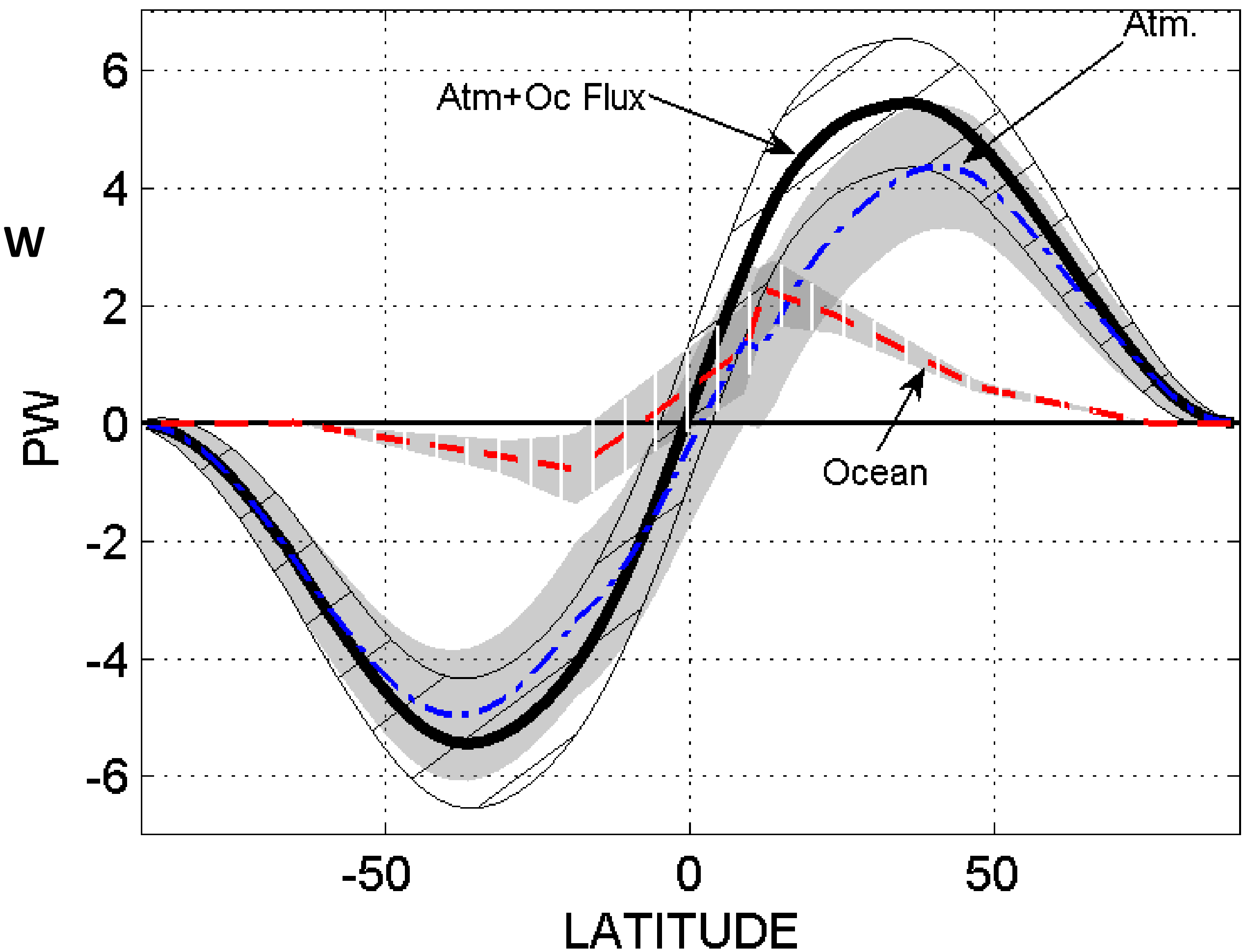


Net Incoming Radiation

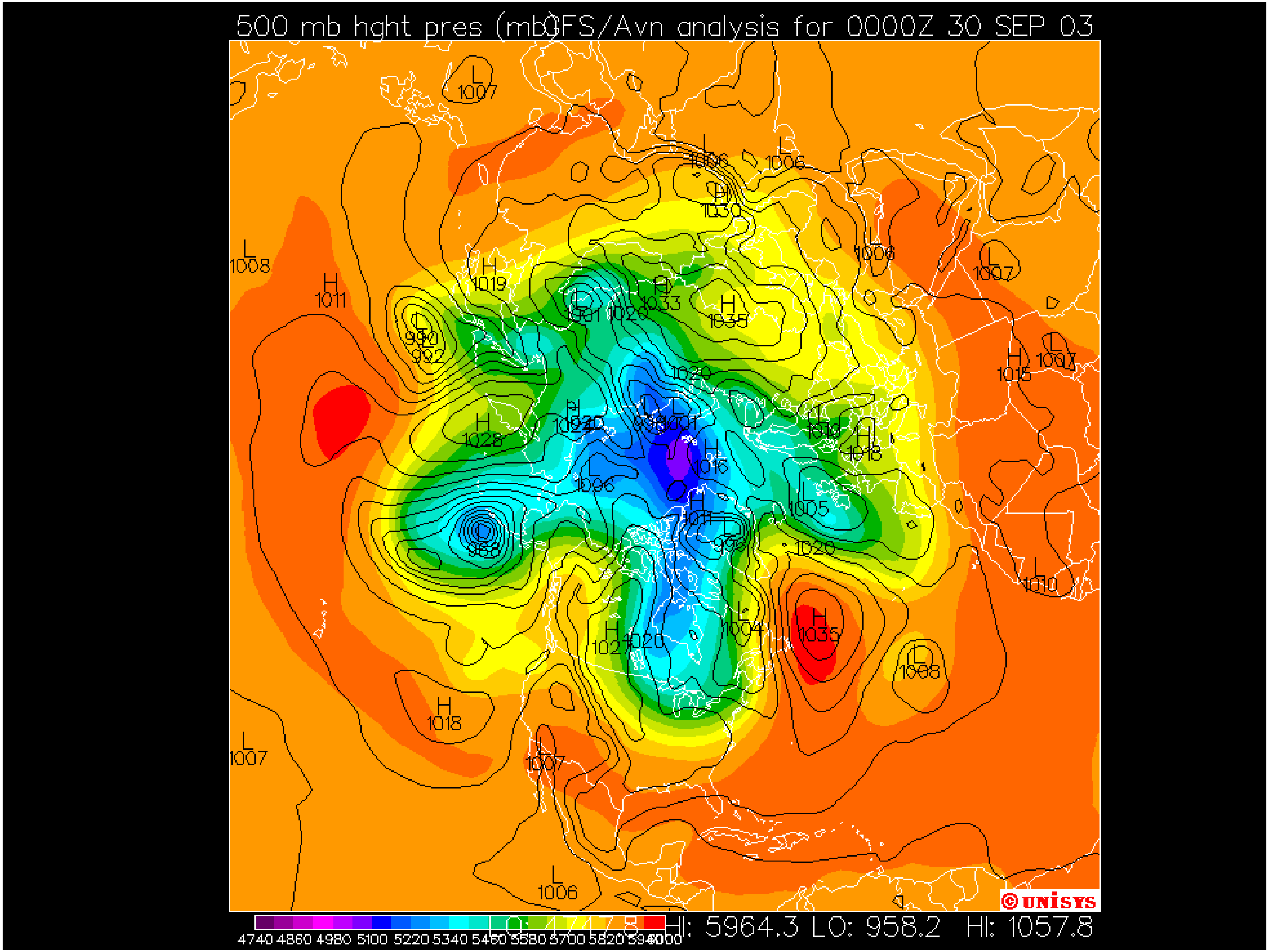


Lateral Heat Transport by Atmosphere and Oceans

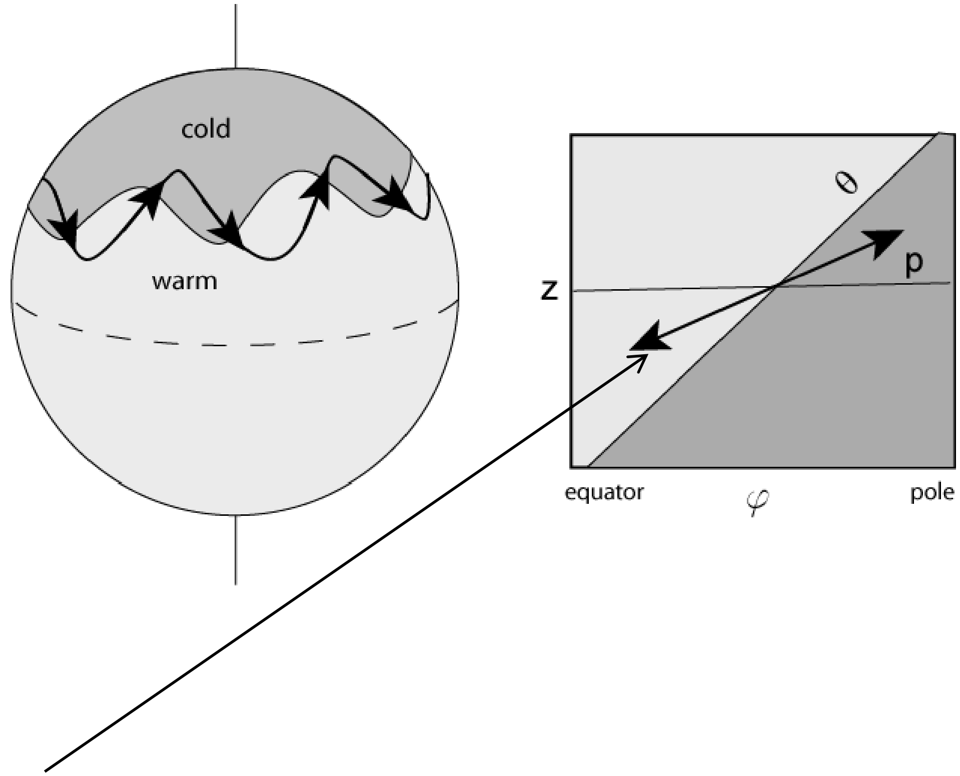
1 PW = 10^{15} W



Circulation of the Atmosphere


500 mb hght pres (mb) GFS/Avn analysis for 0000Z 30 SEP 03
4740 4860 4980 5100 5220 5340 5460 5580 5700 5820 5940 6000
HI: 5964.3 LO: 958.2 HI: 1057.8
© UNISYS

Atmospheric heat transport



Warm air upward and poleward

Cold air downward and equatorward

Efficient poleward heat transport

Steady Flow:

$$\nabla \bullet \left[F_{rad}\hat{k} + F_{conv}\hat{k} + \rho \mathbf{V} E \right] = 0,$$

$$where$$

$$E \equiv c_p T + gz + L_v q + \frac{1}{2} |\, \mathbf{V} \,|^2$$

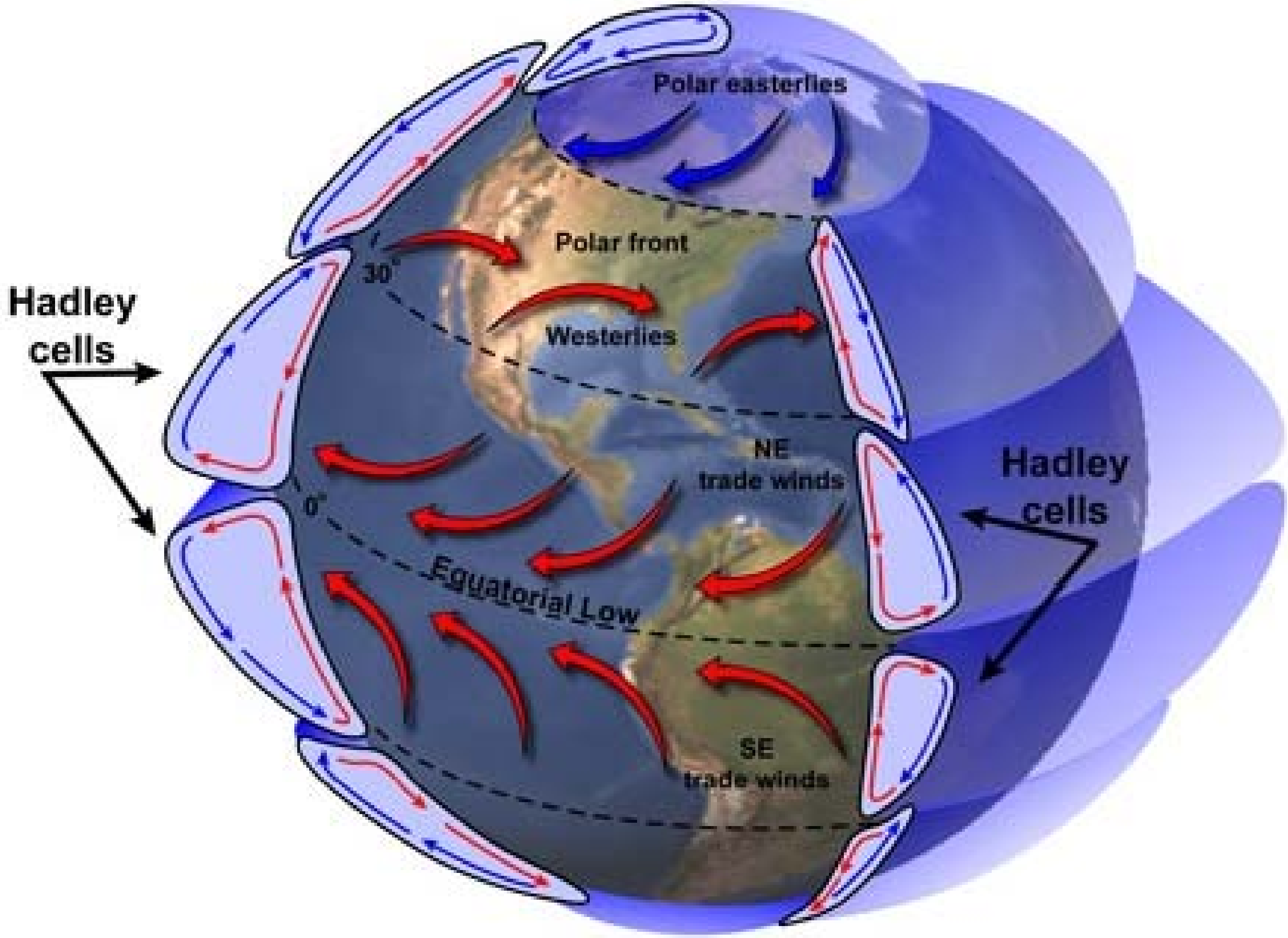
Integrate from surface to top of atmosphere:

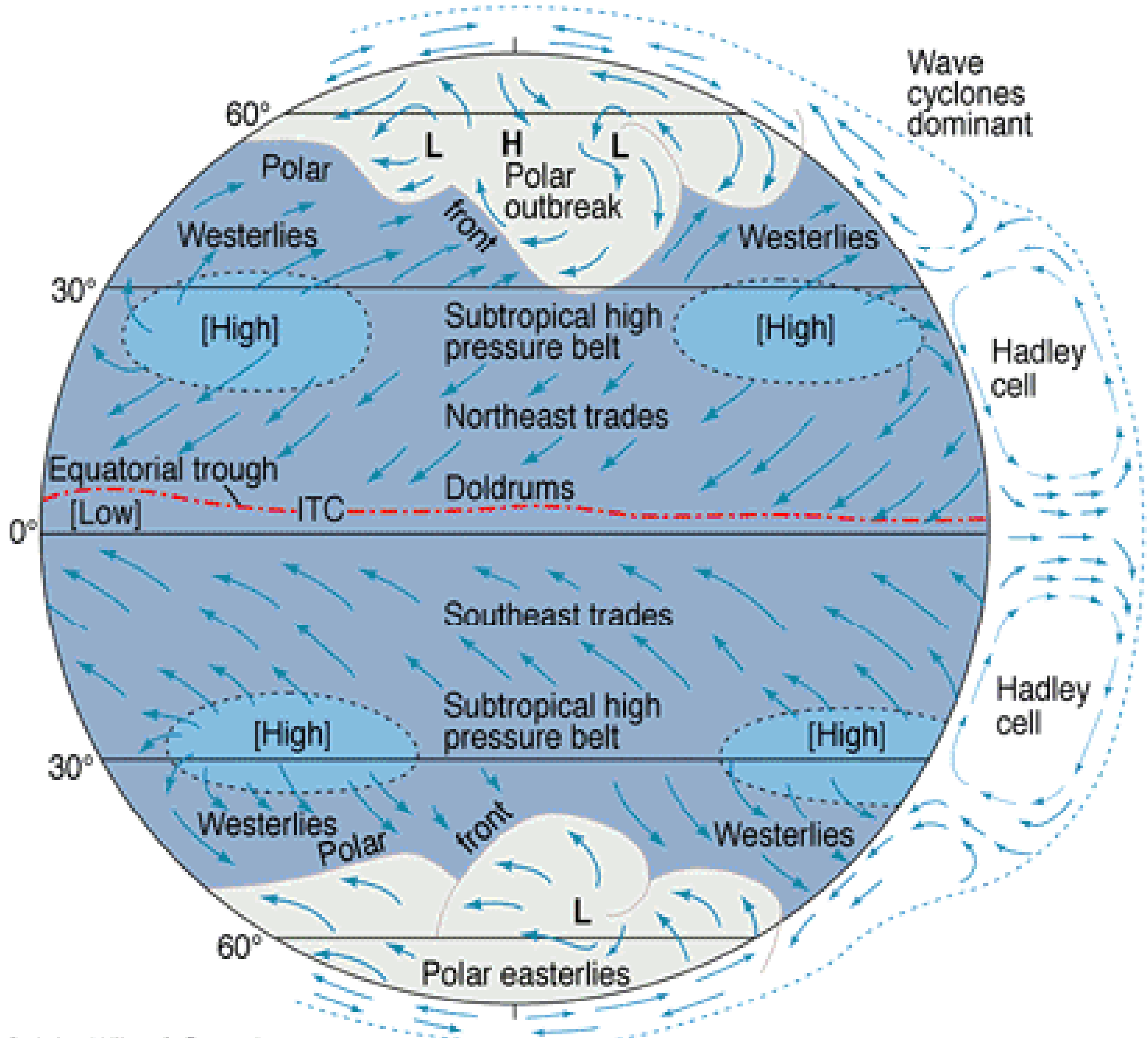
$$\nabla_2 \bullet \overline{\rho \mathbf{V} E} + F_{rad_{TOA}} - \left(F_{rad} + F_{conv} \right)_{surface} = 0$$

What causes lateral enthalpy transport by atmosphere?

1: Large-scale, quasi-steady overturning motion in the Tropics,

2: Eddies with horizontal dimensions of ~ 3000 km in middle and high latitudes


Polar easterlies
Polar front
30°
Westerlies
Hadley cells
NE trade winds
Hadley cells
0°
Equatorial Low
SE trade winds


Wave
cyclones
dominant
60°
L
H
L
Polar
Polar
outbreak
front
Westerlies
Westerlies
30°
[High]
Subtropical high
pressure belt
[High]
Hadley
cell
Northeast trades
Equatorial trough
Doldrums
[Low]
ITC
0°
Southeast trades
Hadley
cell
[High]
Subtropical high
pressure belt
[High]
30°
Westerlies
Polar
front
Westerlies
L
60°
Polar easterlies

First consider a hypothetical planet like Earth, but with no continents and no seasons and for which the only friction acting on the atmosphere is at the surface.

This planet has an exact nonlinear equilibrium solution for the flow of the atmosphere, characterized by:

1. Every column is in radiative-convective equilibrium,

2. Wind vanishes at planet's surface

3. Horizontal pressure gradients balanced by *Coriolis accelerations*

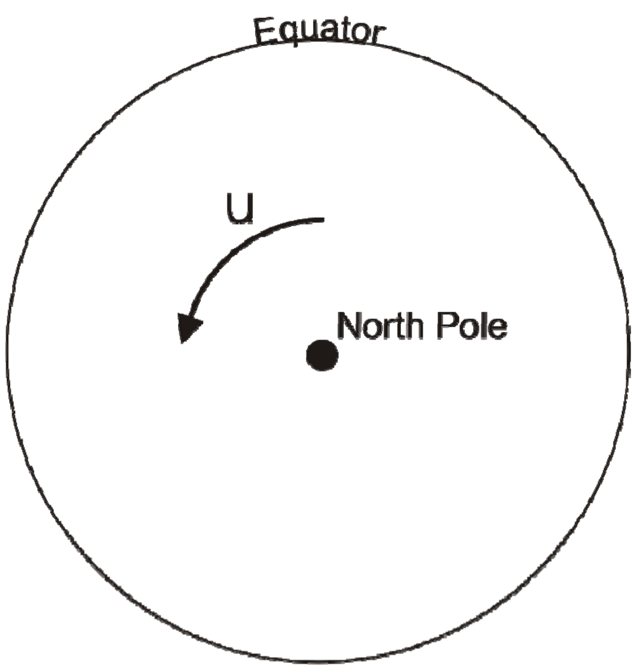
Hydrostatic balance:

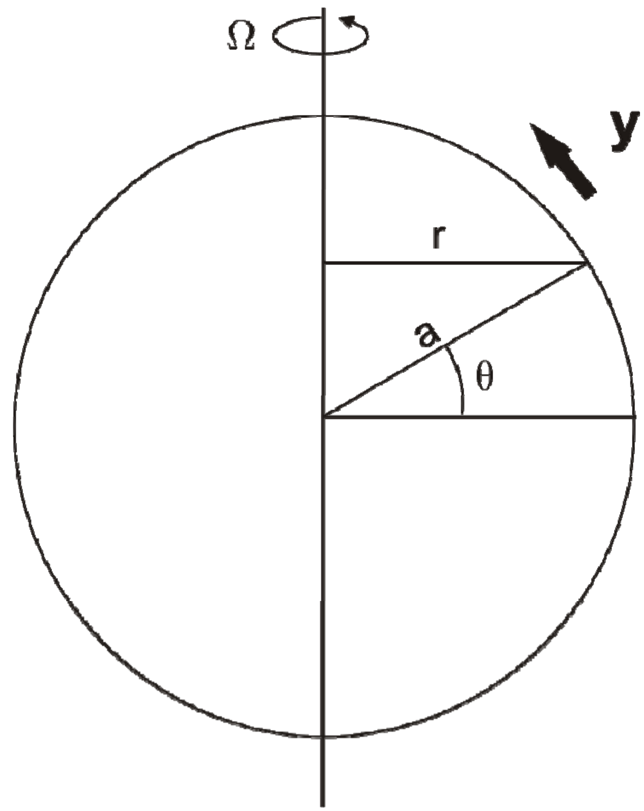
$$\frac{\partial p}{\partial z} = -\rho g$$

Horizontal force balance in *inertial* reference frame:

$$\frac{du}{dt} = -\alpha \frac{\partial p}{\partial x}, \qquad \frac{dv}{dt} = -\alpha \frac{\partial p}{\partial y}$$

Rotating reference frame of Earth:





$$\frac{dv}{dt} = -\alpha\frac{\partial p}{\partial y} - \frac{u^2}{r}\sin\theta$$

$$u = \Omega a\cos\theta + u_{rel}, \qquad r = a\cos\theta$$

$$\rightarrow \frac{dv}{dt} = -\alpha\frac{\partial p}{\partial y} - \underbrace{\Omega^2 a\cos\theta\sin\theta} - 2\Omega\sin\theta u_{rel} - \frac{u_{rel}^2}{a}\tan\theta$$

Bracketed term absorbed into definition of gravity:

$$\frac{dv}{dt} = -\alpha\frac{\partial p}{\partial y} - 2\Omega\sin\theta u_{rel} - \frac{u_{rel}^2}{a}\tan\theta$$

$$\cong -\alpha\frac{\partial p}{\partial y} - 2\Omega\sin\theta u_{rel}$$

$$\equiv -\alpha\frac{\partial p}{\partial y} - fu_{rel}, \quad where\ f \equiv 2\Omega\sin\theta$$

Geostrophic Balance

$$\alpha \frac{\partial p}{\partial y} = -f u_{rel}, \quad where\ f \equiv 2\Omega \sin\theta$$

Similarly,

$$\alpha \frac{\partial p}{\partial x} = f v_{rel}$$

$$\alpha \frac{\partial p}{\partial y} = -fu_{rel} \qquad geostrophic$$

$$\alpha \frac{\partial p}{\partial z} = -g \qquad hydrostatic$$

$Eliminate\ p:$

$$f\frac{\partial u}{\partial z} = -g\left(\frac{\partial \ln(\alpha)}{\partial y}\right) = -g\left(\frac{\partial \ln(T)}{\partial y}\right) \quad Thermal\ wind$$

Zonal wind increases with altitude if temperature decreases toward pole

Two potential problems with this solution:

1. Not enough angular momentum available for required west-east wind,

2. Equilibrium solution may be unstable

Angular momentum per unit mass:

$$M = a\cos\theta\left(\Omega a\cos\theta + u\right)$$

$$a = radius\ of\ earth$$

$$\theta = latitude$$

$$\Omega = angular\ velocity\ of\ earth$$

$$u = west-east\ wind\ speed$$

Violation results in large-scale overturning circulation, known as the Hadley Circulation, that transports heat poleward and drives surface entropy gradient back toward its critical value

