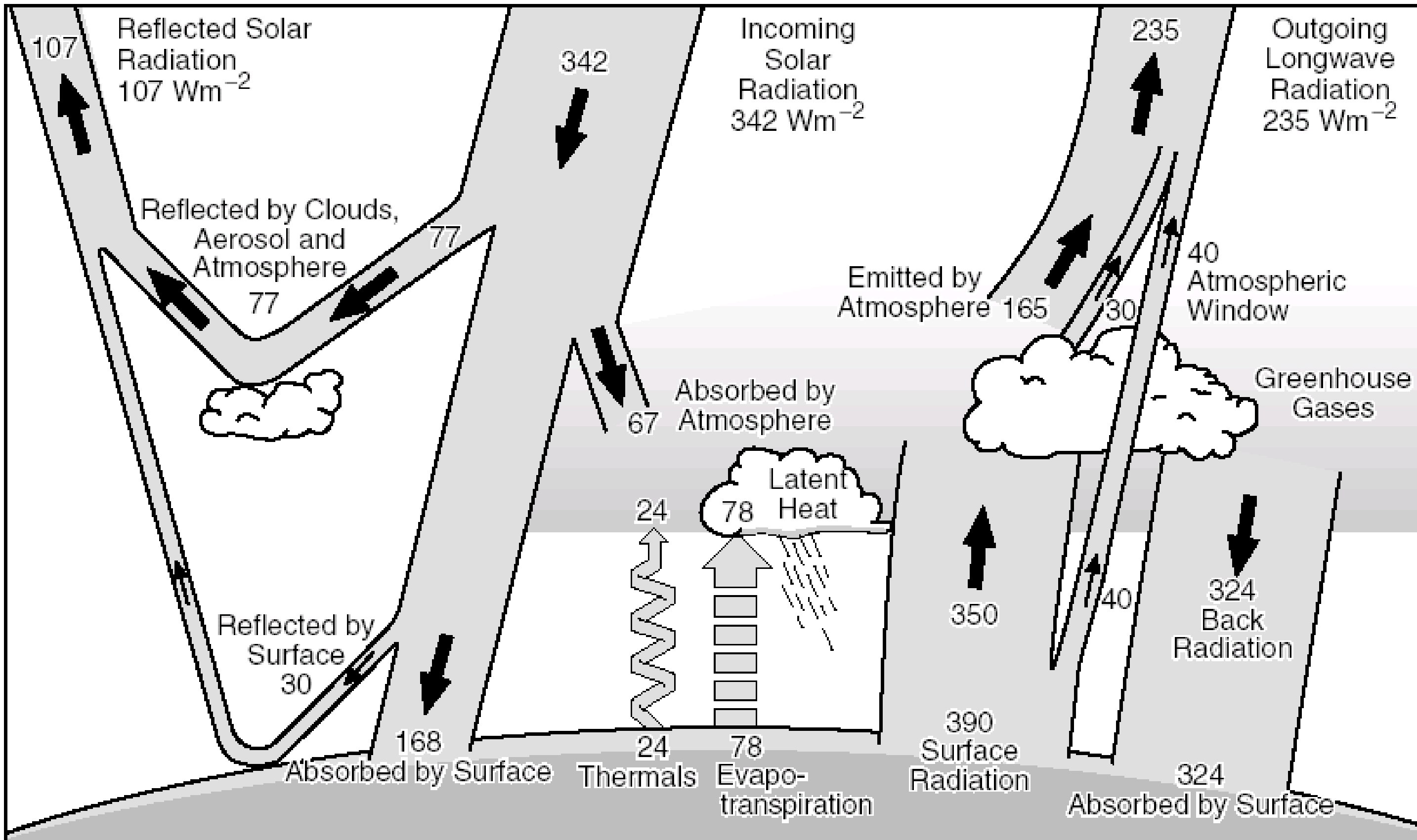
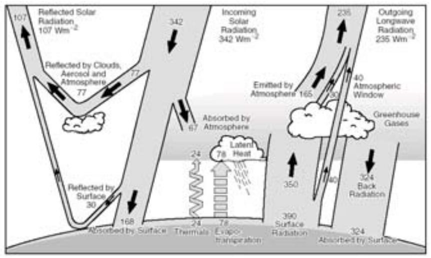
Global energy and water balance

The global energy balance of the climate system



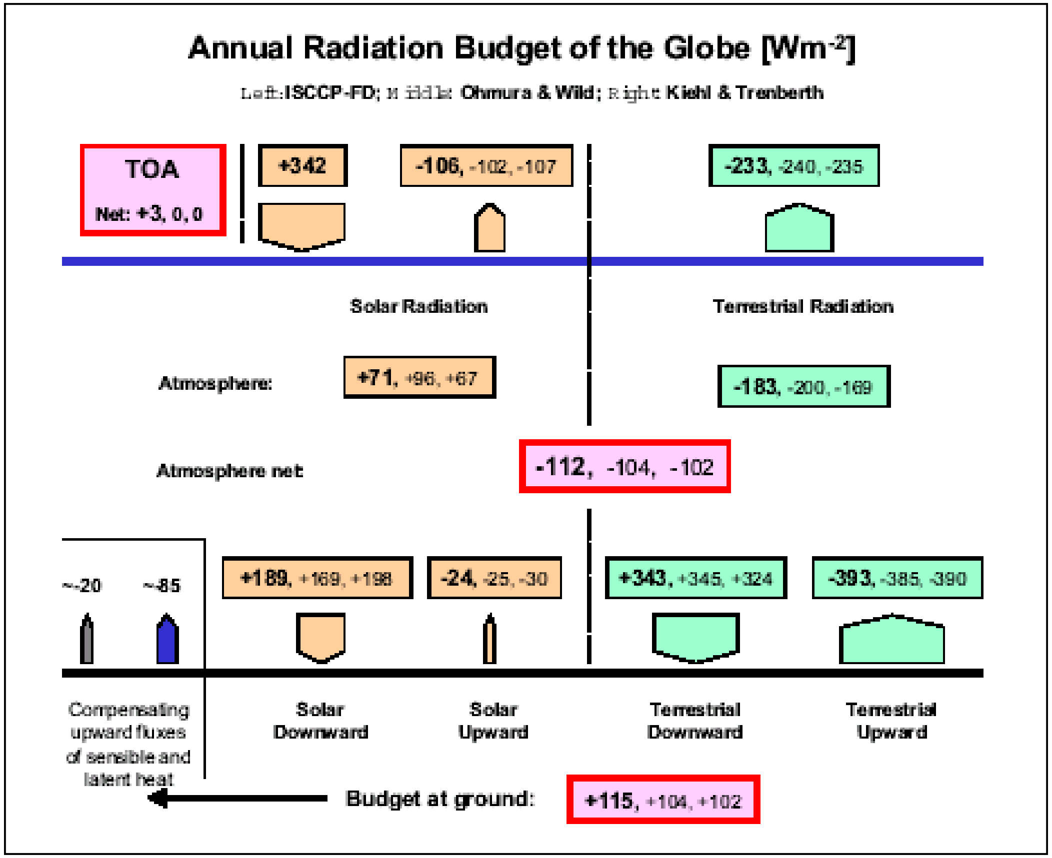
Kiehl and Trenberth (1997)

