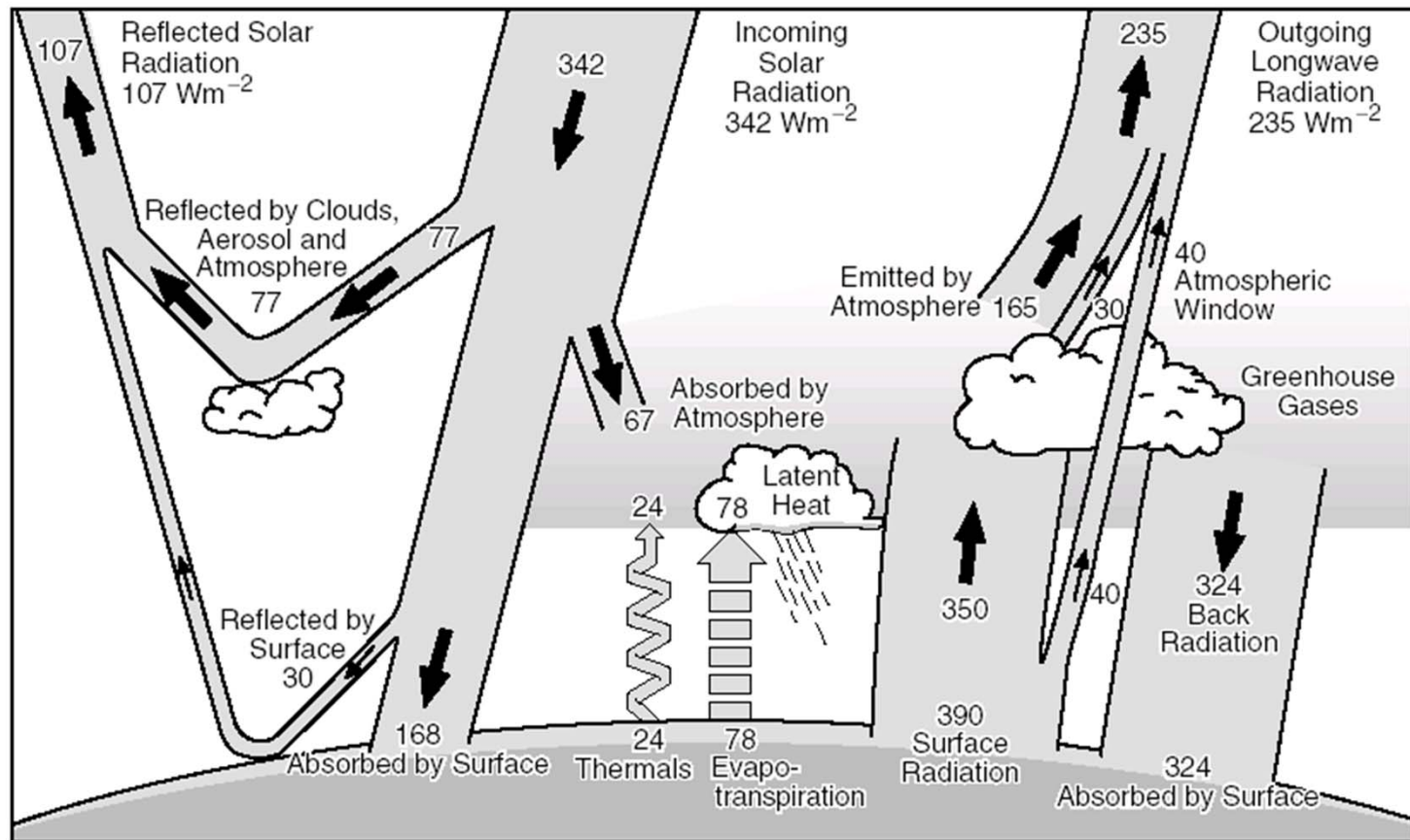


Global energy and water balance

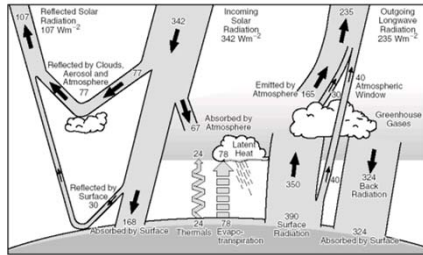
The global energy balance of the climate system



Kiehl and Trenberth (1997)