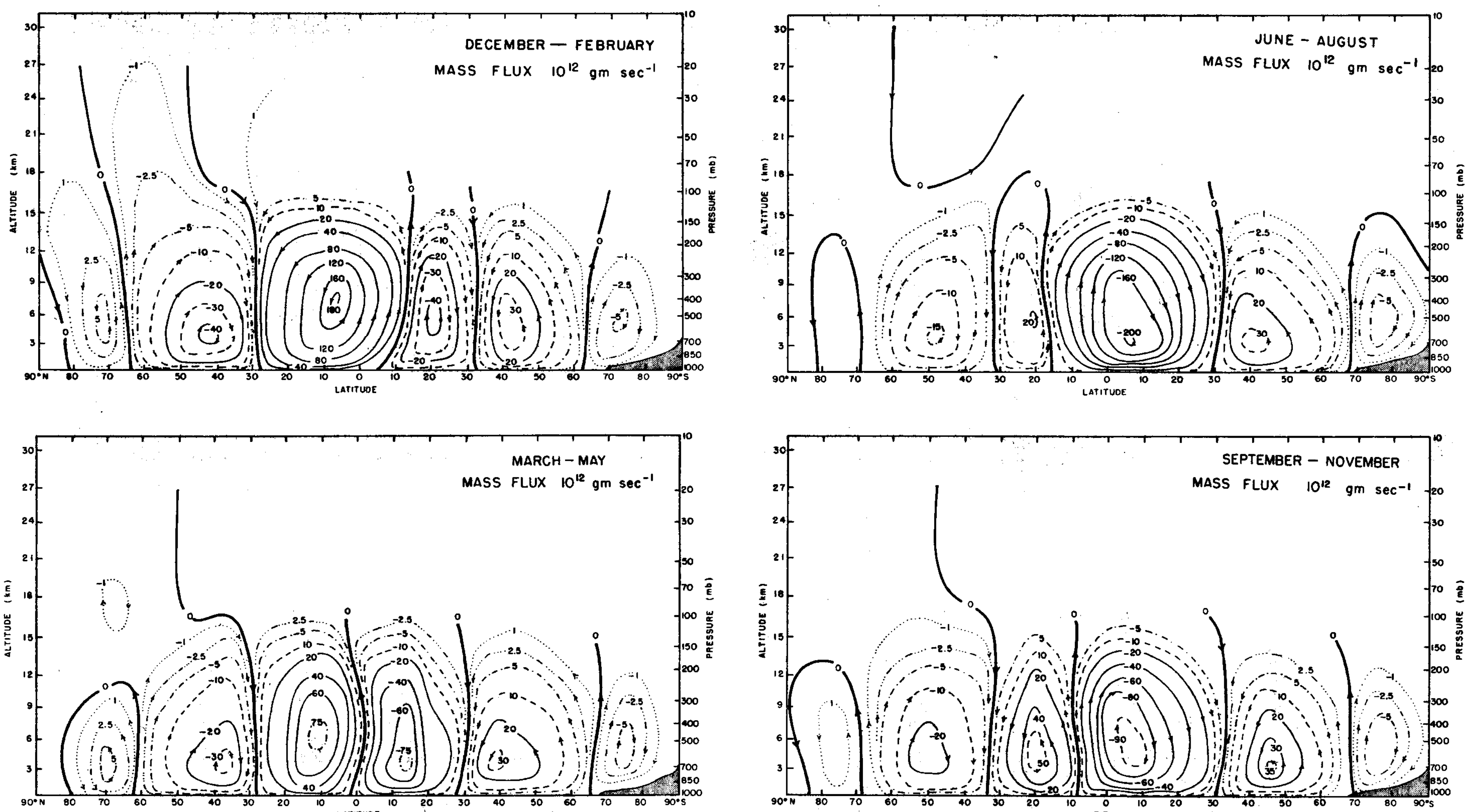


Figure 3.19

Concept of eddy fluxes:

$$\nabla \bullet \overline{\rho \mathbf{V} E} + F_{rad_{TOA}} - \left(F_{rad} + F_{conv}\right)_{surface} = 0$$

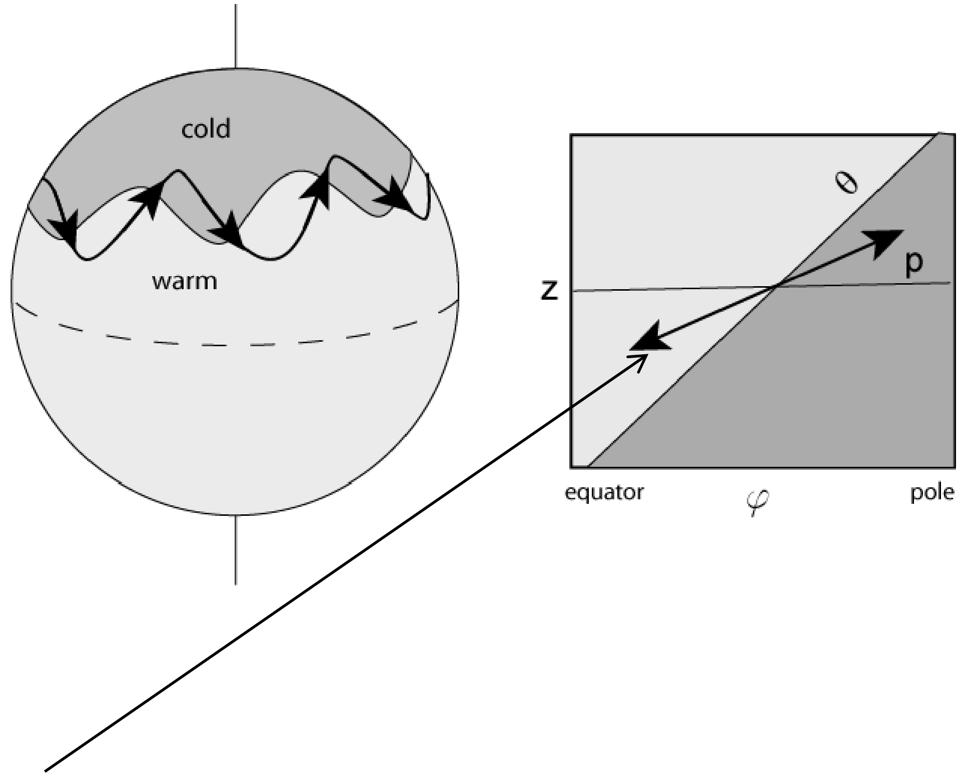
$$\rho \mathbf{V} = \{\rho \mathbf{V}\} + \rho \mathbf{V}',$$

$$E = \{E\} + E',$$

$$where \ \ \{X\} \equiv \frac{1}{2\pi}\int_0^{2\pi} X d\lambda$$

$$\rightarrow \nabla \bullet \left[\overline{\{\rho \mathbf{V}' E'\}} + \overline{\{\rho \mathbf{V}\}\{E\}} \right] + F_{rad_{TOA}} - \left(F_{rad} + F_{conv}\right)_{surface} = 0$$

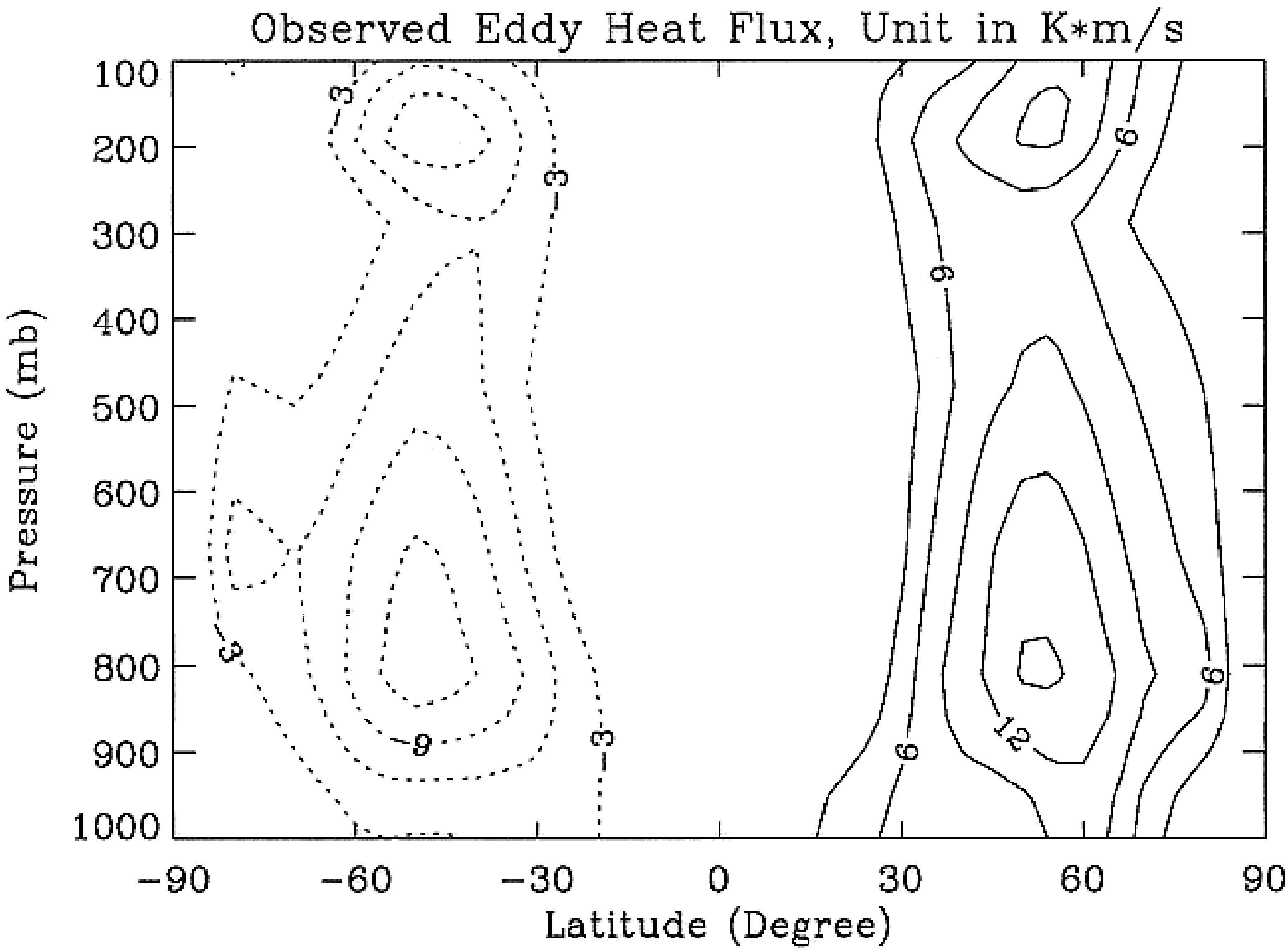
Atmospheric heat transport



Warm air upward and poleward

Cold air downward and equatorward

Efficient poleward heat transport



Observed annual mean eddy heat flux, from Oort and Peixoto, 1983

Eddy heat fluxes not efficient enough to prevent temperature gradients from developing