Important issues



- Short- and longwave fluxes
- Extinction (scattering and absorption), reflection, transmission
- Radiation budget at the top of atmosphere (TOA), $NR_{TOA} \sim 0$ W m^{-2}
- Planetary albedo, $\alpha_P \sim 0.3$
- Surface temperature, $T_S = (LW\uparrow_S / \sigma)^{1/4} = 288$ K ≡ 15 °C
- Radiation budget of the atmosphere, –102 W m^{-2}
- Radiation budget at the Earth's surface, +102 W m^{-2}
- Additional exchange through the turbulent fluxes of sensible, H, and latent heat (evapotranspiration), L_VE, with $H + L_VE = 102$ W m^{-2} (to the atmosphere)
- Bowen ratio, $Bo = H / L_VE \sim 0.2 – 0.3$
- Cloud effects (see course Atmospheric Physics)

Uncertainties

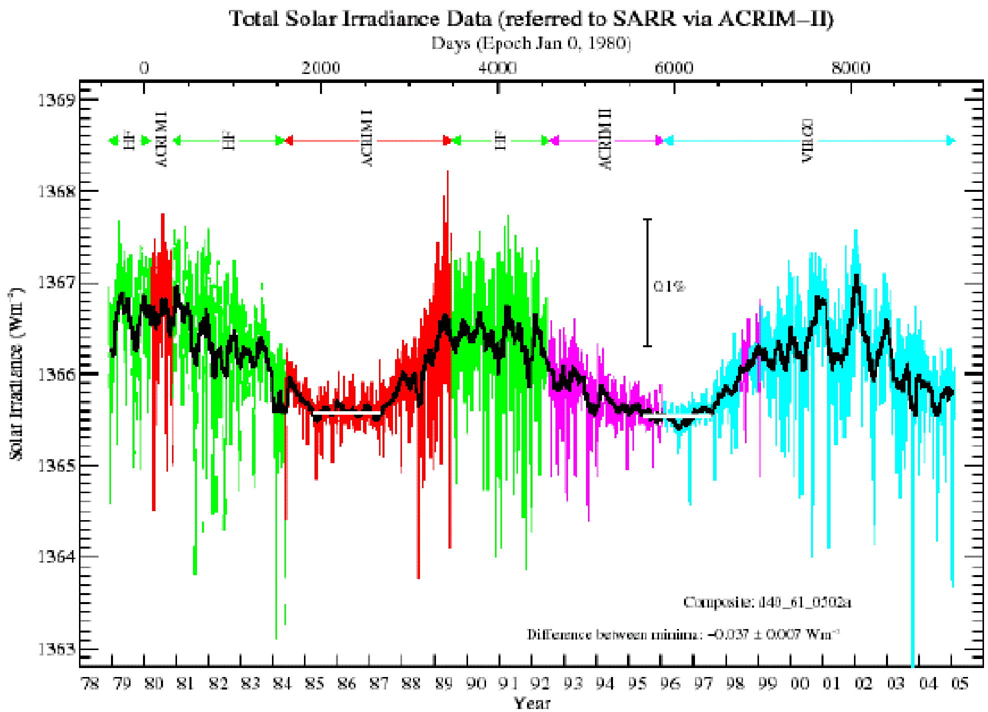


Raschke and Ohmura (2006)

Different estimates of the global energy balance show considerable differences, in particular with respect to the absorption of solar radiation by the atmosphere.

The solar constant

Definition: Value of the solar radiation flux at the top of the atmosphere at the mean distance Earth-Sun



Quinn and Fröhlich (1999)

- average value of $S_o = 1366.5\ W\ m^{-2}$
- convenient estimate of $S_o = 1368\ W\ m^{-2}$, giving $S_o/4 = 342\ W\ m^{-2}$

The Bowen ratio

As seen above, on a global scale Bo = H / $L_V E$ ~ 0.2 – 0.3. How to explain this value? Let us have a look at the relation between Bo and the vertical gradients of the (potential) temperature T and the specific humidity q. It can be shown that:

$$Bo \cong \frac{C_p \; \partial T / \partial z}{L_V \; \partial q / \partial z} \equiv \frac{C_p \, p}{L_V \, \varepsilon} \left(\frac{\partial e}{\partial T} \right)^{-1}$$

where

L_V	=	$2.5 \; 10^6$ J kg^{-1} latent heat of vaporization
C_p	=	1005 J kg^{-1} K^{-1}, specific heat of air at constant temperature
p	=	pressure
ε	=	M_w/M = 0.622, ratio of molecular weights of water and dry air
e	=	water vapor pressure

and where the ratio $\gamma \equiv (C_p p)/(L_V \varepsilon) \approx 0.65$ hPa K^{-1} (at T = 20 C and p = 1000 hPa) is called the psychrometric constant (although it is not constant).

The Bowen ratio (2)

The vapor pressure is related to its saturated value through:

$$e \cong e_{sat} U$$

where U is the relative humidity.

It follows that:

$$Bo \cong \gamma \left(U \frac{de_{sat}}{dT} + e_{sat} \frac{\partial U}{\partial T} \right)^{-1}$$

where by the Clausius-Clapeyron equation

$$\frac{de_{sat}}{dT} = \frac{\varepsilon L_V e_{sat}}{R_d T^2}$$

and

R_d = 287.04 J kg^{-1} K^{-1}, gas constant for dry air.

The Bowen ratio (3)

The above equation shows that the Bowen ratio is mainly determined by the slope of the saturation pressure curve, de_{sat}/dT, and by how the relative humidity U varies with T.

In a saturated atmosphere $U = 1$ and $de/dT = de_{sat}/dT$. Therefore:

$$\mathrm{Bo} \cong \gamma \left(\frac{de_{sat}}{dT} \right)^{-1}$$

At $T = 288$ K ($\equiv 15$ C), $e_{sat} \approx 17$ hPa and $de_{sat}/dT \approx 1$ hPa. Hence:

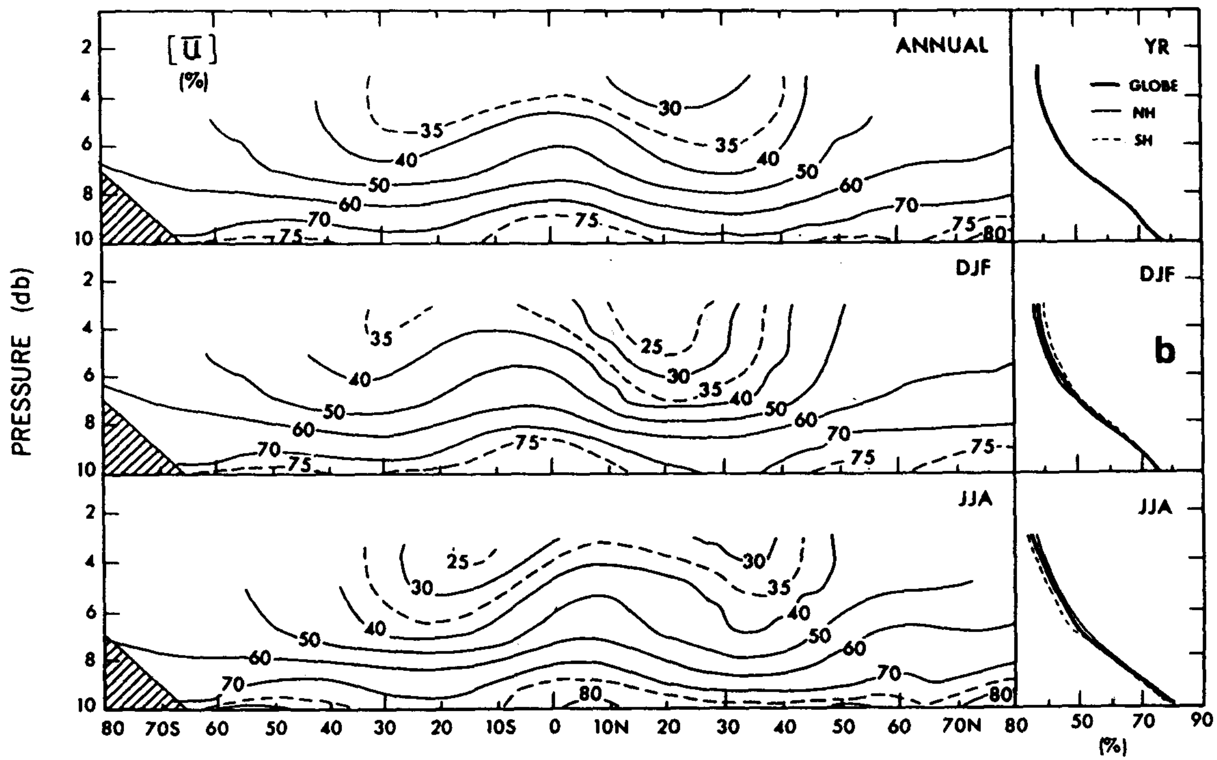
$$\mathrm{Bo} \approx 0.67 \gg 0.2 - 0.3$$

To obtain more realistic estimates, one has to consider the Bowen ratio in a unsaturated atmosphere ($U < 1$) overlying a saturated surface (ocean). Under such circumstances $U \sim 0.8$ and close to the surface $\partial U/\partial T \sim 0.2\ \mathrm{K}^{-1}$, giving:

$$\mathrm{Bo} \cong \gamma \left(U \frac{de_{sat}}{dT} + e_{sat} \frac{\partial U}{\partial T} \right)^{-1} \sim 0.2$$

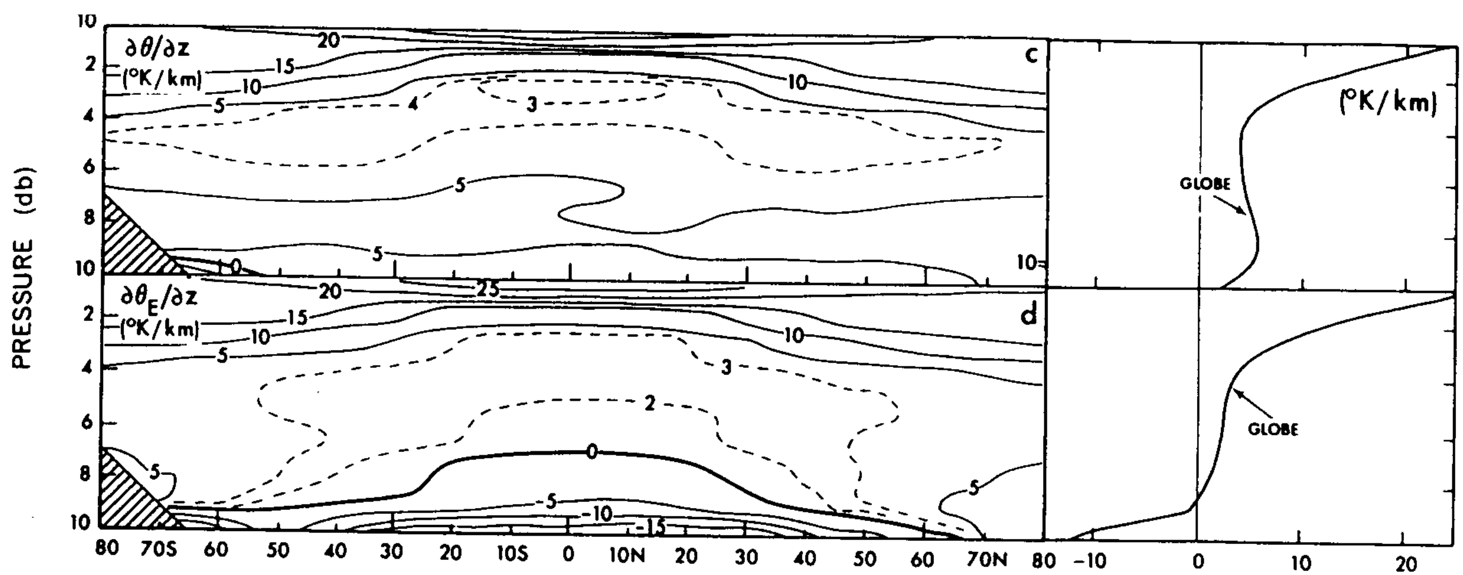
The Bowen ratio (4)

It is clear, however, that the exact value of the Bowen ratio very much depends on the precise relation between relative humidity and temperature. According to Peixoto and Oort (1992) the global distribution of the relative humidity is as follows:



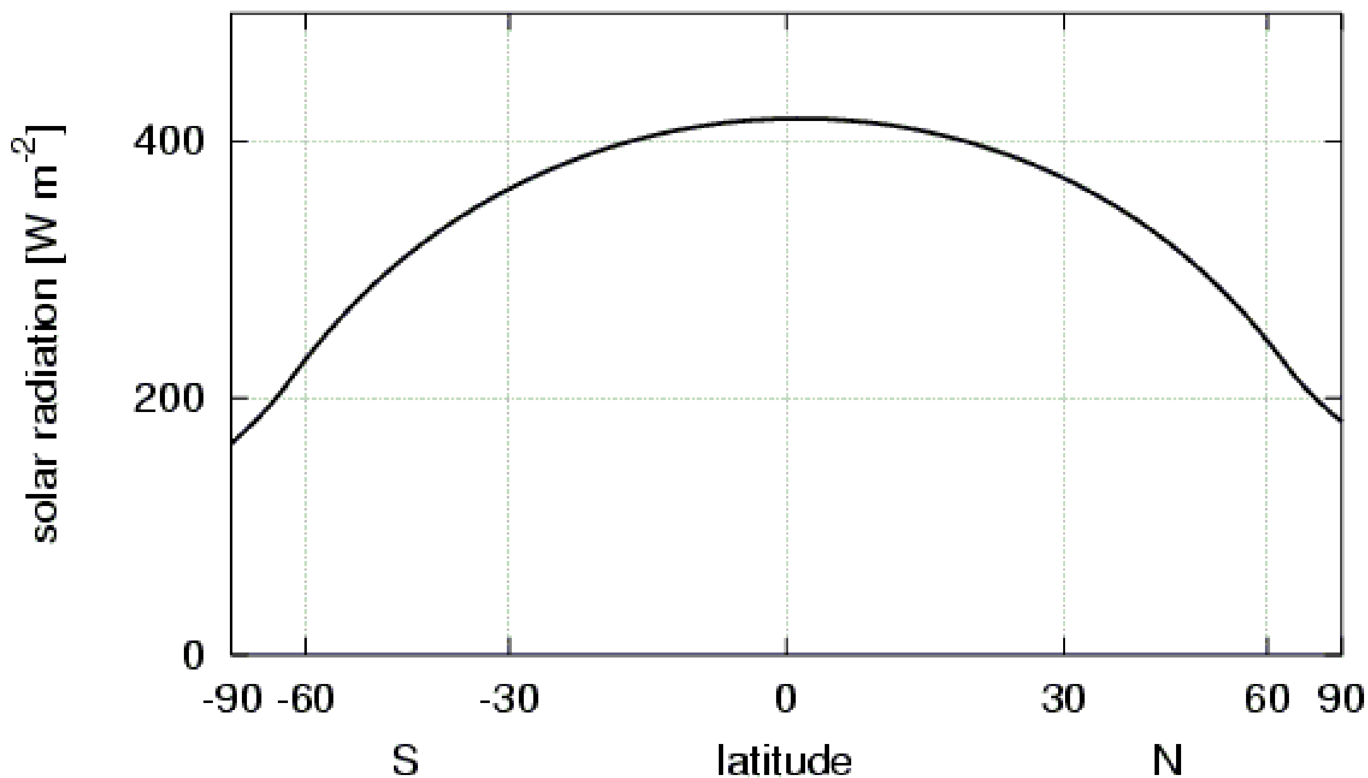
Static stability

Water vapor is essential not only as a carrier of latent heat, but also because of its effects on the static stability of the atmospheric boundary layer (see course on Atmospheric Physics). The following picture, after Peixoto and Oort (1992), emphasizes the differences between the vertical gradients of potential and equivalent potential temperature.



Meridional distribution of the radiative fluxes

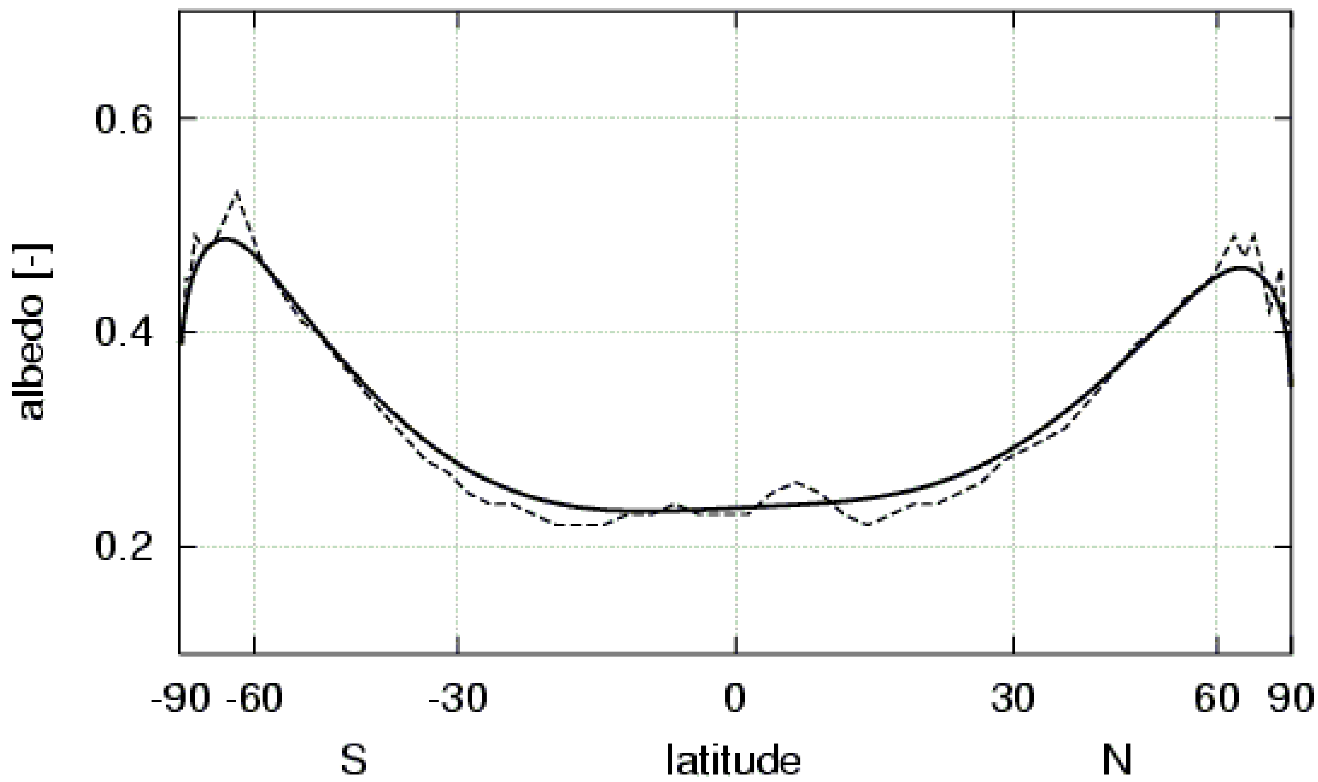
The zonal mean values shown in the following figures were obtained from the Earth Radiation Budget Experiment (ERBE, Barkstrom et al., 1990). The plots refer to the conditions at the top of the atmosphere (TOA), the atmospheric outer boundary.



Barkstrom B. et al., 1990: Earth Radiation Budget Experiment. Preliminary Seasonal Results. *EOS*, Feb. 27