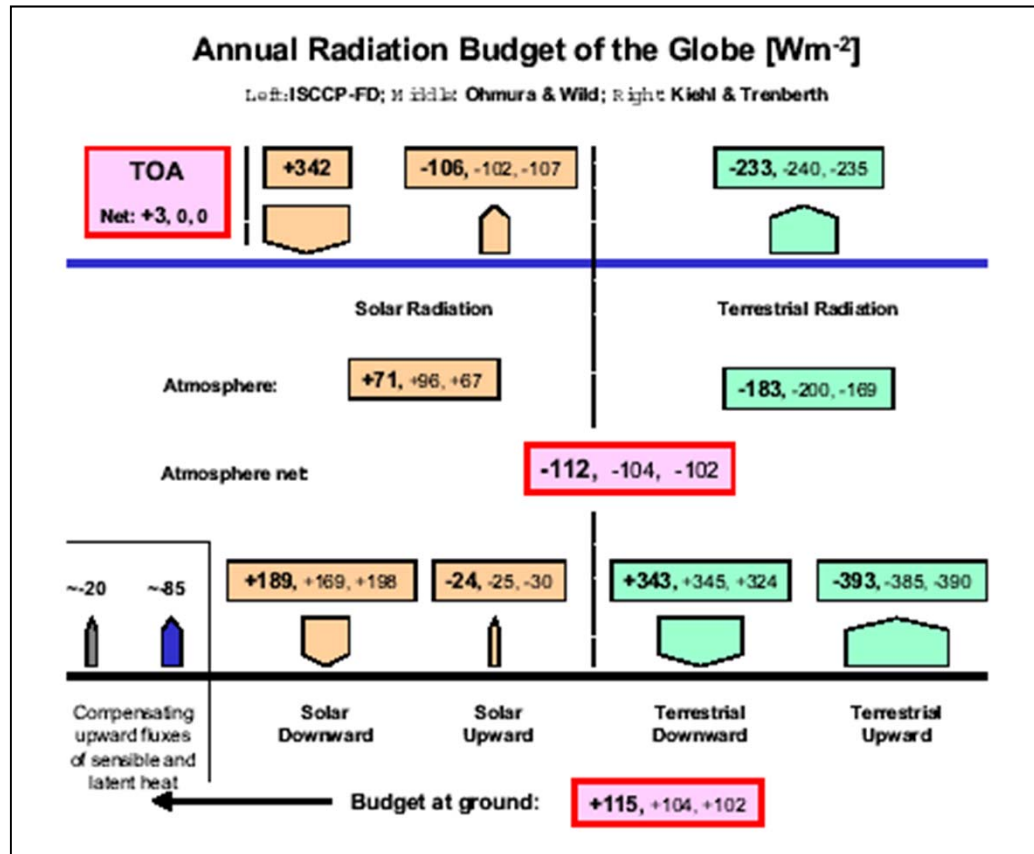


Important issues



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

Uncertainties



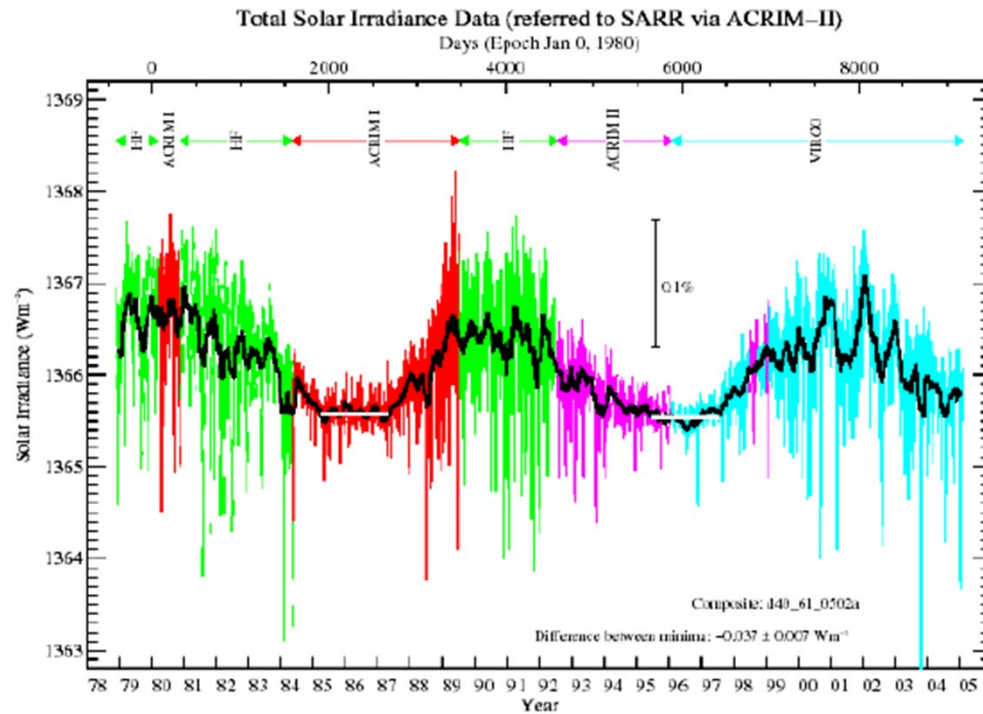
Raschke and Ohmura (2006)