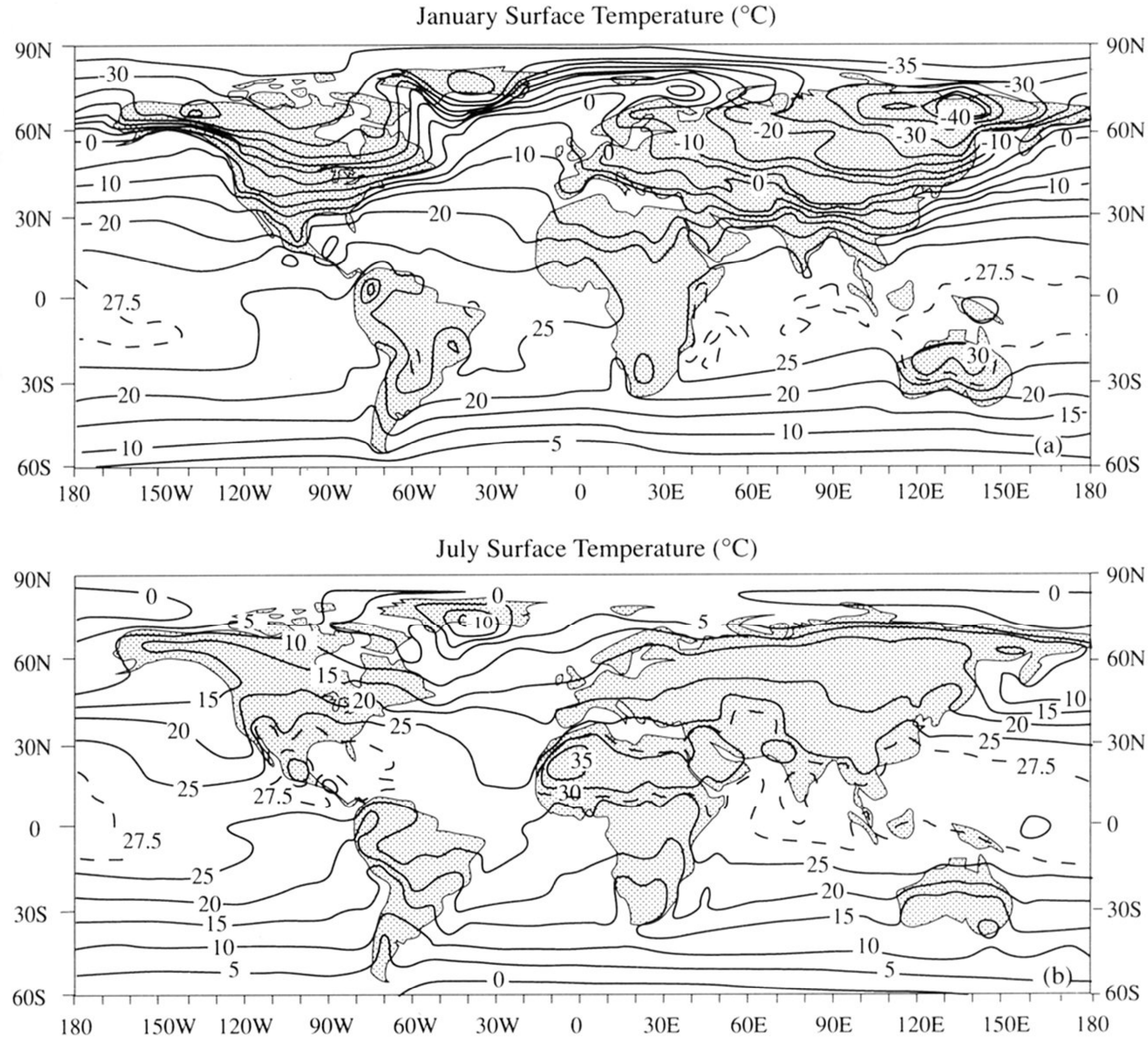
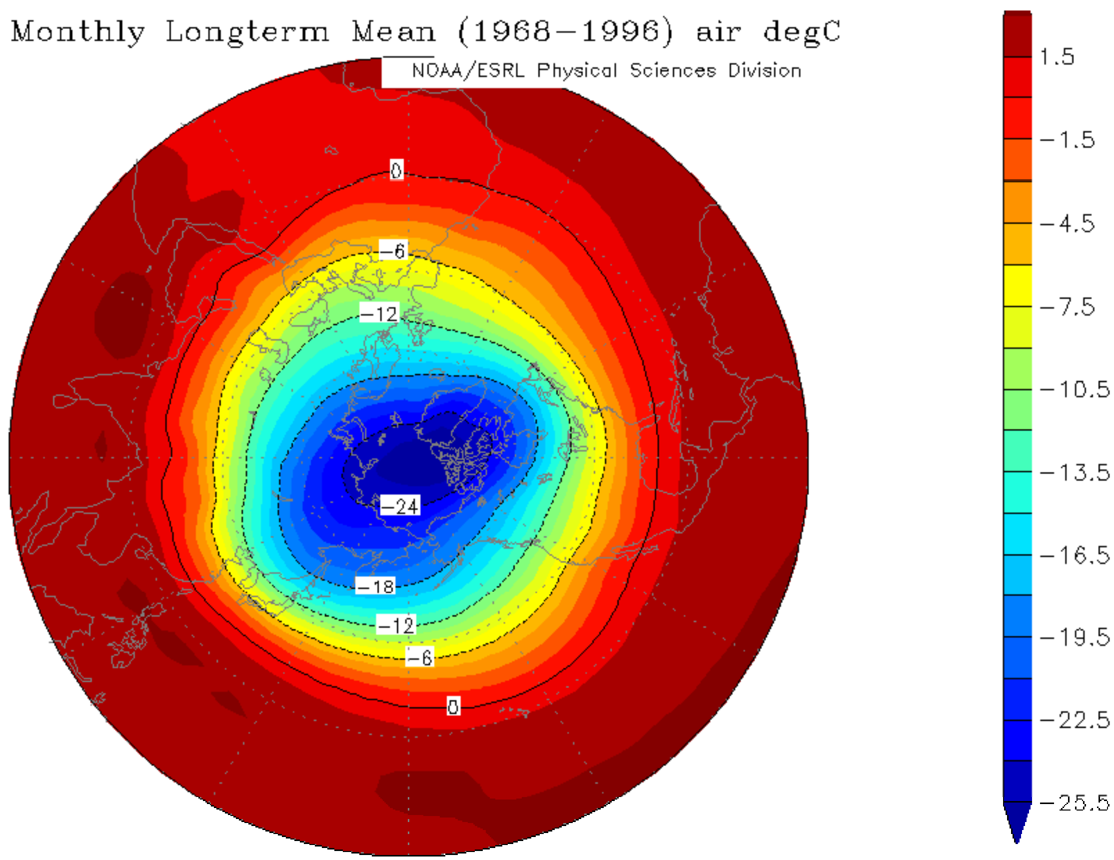


Fig. 1.6 Global map of the (a) January and (b) July surface temperature. [From Shea (1986). Reproduced with permission from the National Center for Atmospheric Research.]

What controls the temperature gradient in middle and high latitudes?

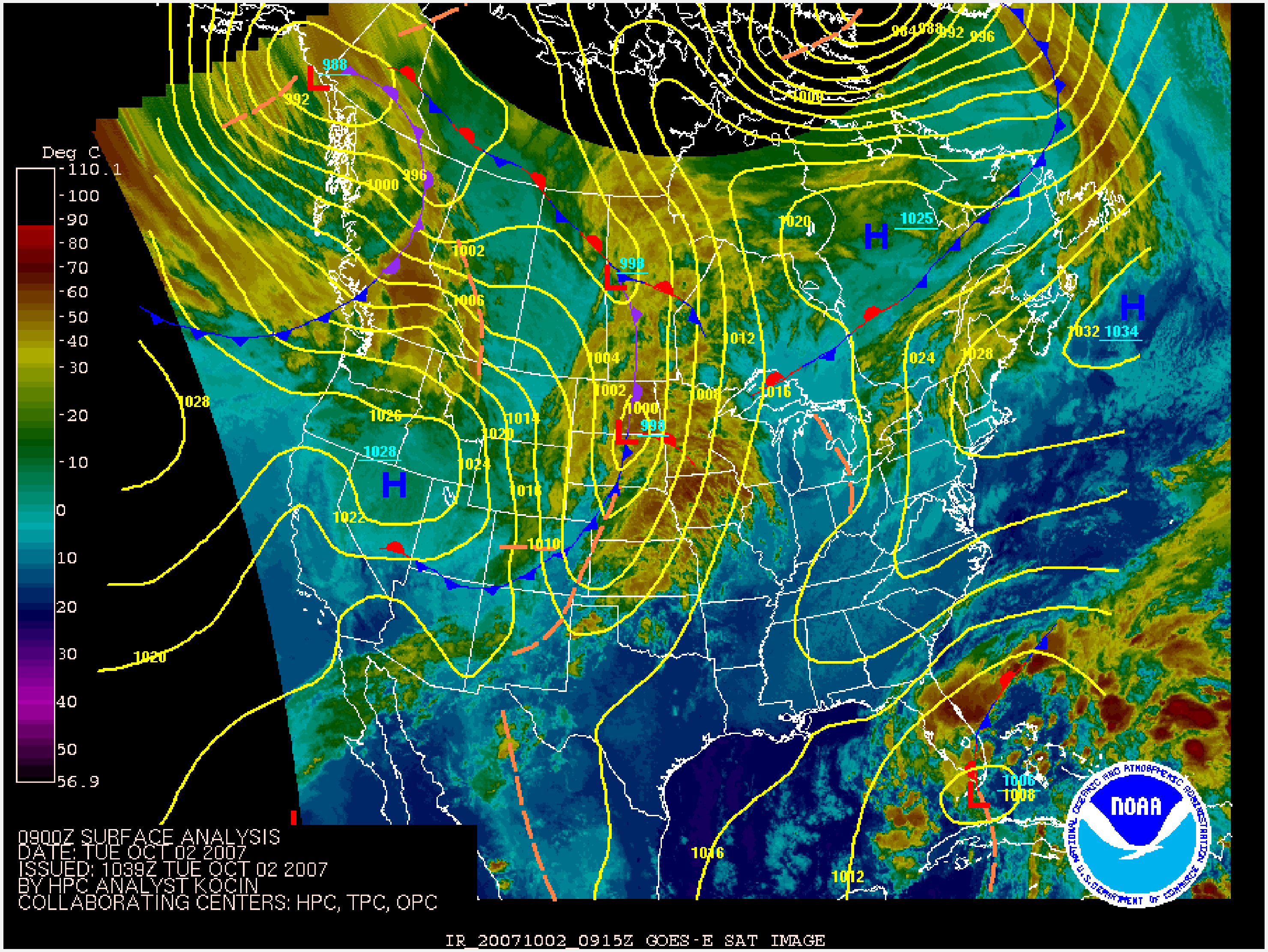


Annual mean temperature, northern hemisphere, about 4 km altitude

Issues

- Temperature gradient controlled by eddies of horizontal dimensions ~ 3000 km
- Familiar highs and lows on weather maps
- Eddy physics not simple
- Concept of criticality does not apply...critical T gradient = 0 (not observed)
- While eddies try to wipe out T gradient, they do not succeed

Example of surface pressure distribution



The Oceans and Climate

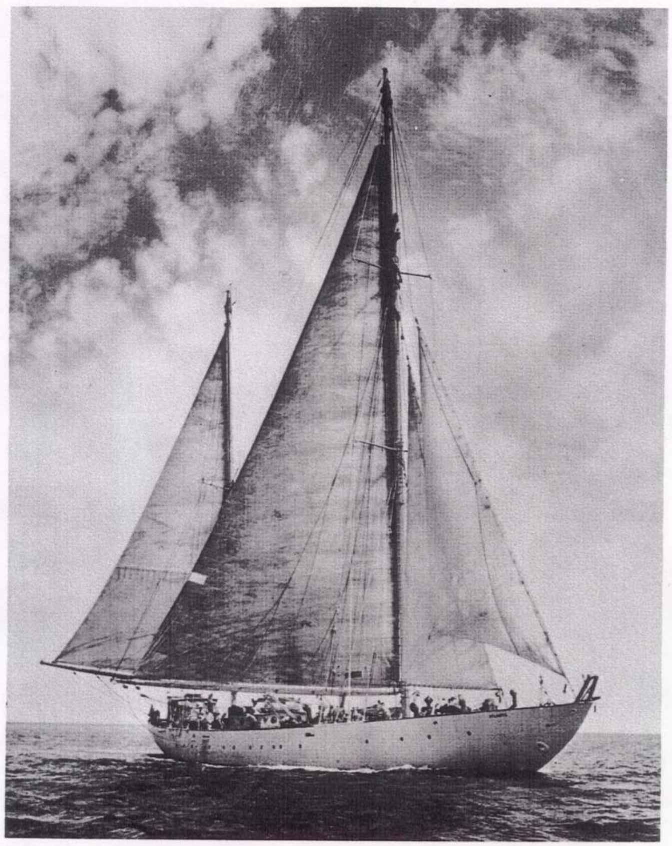
Major elements of the connection of the oceans with the climate system:

(1) The ocean has almost all of the water on the planet
(2) It has an immense heat capacity compared to the atmosphere
(3) It retains a memory of past disturbances that can extend to many thousands of years

Consequently and in addition:
(3) It exchanges energy with the atmosphere (heat, moisture) and *transports it* in very large amounts
(4) It absorbs, stores, and ejects carbon dioxide in very large amounts
(5) It is the site of a large fraction of the biological activity on Earth.
(6) It is a major component of the biogeochemical cycles (nitrogen, phosphorous, etc.) on Earth
(7) When frozen, it can undergo a very large albedo change

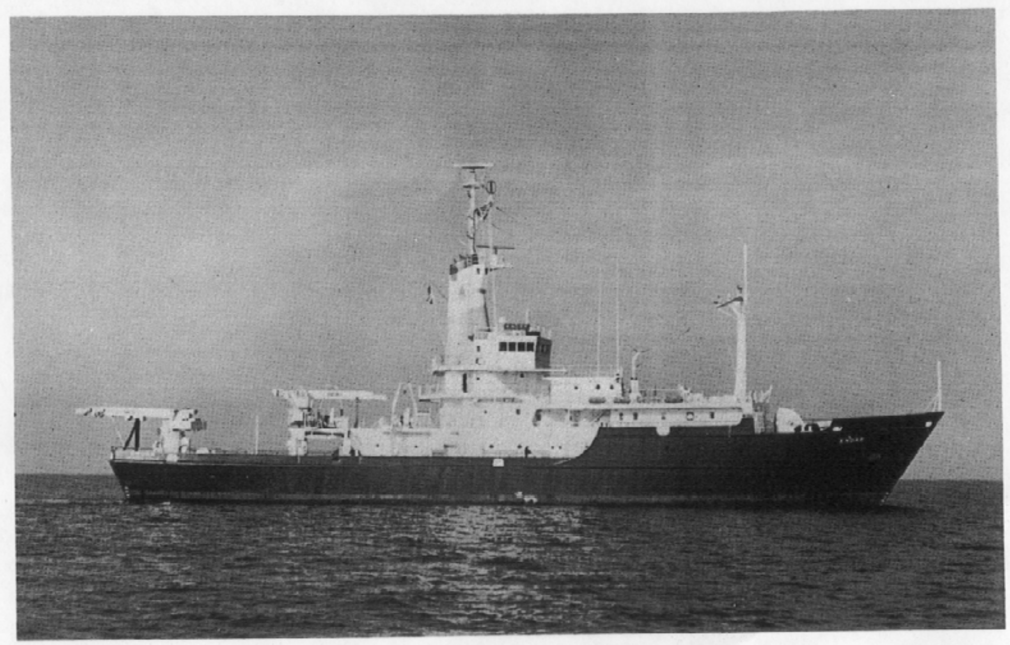
- Similarities to atmosphere:
  - Global-scale fluid on a rotating earth
  - Governing equations nearly identical
- Major differences:
  - Continental barriers to east-west motions
  - Ocean virtually opaque to radiation
  - The ocean is heated (and cooled) at its upper surface
  - No equivalent in the ocean of moist convection (although salinity introduces some analogous issues)
  - Much more difficult to observe the ocean

How does one determine what the ocean does? Dynamically, the ocean is much like the atmosphere, but the problem of observing it is radically different, and the difference in observational technologies has strongly influenced inferences about the ocean circulation. It is essential to understand what is really known, and what is surmised from limited data.