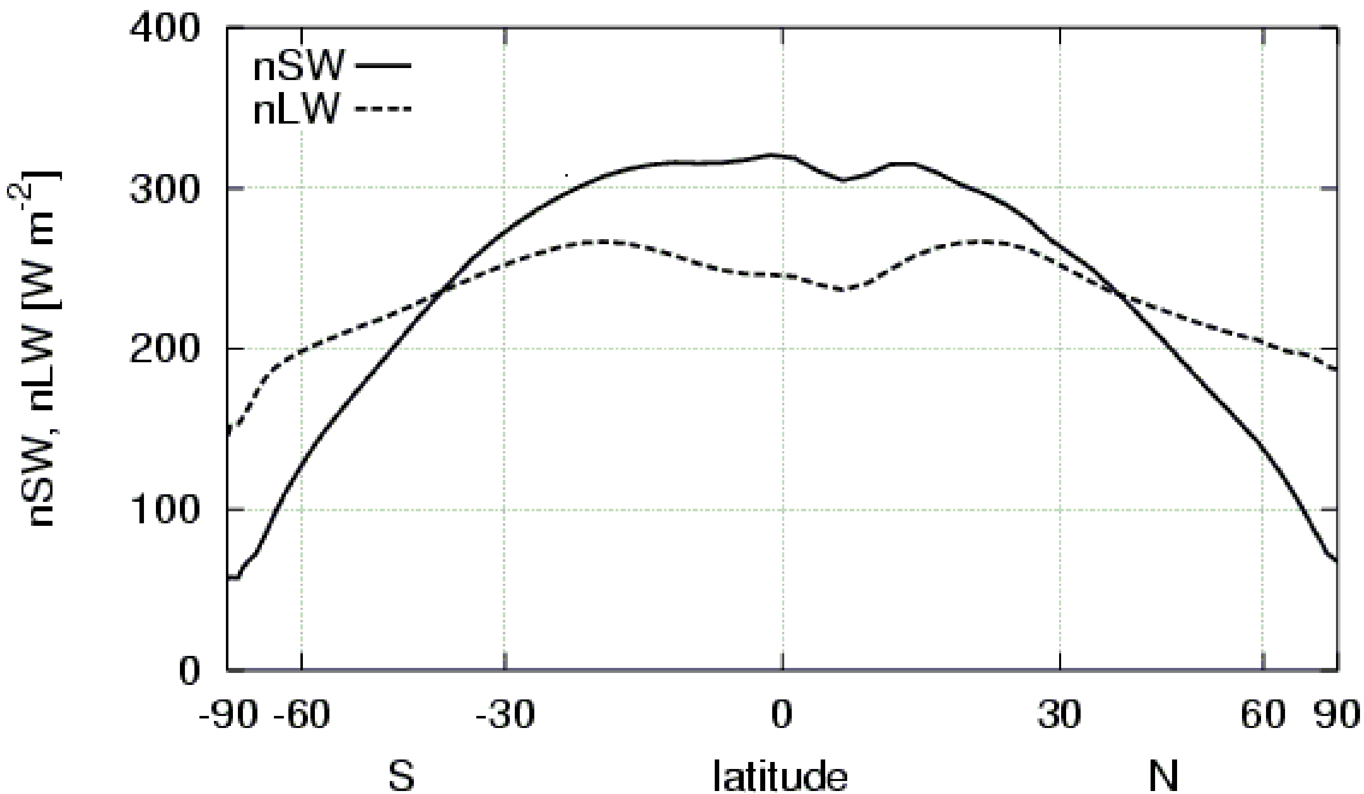
Meridional distribution of the radiative fluxes (2)

Same as before but for the albedo.



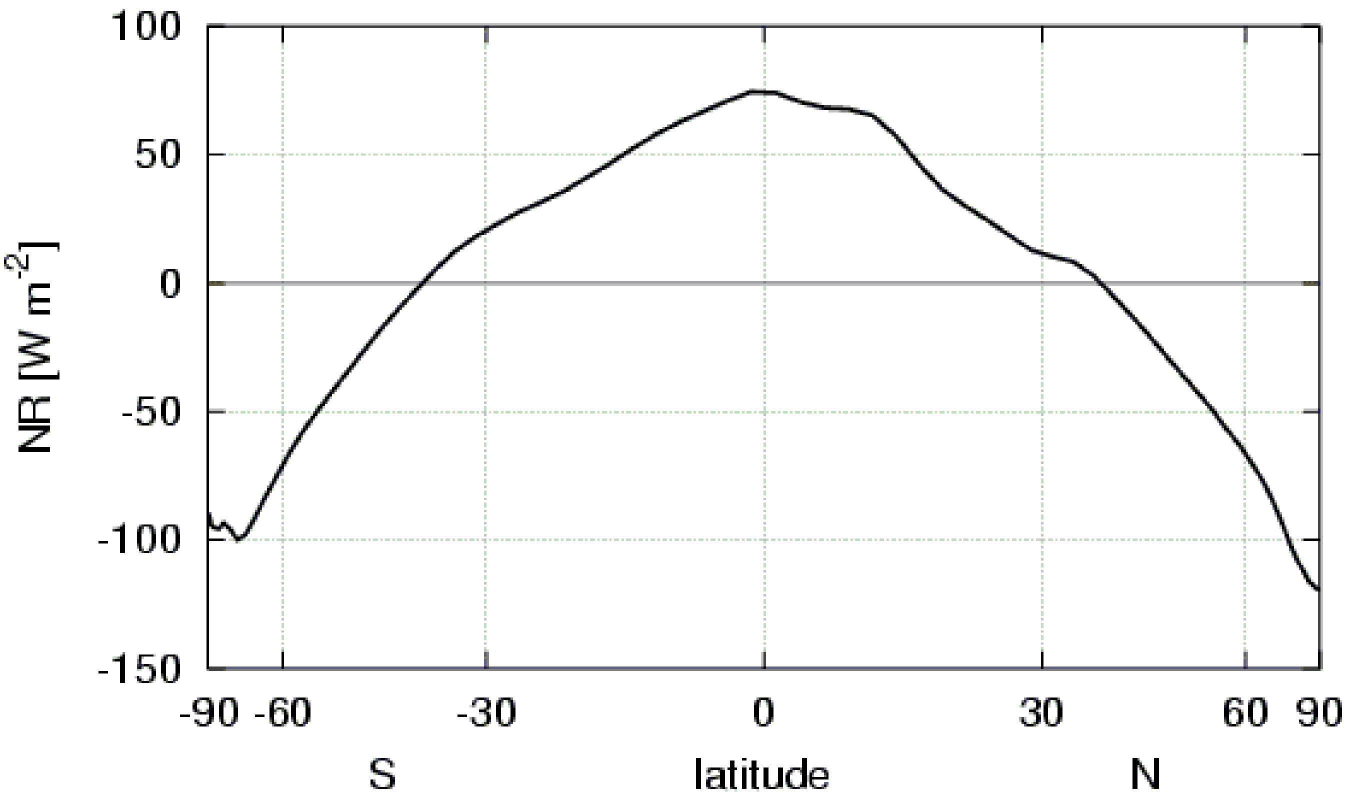
Meridional distribution of the radiative fluxes (3)

Same as before but for the net shortwave and net longwave radiation. The net shortwave radiation is the difference between incoming solar radiation and reflected radiation (given the albedo of the previous picture). The net longwave radiation is, in this case, simply the longwave emission.



Meridional distribution of the radiative fluxes (4)

Same as before but for the net radiation (radiation budget), the difference between the net shortwave and net longwave fluxes of the previous figure.



Meridional distribution of the radiative fluxes (5)

In the above plots the abscissa is actually the sinus of the latitude, sin(φ), despite the fact that the axis is labelled with φ. Recall in fact that the area S_{12} between two latitudes, say φ_1 and φ_2, is given by:

$$S_{12} = 2\pi r_E^2 \int_{\varphi_1}^{\varphi_2} \cos\varphi \, d\varphi = 2\pi r_E^2 \left[\sin(\varphi_2) - \sin(\varphi_1)\right]$$

Meridional energy transport

On an annual mean basis all latitudes appear to be in thermal equilibrium:

$$\frac{\partial E}{\partial t} = 0$$

This implies a meridional transport of energy to compensate for the positive/negative radiation balance of the low/high latitudes.