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 $S_0 = 1366.5 \text{ W m}^{-2}$
- convenient estimate of $S_0 = 1368 \text{ W m}^{-2}$, giving $S_0/4 = 342 \text{ W m}^{-2}$



The Bowen ratio

As seen above, on a global scale $Bo = H / L_v E \sim 0.2 - 0.3$. 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

$L_v = 2.5 \cdot 10^6 \text{ J kg}^{-1}$ latent heat of vaporization

$C_p = 1005 \text{ J kg}^{-1} \text{ K}^{-1}$, specific heat of air at constant temperature

$p =$ pressure

$\varepsilon = M_w / M = 0.622$, ratio of molecular weights of water and dry air

$e =$ water vapor pressure

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



The Bowen ratio (2)

The vapor pressure is related to its saturated value through:

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

where U is the relative humidity.

It follows that:

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

where by the Clausius-Clapeyron equation

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

and

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

Clausius-Clapeyron basically says that the water-holding capacity of air increases by about 7% per degree C increase in temperature.



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 $de/dT = de_{\text{sat}}/dT$. Therefore:

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

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

$$Bo \approx 0.67 \gg 0.2 - 0.3$$

To obtain more realistic estimates, one has to consider the Bowen ratio in a unsaturated atmosphere ($U < 1$) overlying a saturated surface (ocean).

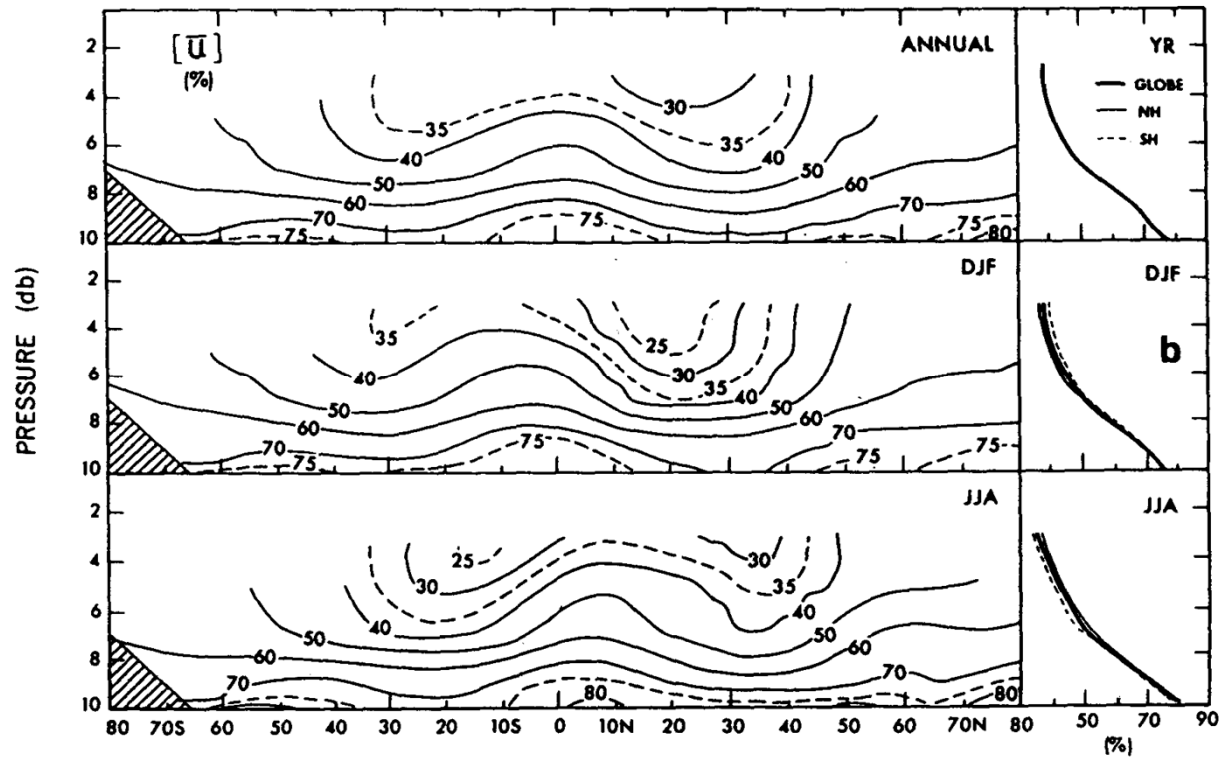
Under such circumstances $U \sim 0.8$ and close to the surface $\partial U/\partial T \sim 0.2 \text{ K}^{-1}$, giving:

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



The Bowen ratio (4)

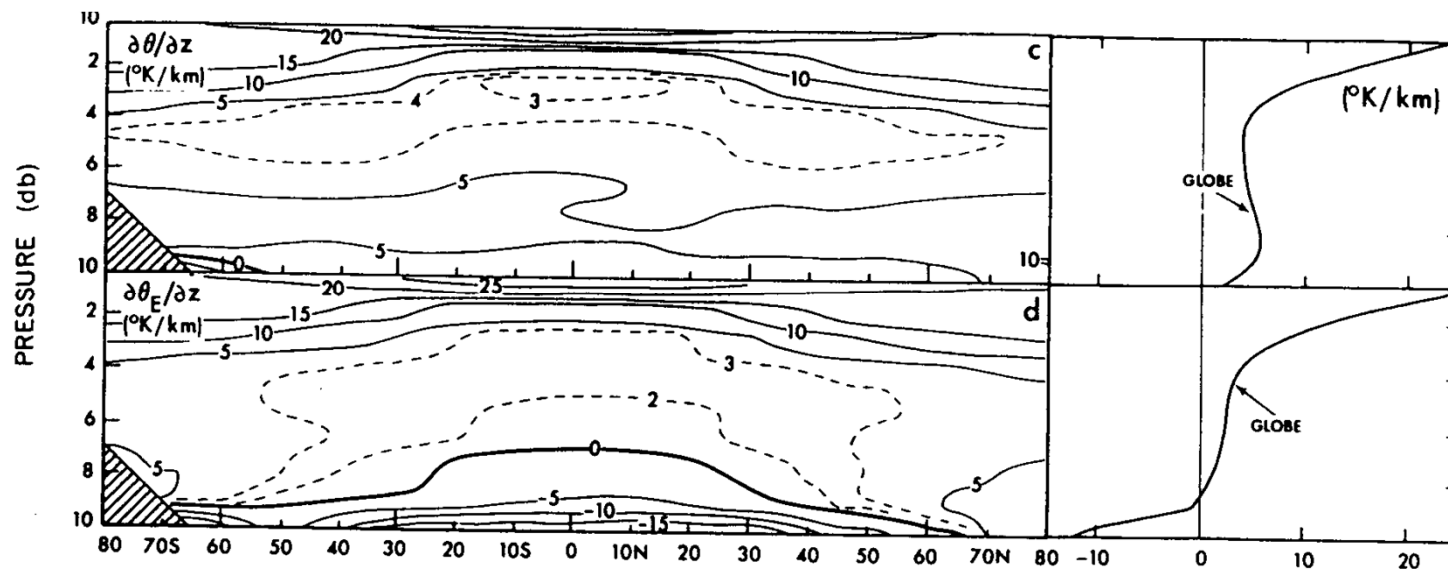
It is clear, however, that the exact value of the Bowen ratio very much depends on the precise relation between relative humidity and temperature. According to Peixoto and Oort (1992) the global distribution of the relative humidity is as follows:





Static stability

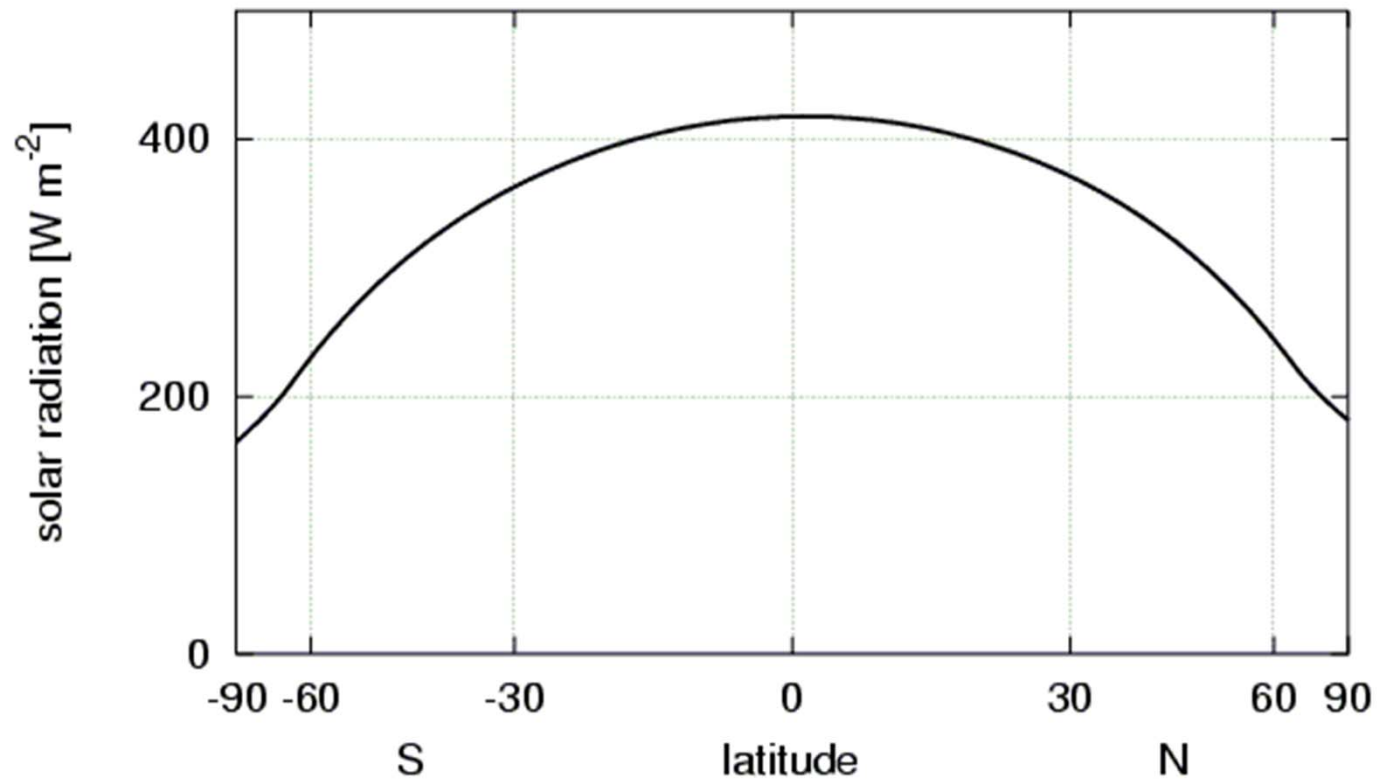
Water vapor is essential not only as a carrier of latent heat, but also because of its effects on the static stability of the atmospheric boundary layer (see course on Atmospheric Physics). The following picture, after Peixoto and Oort (1992), emphasizes the differences between the vertical gradients of potential and equivalent potential temperature (the temperature considering the energy from evaporation water vapor in the parcel). Water vapor can make the vertical gradient of equivalent potential temperature negative and therefore the lower atmosphere unstable.