R/V *Atlantis I*

A 142-foot steel ketch built in Copenhagen in 1931 for Woods Hole Oceanographic Institution. In 1964 *Atlantis* was sold to the Consejo Nacional de Investigaciones Cientificas y Technicas, Buenos Aires, Repulica de Argentina, where she was renamed *El Austral.*



R/V *Knorr* (AGOR 15)

Built in 1970 by Defoe Shipbuilding Company of Bay City, Michigan, the *Knorr*, a sister ship of the R/V *Melville* (AGOR 14), features the use of cycloidal propellers for propulsion, position keeping, and maneuverability. The Navy has assigned her to Woods Hole Oceanographic Institution.

Length: 245' Displacement: 2,075 tons

Hanging a Nansen bottle on a wire before lowering it into the sea.

Effect of Long Memory

Suppose that the ocean is just a non-moving slab that stores heat coming in from the atmosphere and gives it back when required. Such a system might be modeled by a simple equation like

$$\kappa \frac{\partial T}{\partial t} = q$$

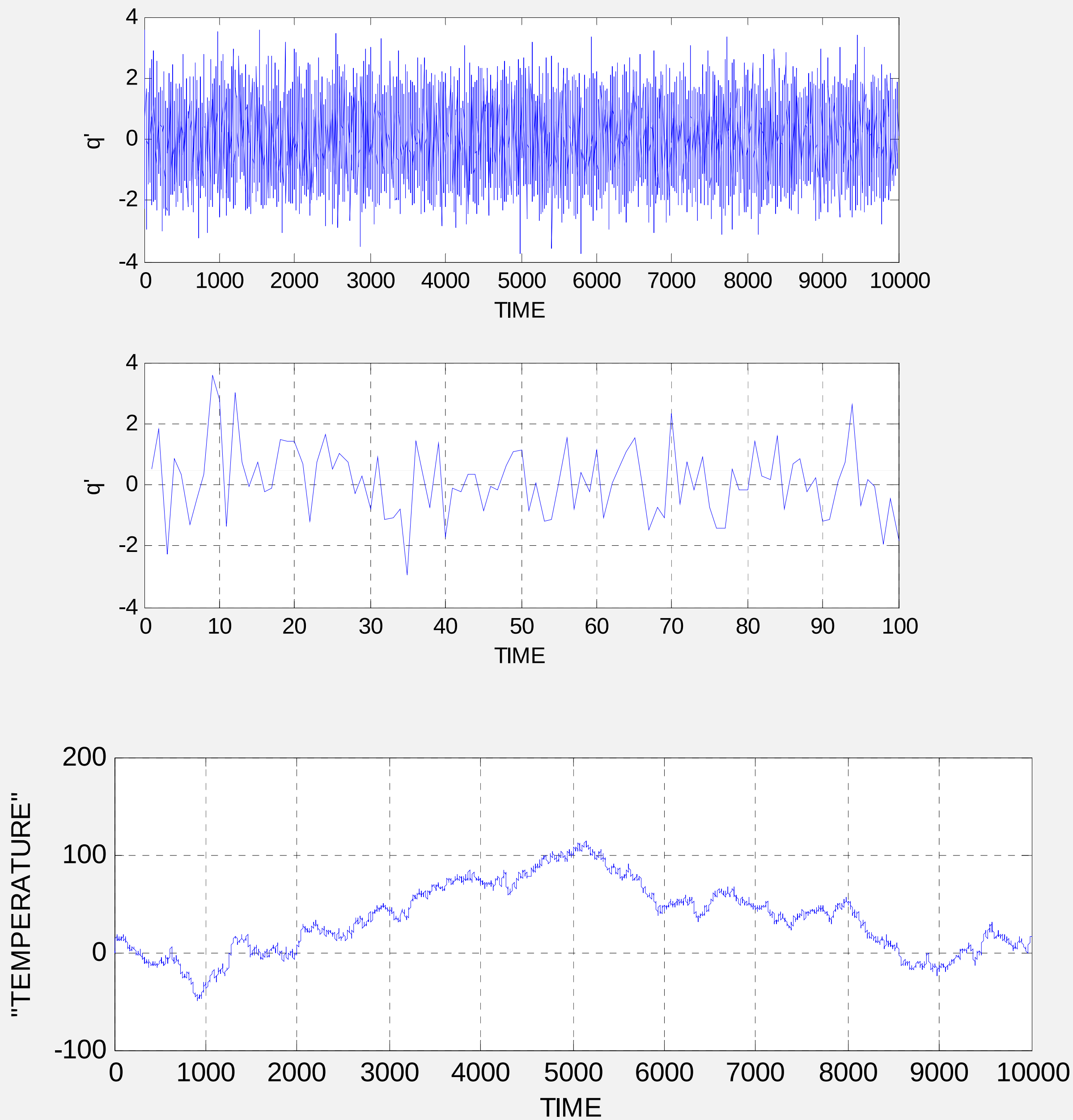
Re-write it as a simple finite difference,

$$T((n+1)\Delta t) - T(n\Delta t) = \frac{\Delta t}{\kappa} q(n\Delta t)$$

or,

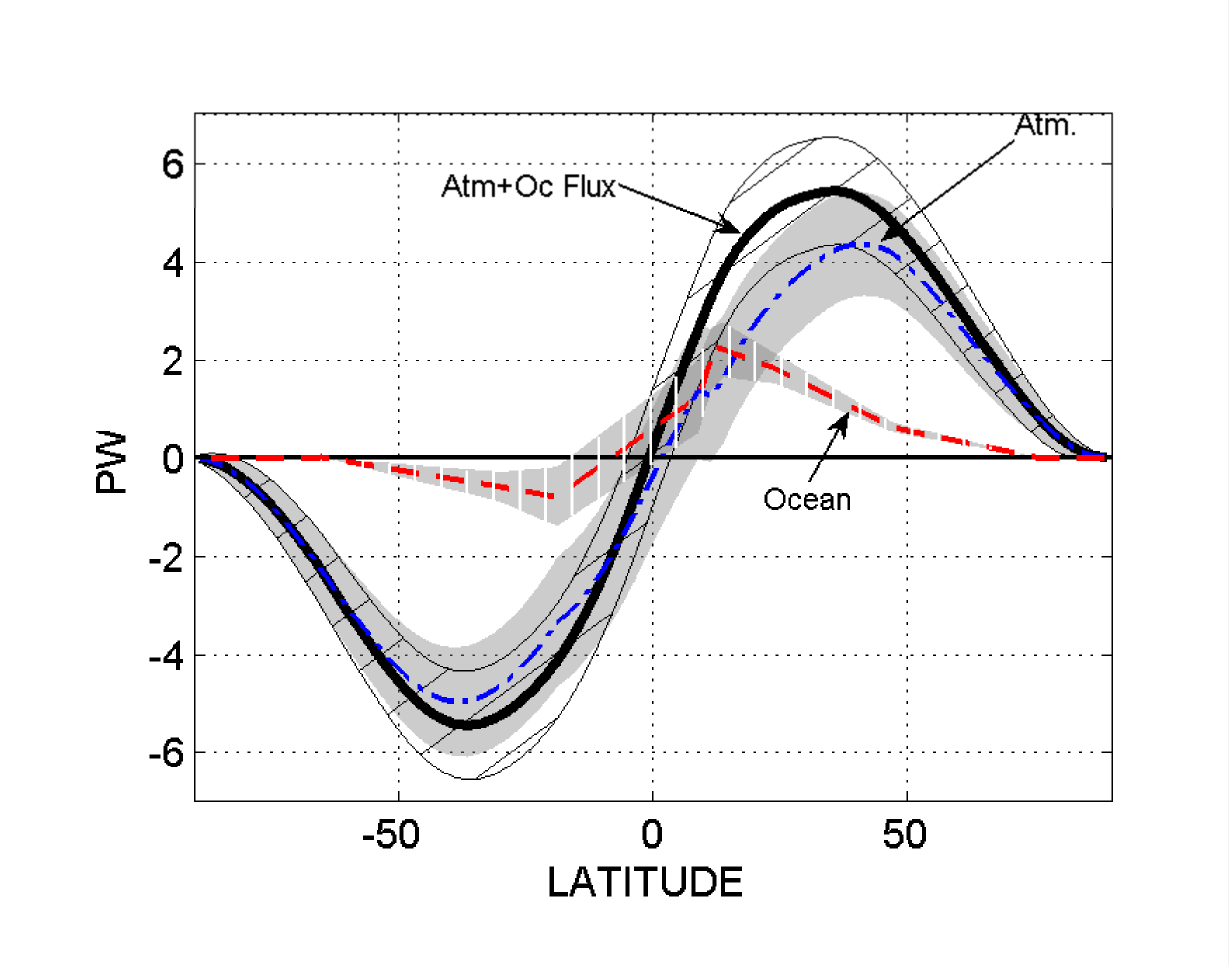
$$T((n+1)\Delta t) = T(n\Delta t) + q'(n\Delta t)$$

where we think of q' as being "weather noise" and treat it as completely random. Easy then to compute $T((n+1)\Delta t)$ from $T(n\Delta t)$.



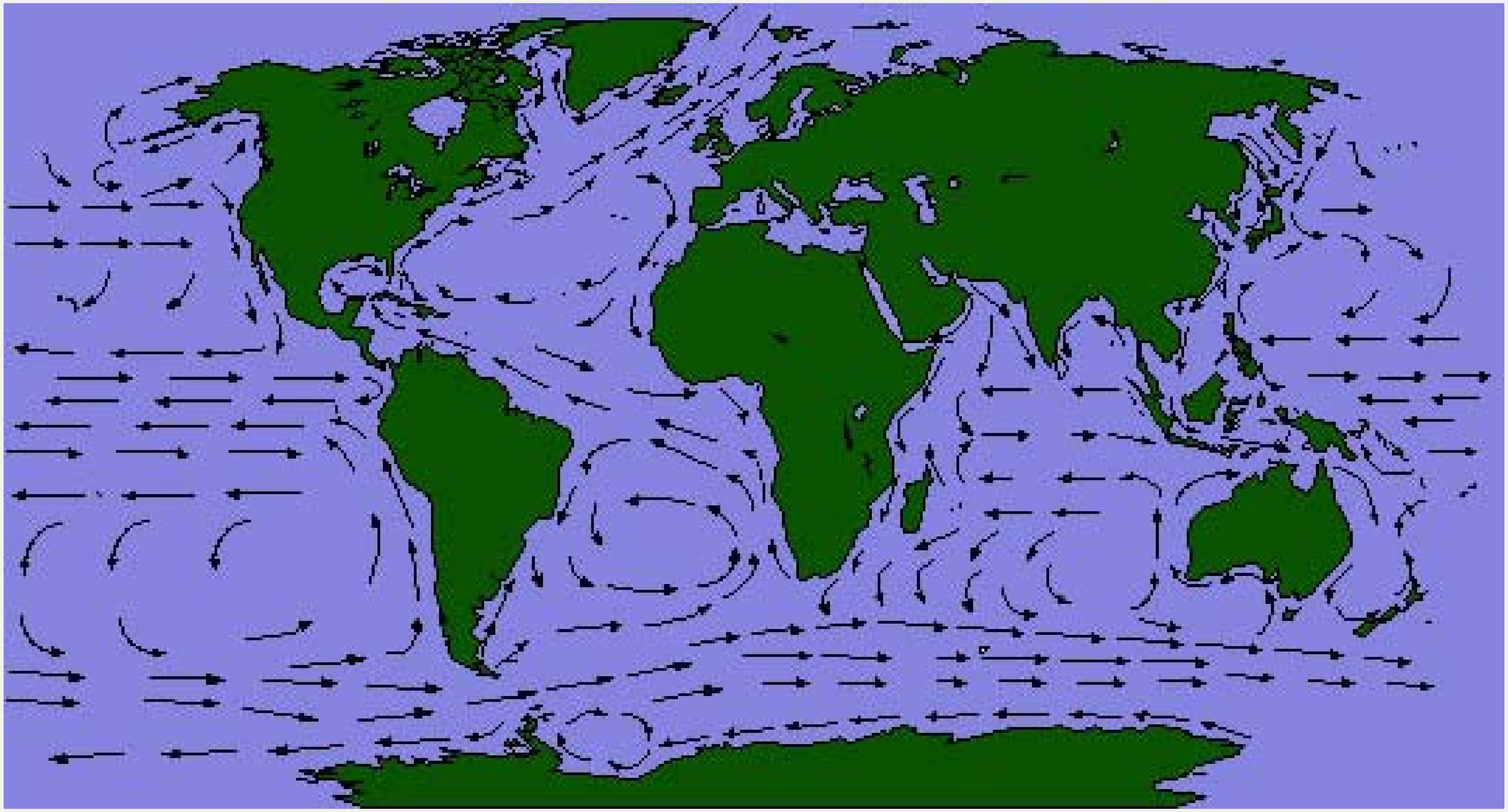
q'
TIME
q'
TIME
"TEMPERATURE"
TIME

1 PW= 1 PetaWatts=10^{15} W



Example of what the ocean does---meridional transport of heat away from the equator. Does this change? Can it change?