The energy transport at a given latitude φ can be computed by integrating the energy budget of a latitudinal zone, which reads:

$$NR = nSW - nLW = div_{\varphi}(EF)$$

where

NR = zonal mean net radiation

nSW = zonal mean net shortwave radiation

nLW = zonal mean net longwave radiation

EF = energy flux or energy transport

div_{φ} = divergence operator (only variations in φ are relevant here)

Meridional energy transport (2)

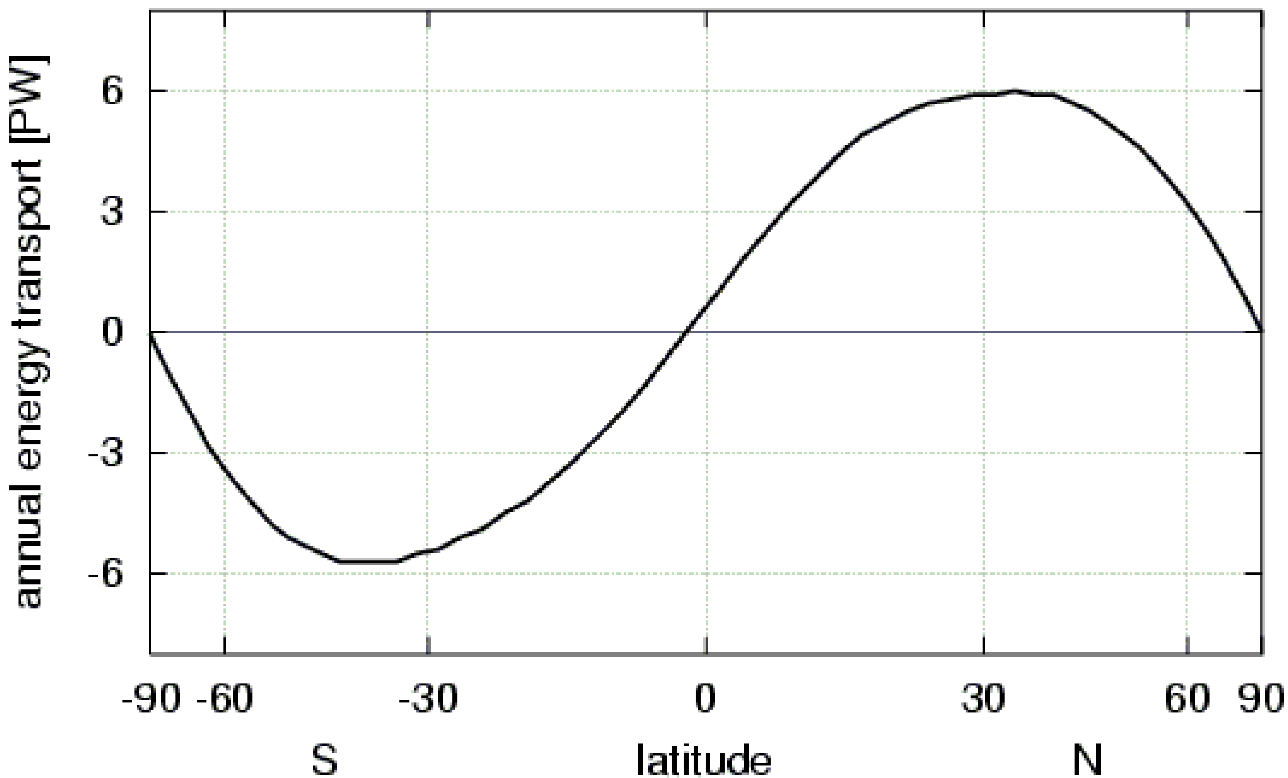
It follows that:

$$EF(\varphi) = 2\pi r_E^2 \int_{-\pi/2}^{\varphi} NR \cos\varphi \, d\varphi$$

Since the zonal mean radiation balance NR is in units of [W m^{-2}], the meridional energy transport is in units of [W]. For convenience it is usually expressed in [PW]. 1 PW ≡ 10^{15} W.

The result of the integration of the ERBE data is shown in the following figure.

Meridional energy transport (3)



Meridional energy transport (4)

The transport is partly accomplished by the atmosphere, partly by the ocean. The classical picture, with about half of the meridional transport taking place with the atmospheric motion, the other half with the oceanic currents, is the one presented in Peixoto and Oort (1992).

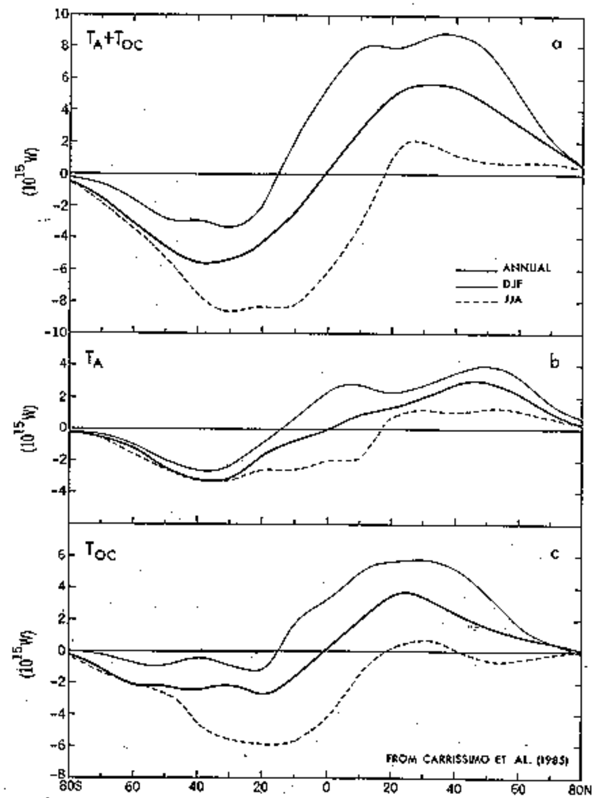


FIGURE 13.19. Meridional profiles of the northward transport of energy in the total system (a), the atmosphere (b), and the oceans (c) for annual, DJF, and JJA mean conditions in units of 10^{15} W (based on data from Carissimo *et al.*, 1985).

Meridional distribution of the energy transport (5)

New data and analysis techniques indicate that most of the transport is realized by the atmospheric circulation (see e.g. Trenberth and Caron, 2001).