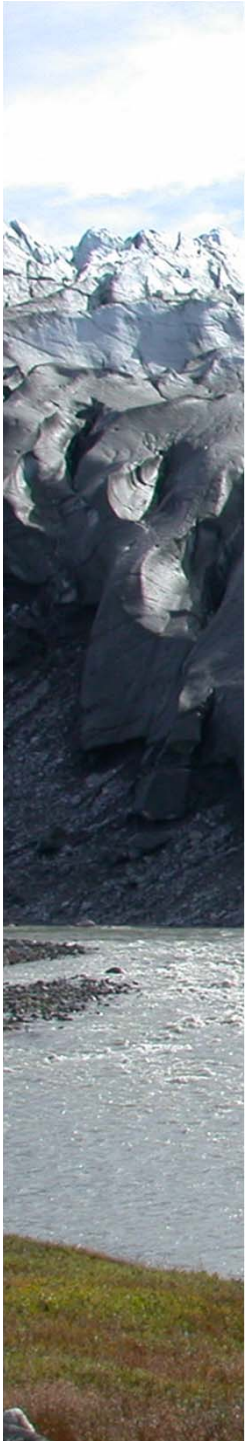




Meridional distribution of the radiative fluxes

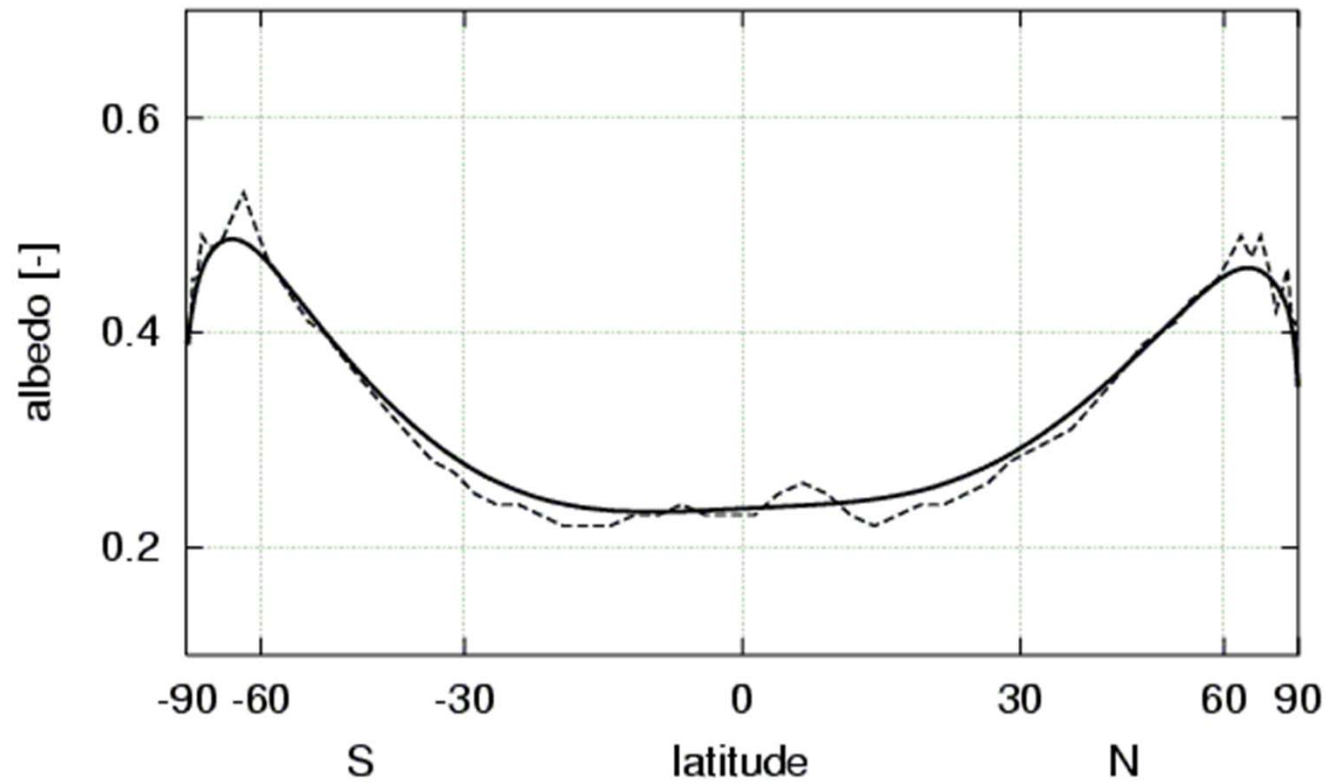
The zonal mean values shown in the following figures were obtained from the Earth Radiation Budget Experiment (ERBE, Barkstrom et al., 1990). The plots refer to the conditions at the top of the atmosphere (TOA), the atmospheric outer boundary.

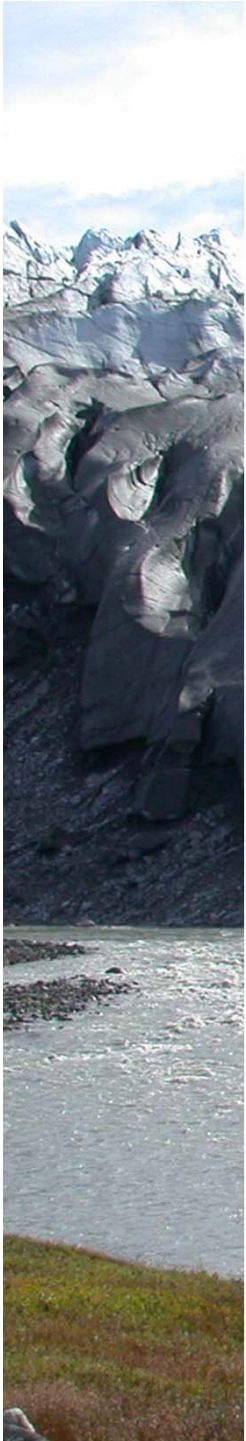




Meridional distribution of the radiative fluxes (2)