Wind-Driven Circulation



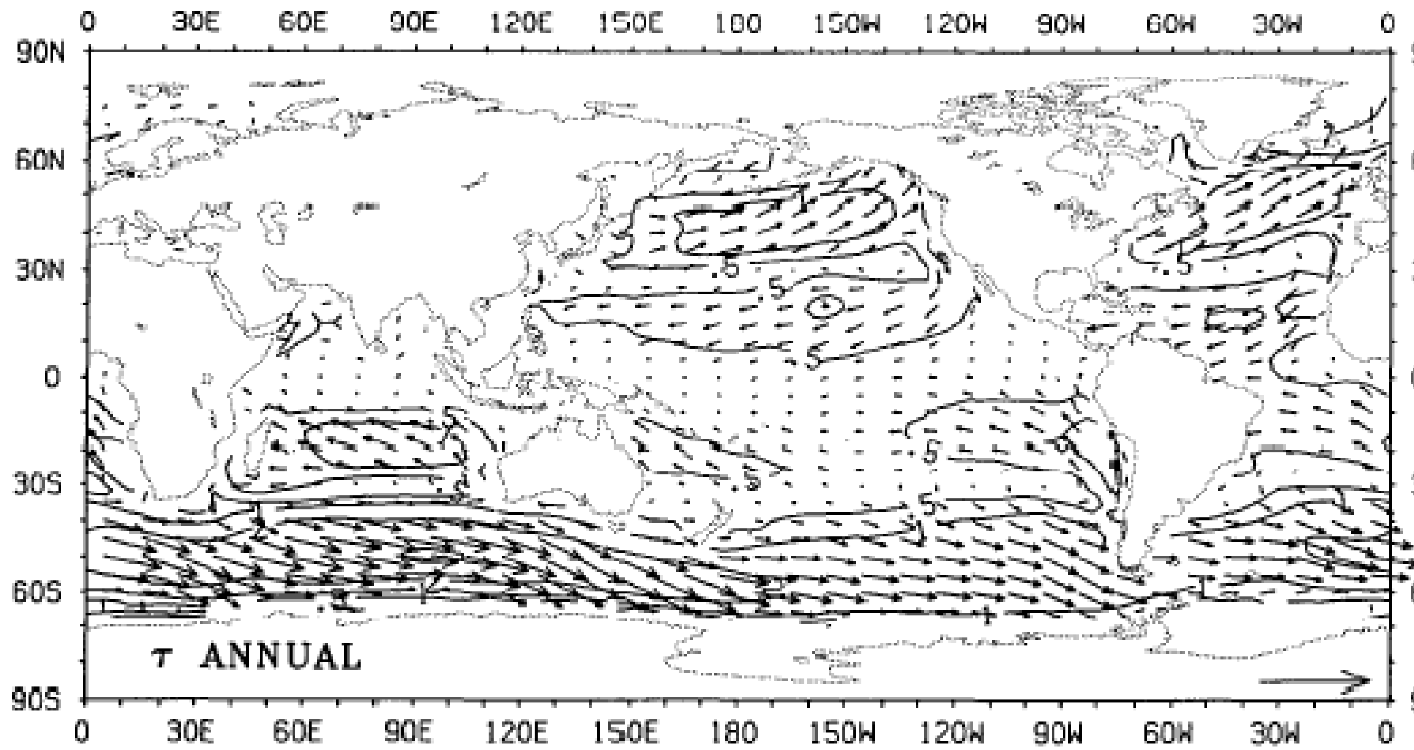
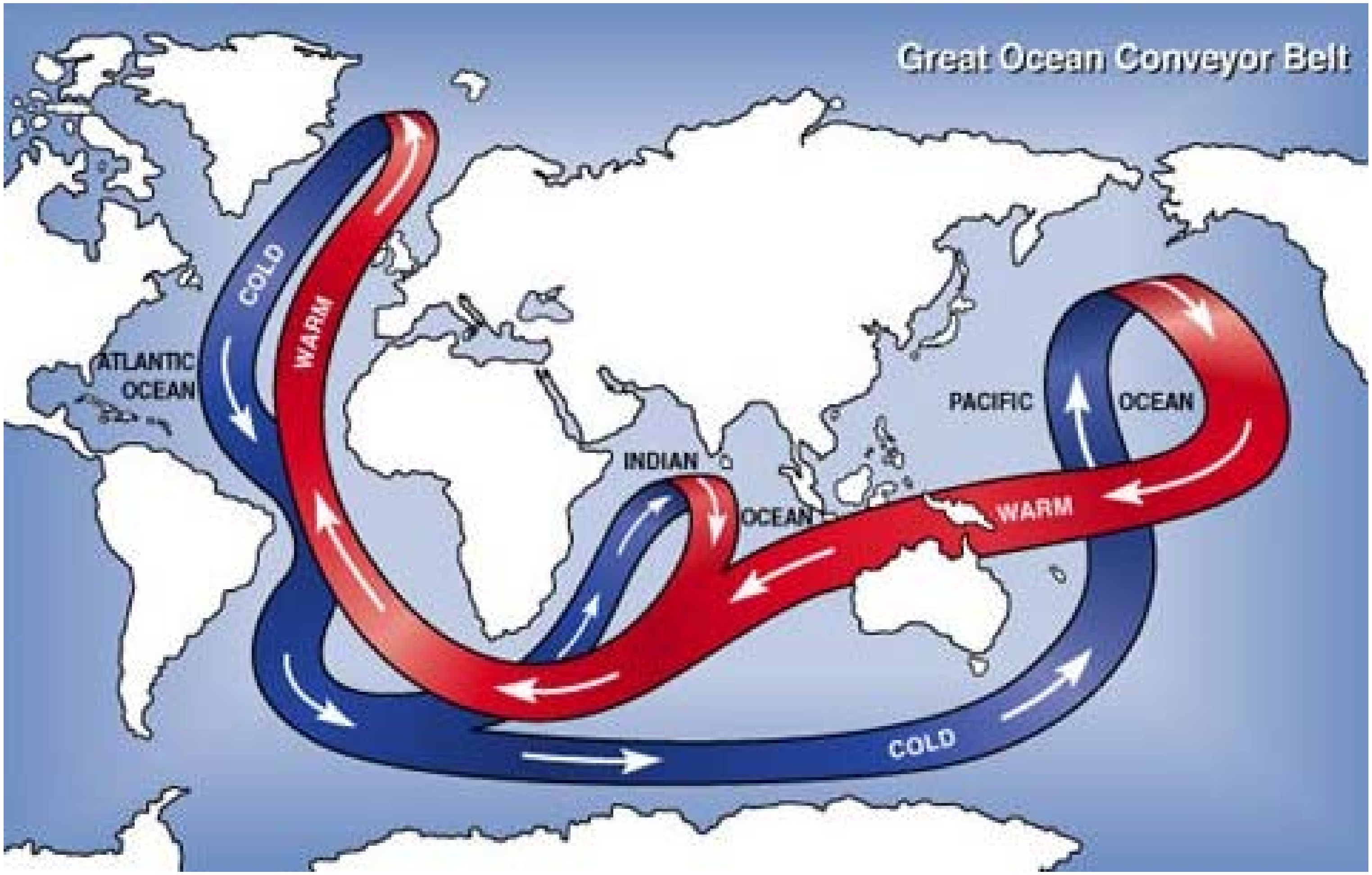
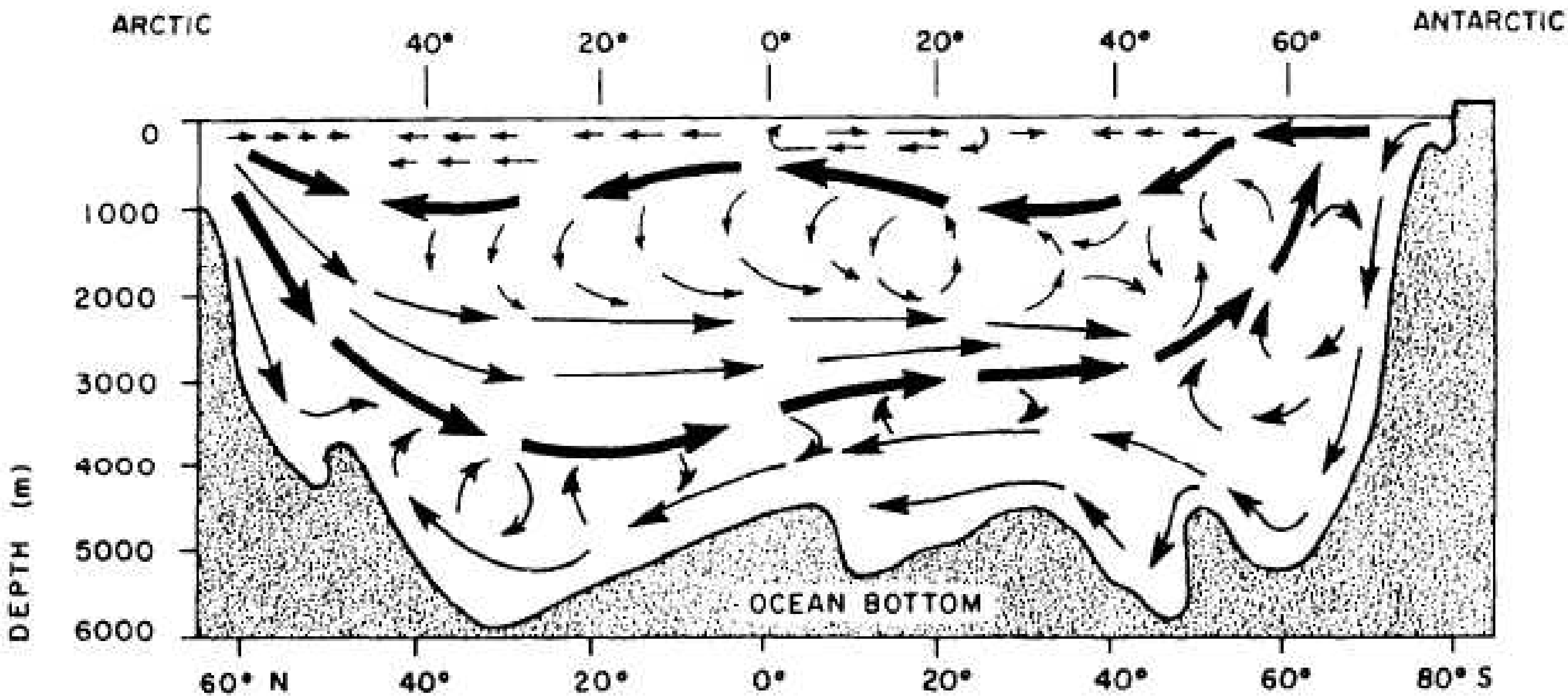


FIG. 1. Annual mean wind stress over the global oceans depicted as vectors. The arrow at bottom right corresponds to 5 dyn cm^{-2} and contours of magnitude of 0.5, 1, 2 and 3 dyn cm^{-2} are plotted.

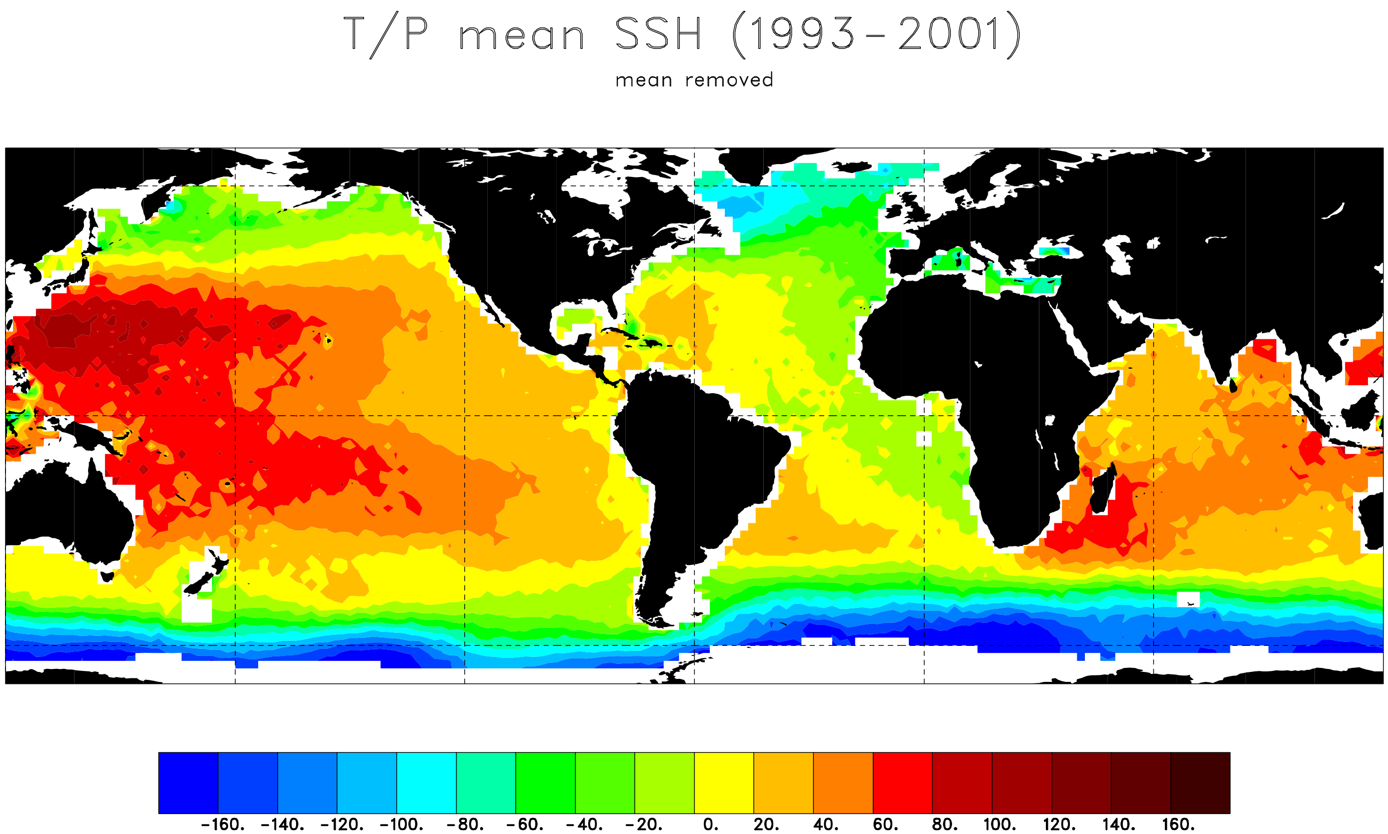
Climatological wind field over the ocean.
Trenberth et al. (1990, JPO) from ECMWF. Patterns of dominant westerlies and easterlies. Low-latitude easterlies are called "trade-winds".