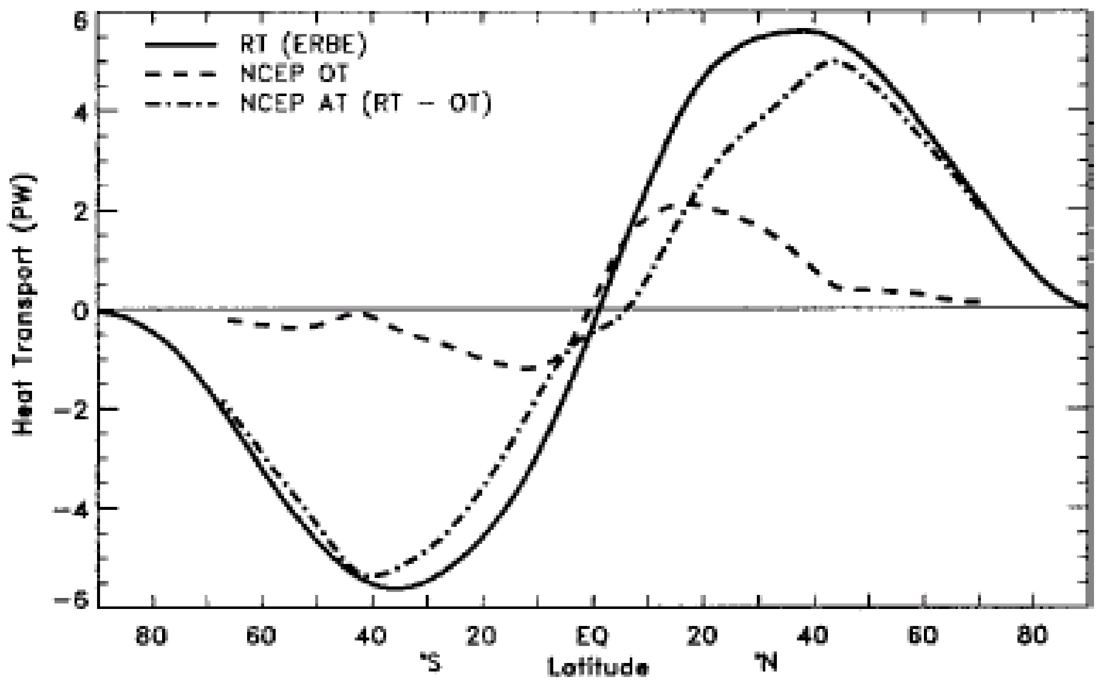
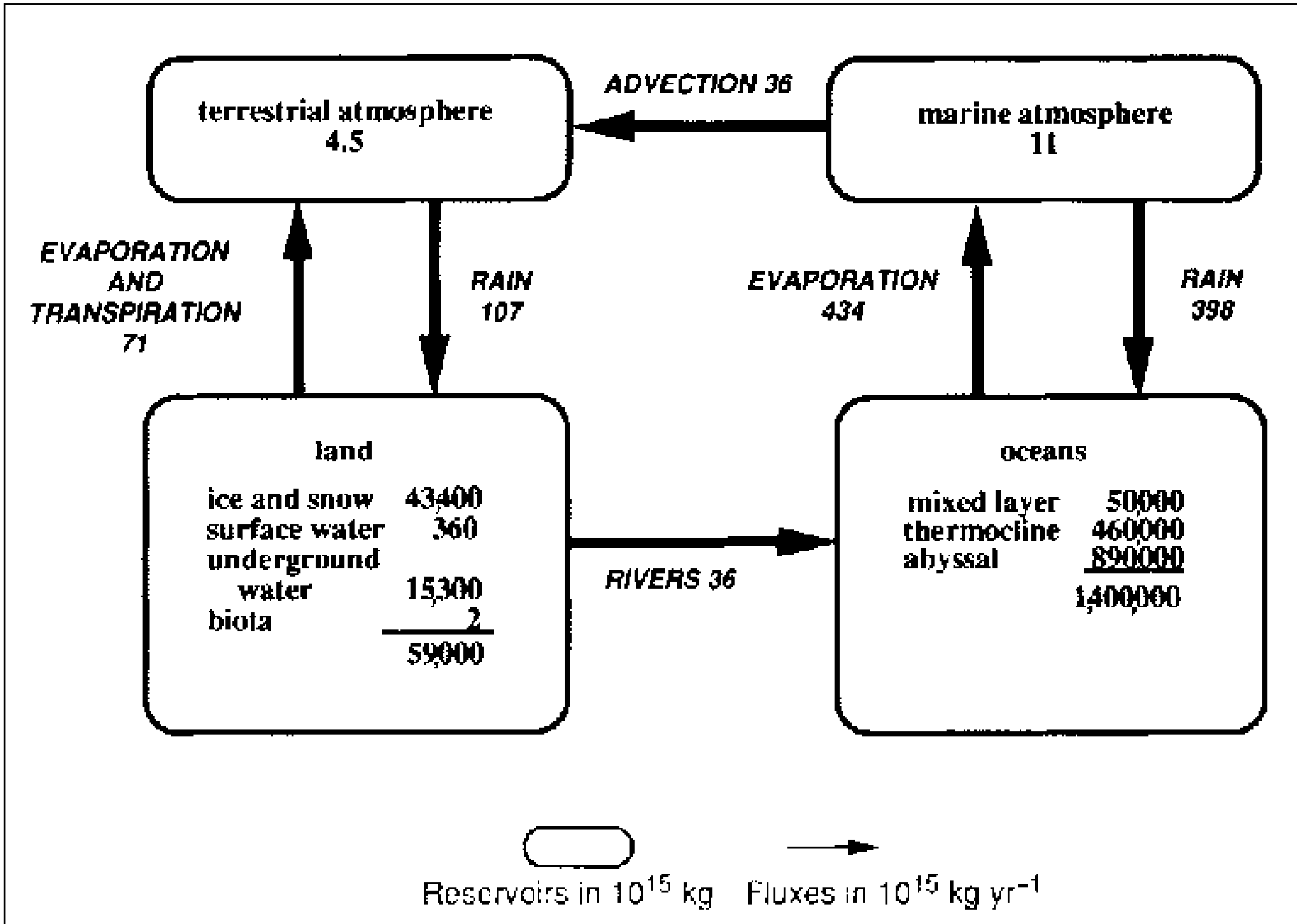


FIG. 7. The required total heat transport from the TOA radiation RT is compared with the derived estimate of the adjusted ocean heat transport OT (dashed) and implied atmospheric transport AT from NCEP reanalyses (PW).