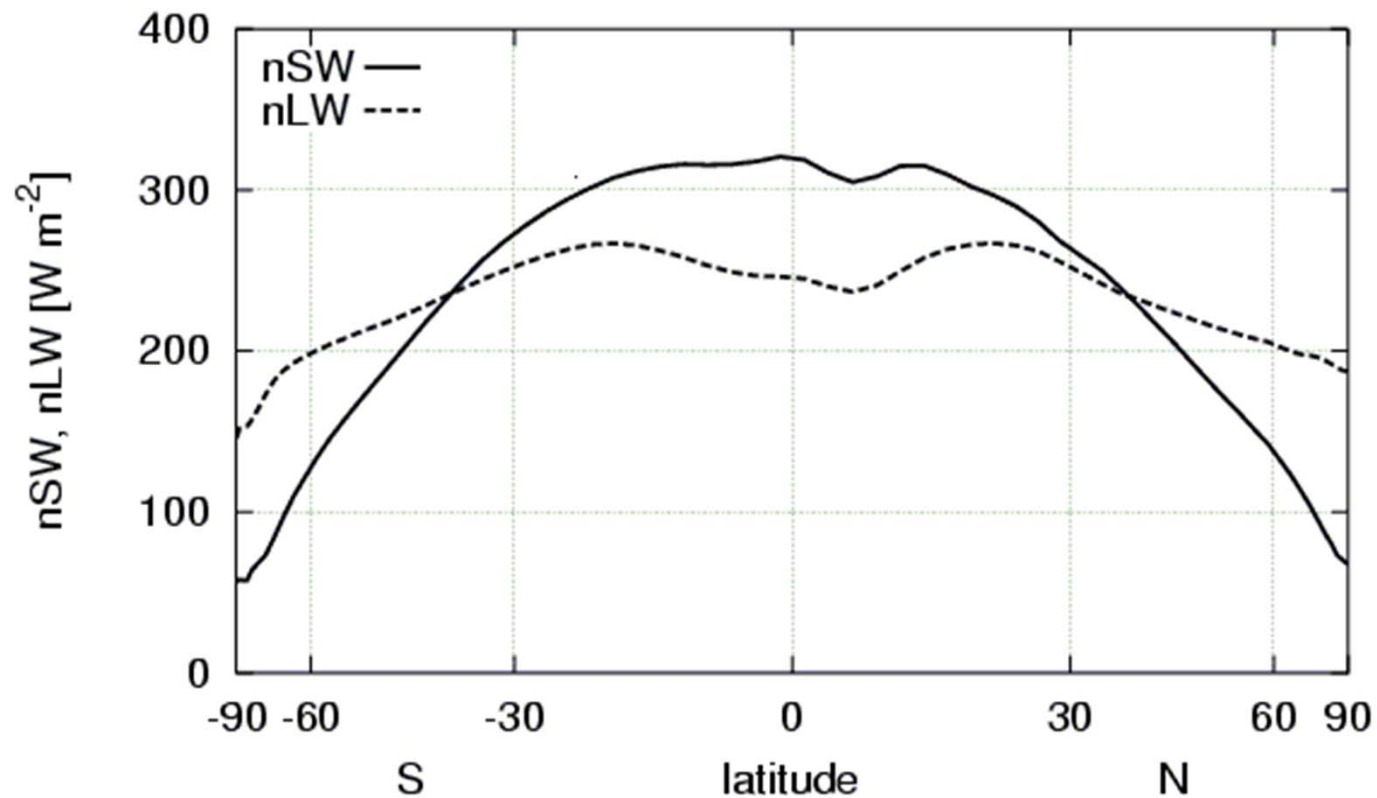
Same as before but for the planetary albedo.





Meridional distribution of the radiative fluxes (3)