Deep Overturning



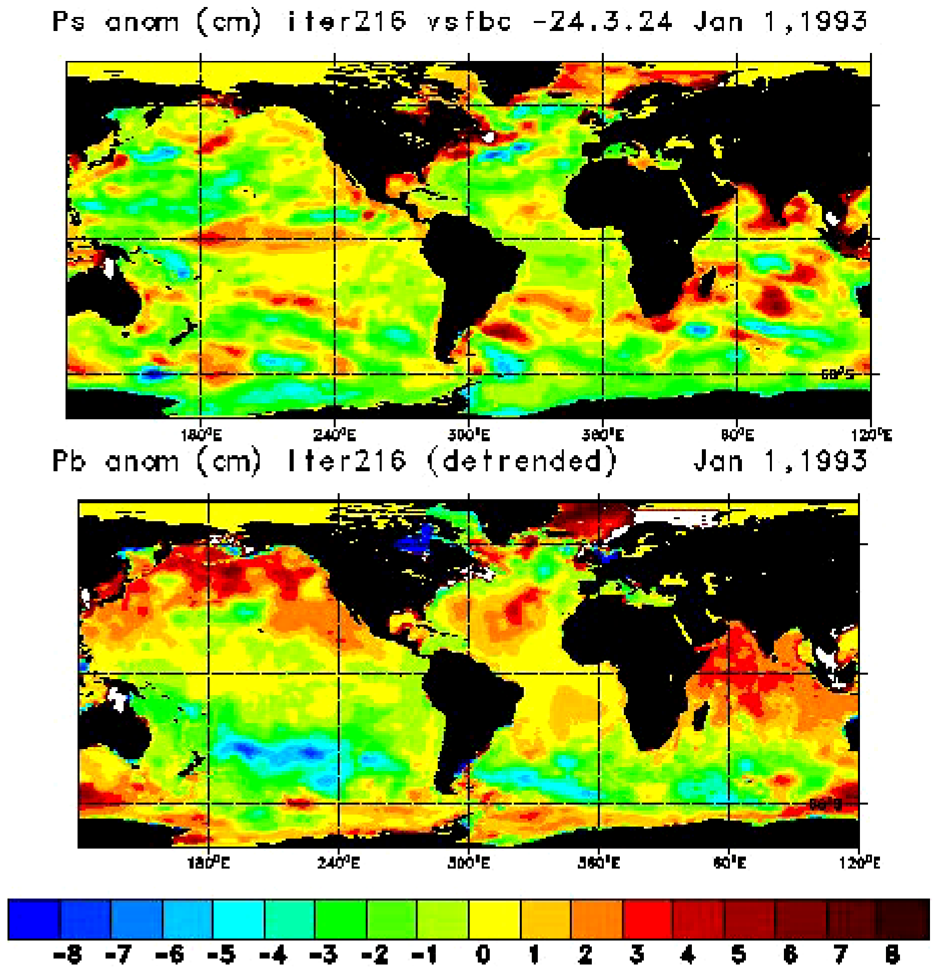

ARCTIC
ANTARCTIC
40°
20°
0°
20°
40°
60°
0
1000
2000
3000
4000
5000
6000
DEPTH (m)
OCEAN BOTTOM
60° N
40°
20°
0°
20°
40°
60°
80° S

By 1992 it had become technically possible to measure the absolute topography of the seasurface from space using orbiting altimeters. There are still major problems, many having to do with the inability to determine the shape of the Earth accurately enough, but the system is adequate to show the major inferred oceanic features:



Heights are in centimeters relative to a surface defined to be a gravitational equipotential.

An animation. See http://ecco-group.org

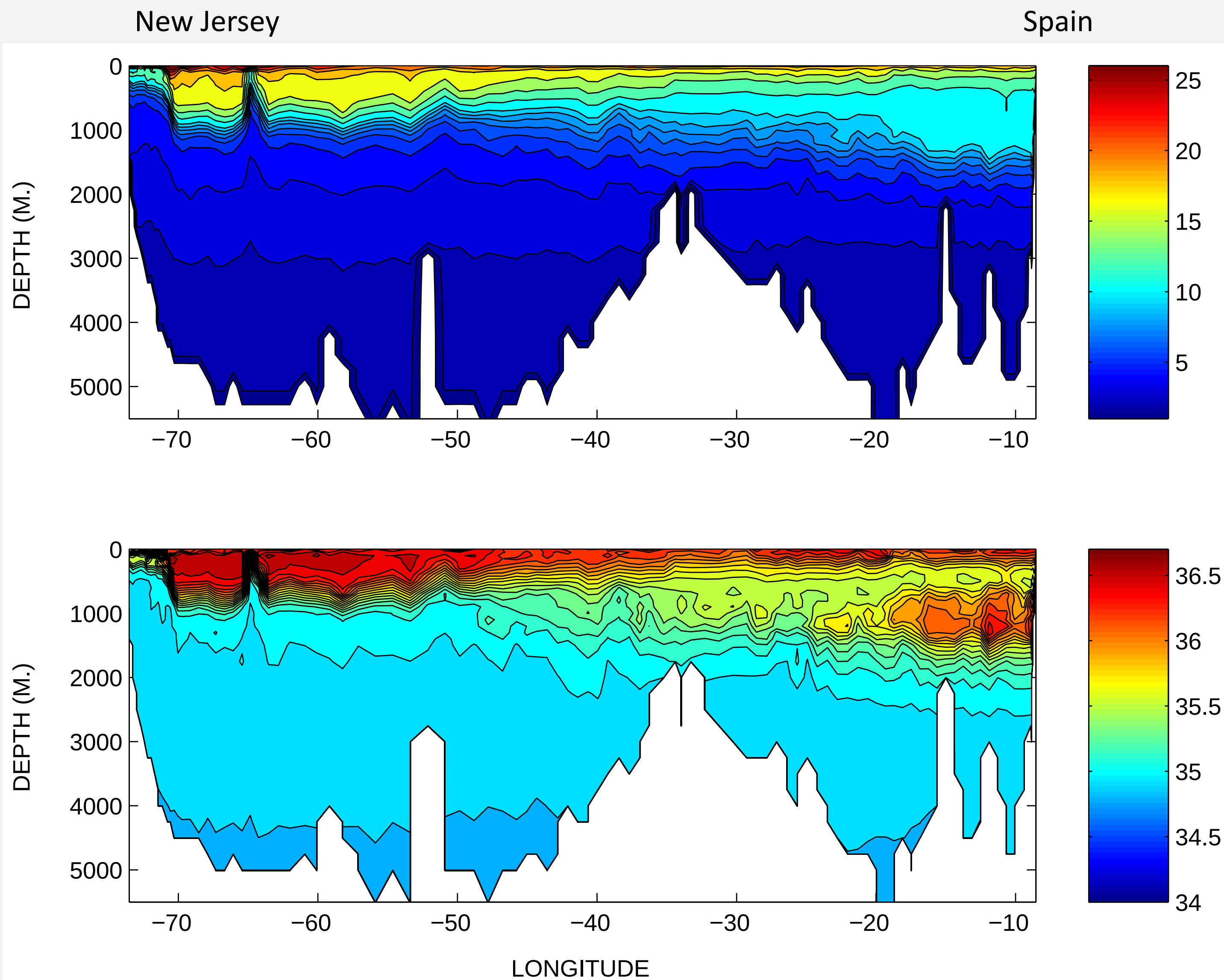


surface elevation, cms

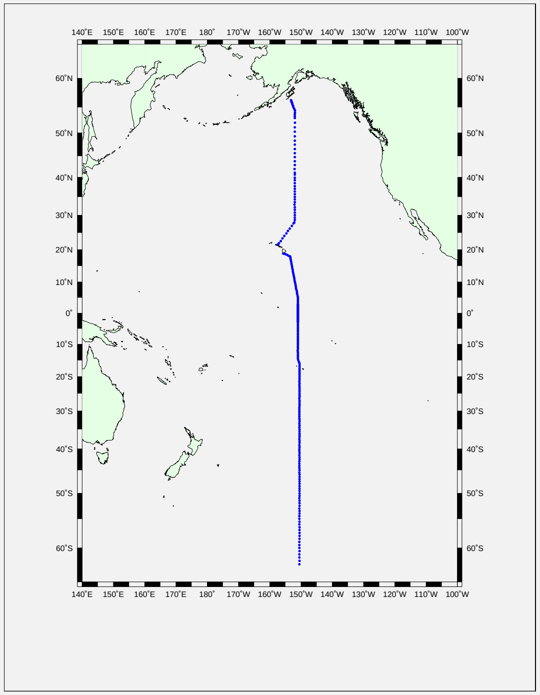
bottom pressure, cms

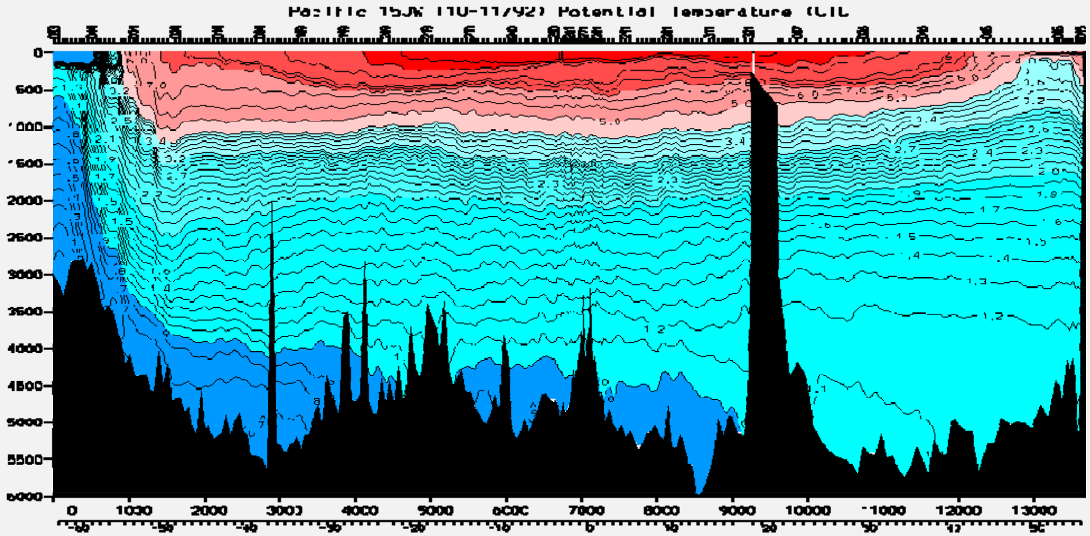
From Estimating the Circulation and Climate of the Ocean (ECCO) consortium results (see posters on 15th floor).