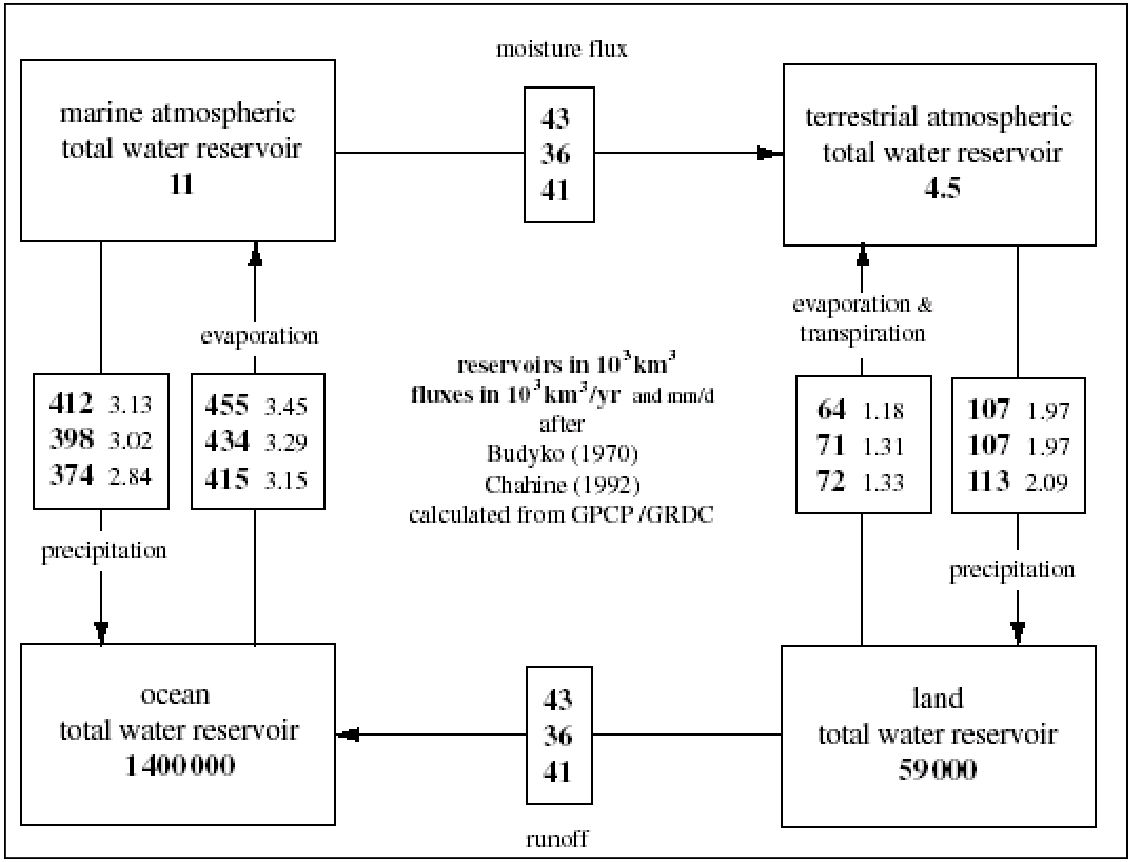
Trenberth, K.E. and J.M. Caron, 2001: Estimates of meridional atmosphere and ocean heat transports. *J. Climate*, **14**, 3433-3443.

Global water balance (1)



Chahine (1992)

Global water balance (2)



Rudolf and Rubel (2006)

As for the global energy budget, uncertainties in the individual terms can be considerable. A water density of 1000 kg m^{-3} and an earth's surface of $4 \pi r_E^2 = 5.1 \cdot 10^{14}$ $m^2 = 5.1 \cdot 10^8$ km^2 are necessary to convert from 10^{15} kg to 10^3 km^3 to mm.

Global precipitation and the surface radiation budget (1)

Why would you expect a strong relationship between annual precipitation and the radiation budget at the Earth's surface?

On an annual basis global precipitation ≡ global evaporation. Moreover over the oceans the surface energy balance can be very crudely approximated as $NR = L_V E + H + G \sim L_V E$. Over land, too, the maximum evapotranspiration is related to the net radiation. Consider for instance the expression for the so-called equilibrium evapotranspiration (Brutsaert, 1982):

$$E_{equil} = \frac{1}{L_V} \frac{\Delta}{\Delta + \gamma} NR$$

where

L_V = $2.5\ 10^6$ J kg^{-1}, latent heat of vaporization and

Δ = de_{sat}/dT, gradient of the saturation water vapour pressure

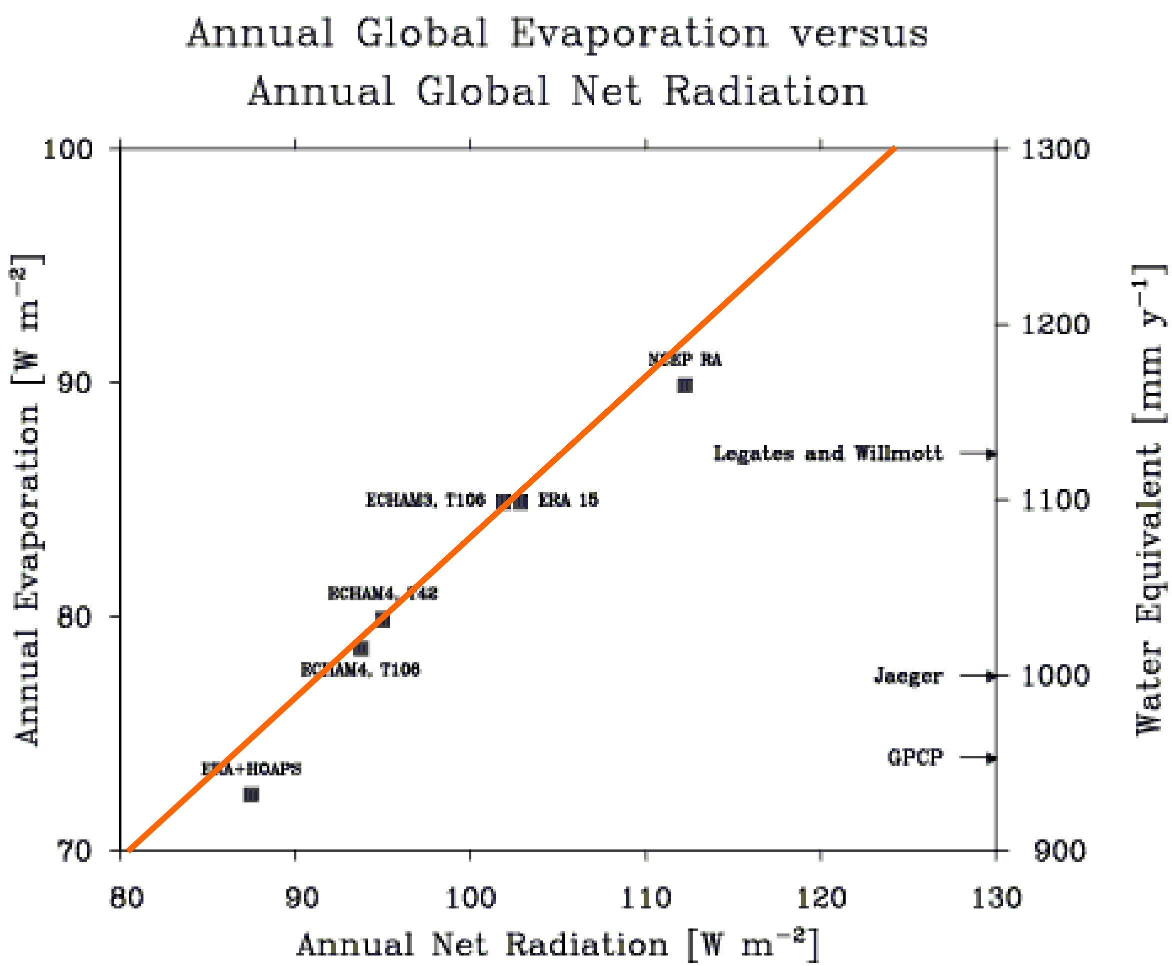
γ = $(C_p\, p)/(\varepsilon\, L_V)$, psychrometric constant

ε = $M_w/M = 0.622$, ratio of molecular weights of water and dry air.

In sum:

$$P \sim ET \sim NR$$

Global precipitation and the surface radiation budget (2)



Ohmura (pers. comm.)