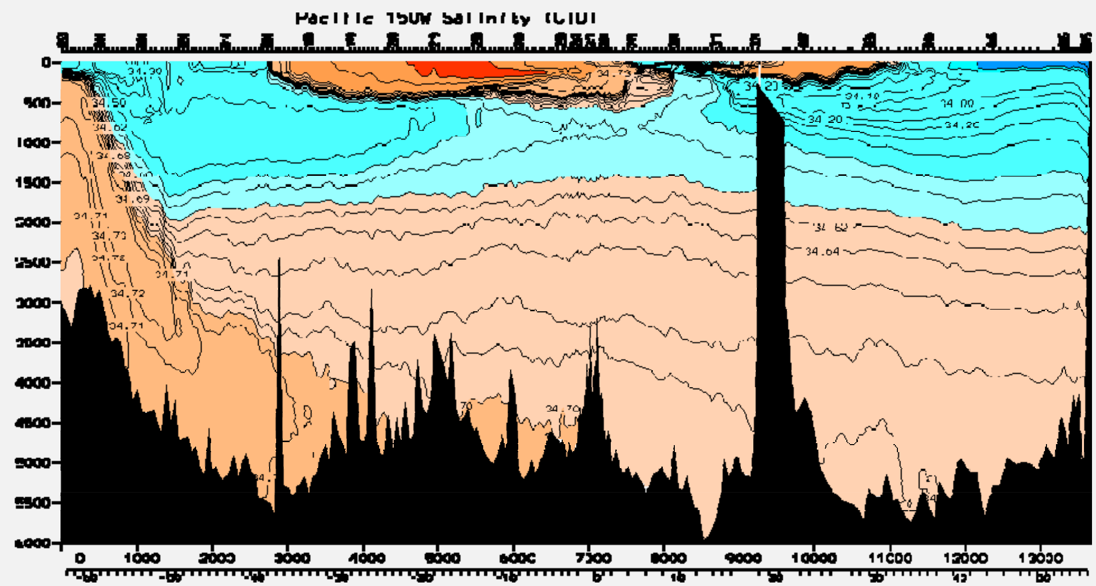
36N potential temperature (°C) and salinity. Roemmich & Wunsch, 1985



Salinity is formally dimensionless, but is very close to being mass in parts/thousand.

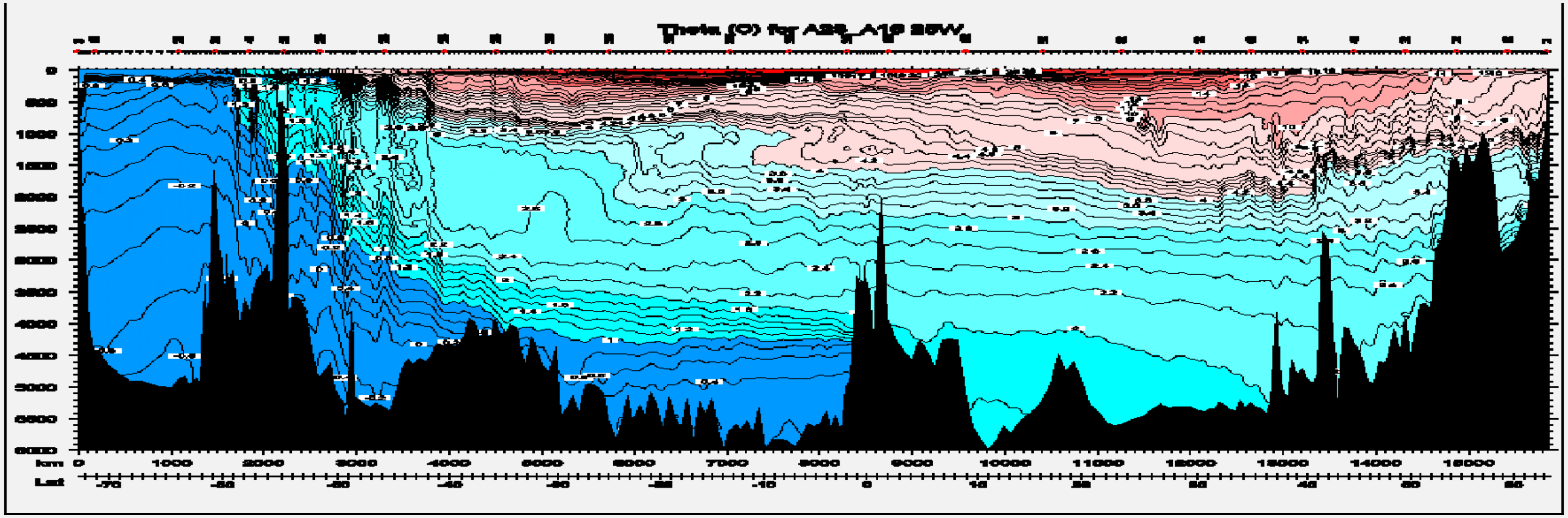


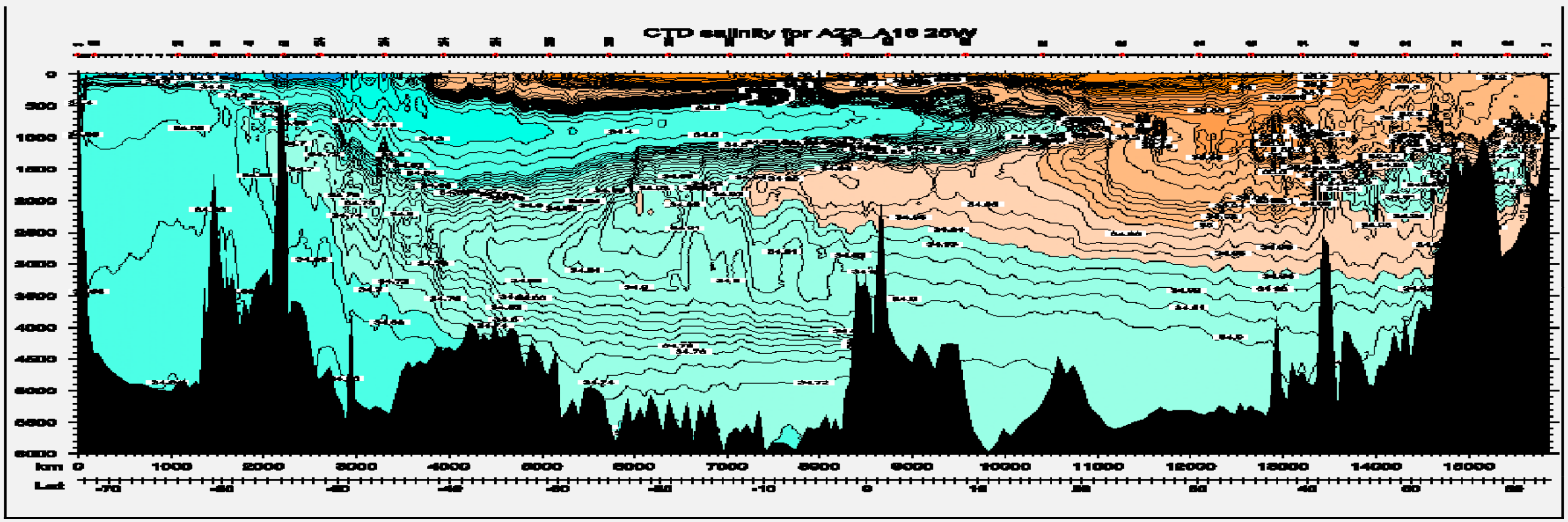


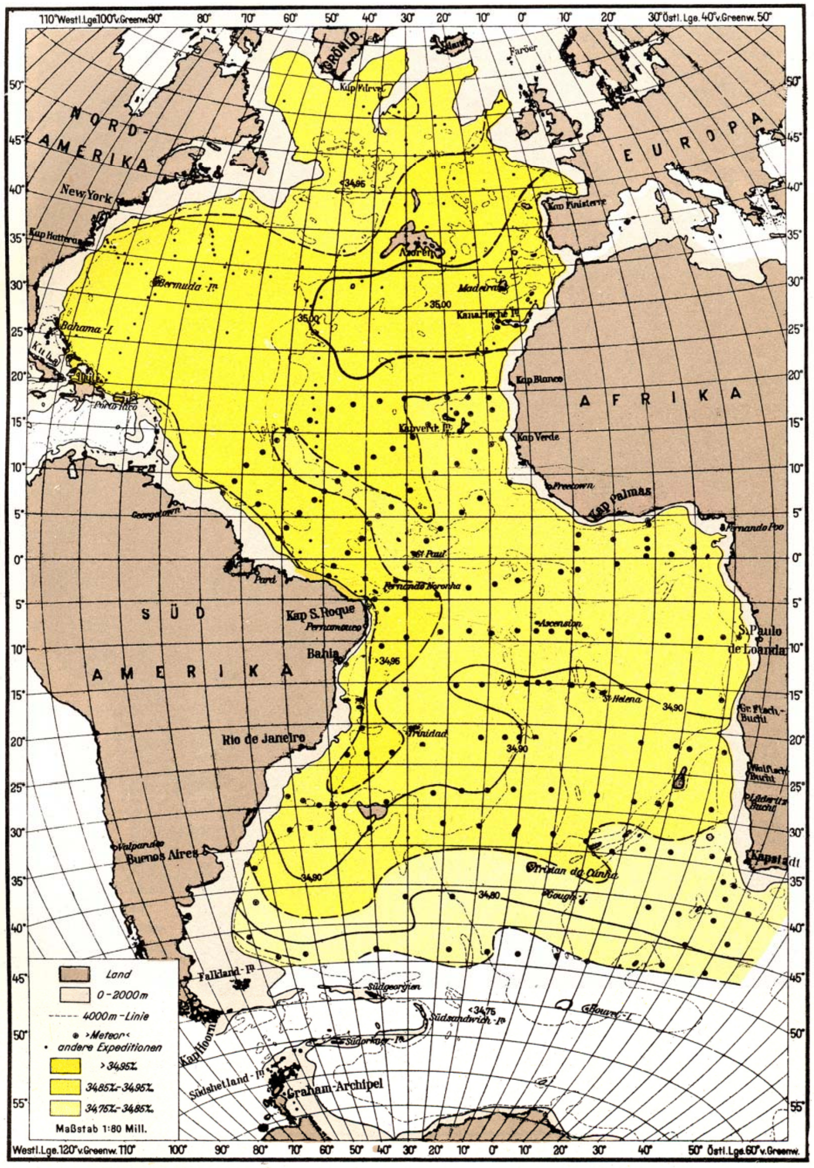


From the World Ocean Circulation Experiment (WOCE). Global coverage from about 1992-1997.

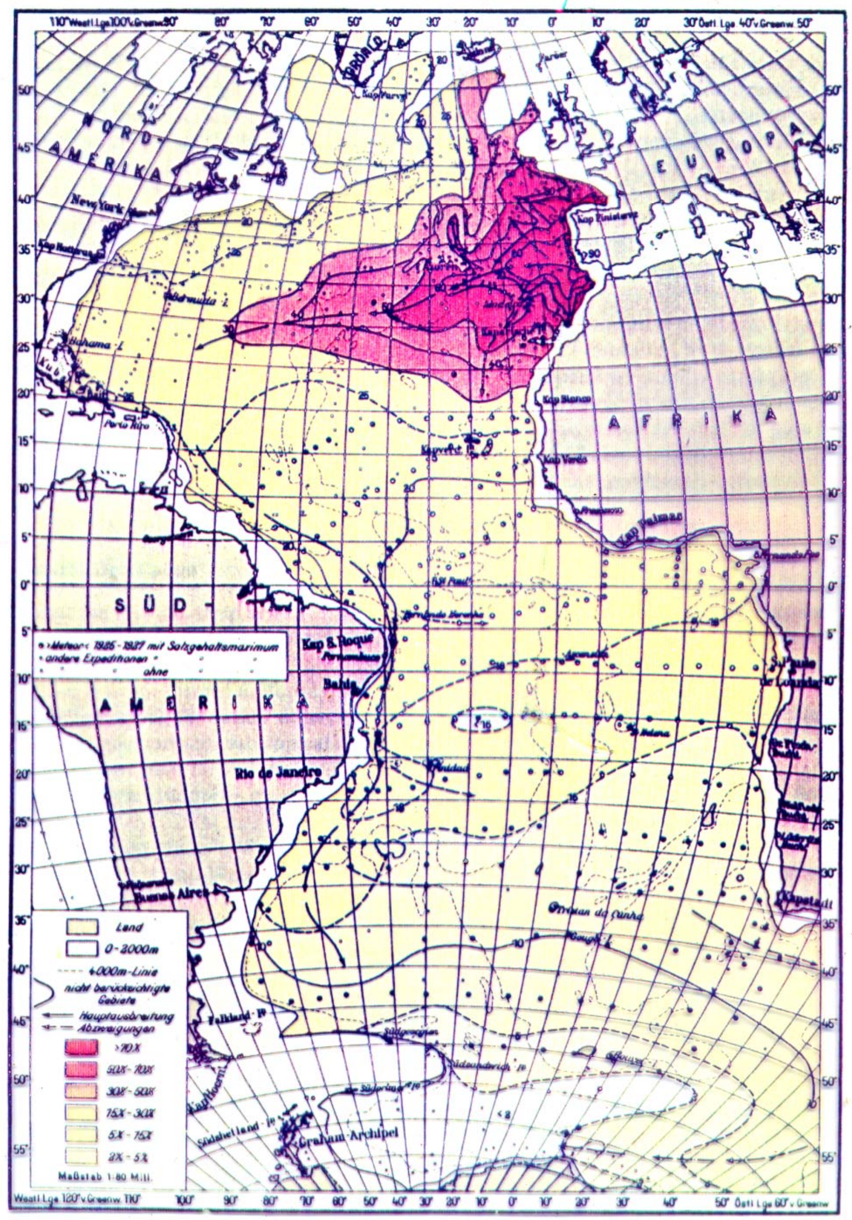
A6, A23 30W in N. Atlantic Pot. Temp and Salinity







Salinity (‰) of the core layer (intermediate oxygen maximum) of the Middle North Atlantic Deep Water.



Spreading of the Upper North Atlantic Deep Water in the core layer (intermediate salinity maximum), represented by the percentage of the Mediterranean component M.