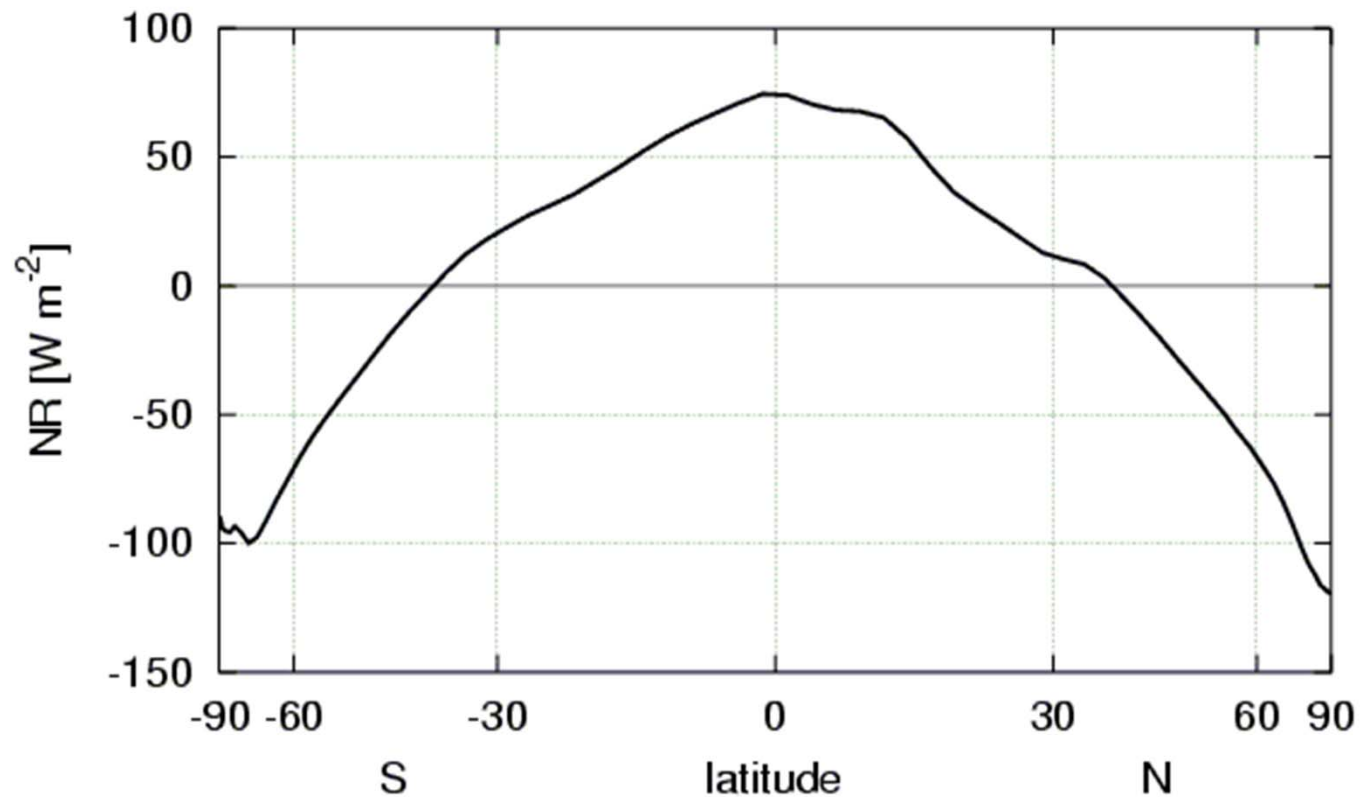
Same as before but for the net shortwave and net longwave radiation. The net shortwave radiation is the difference between incoming solar radiation and reflected radiation (given the albedo of the previous picture). The net longwave radiation is, in this case, simply the longwave emission.

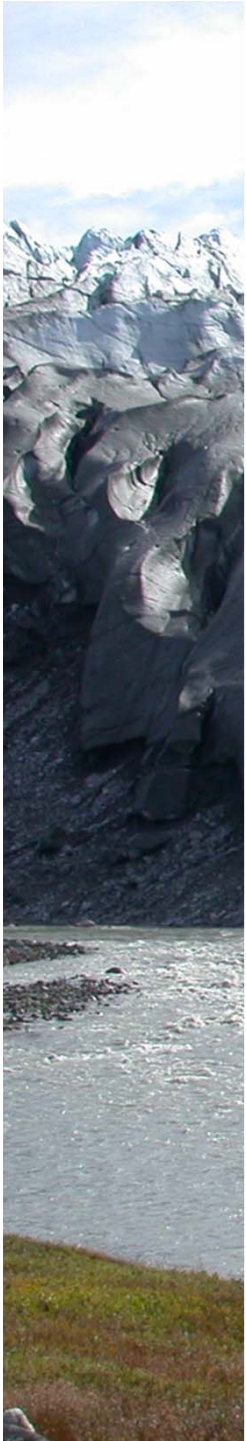




Meridional distribution of the radiative fluxes (4)

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, $\sin(\varphi)$, 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:

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



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:

$$NR = nSW - nLW = \text{div}_{\varphi}(EF)$$

where

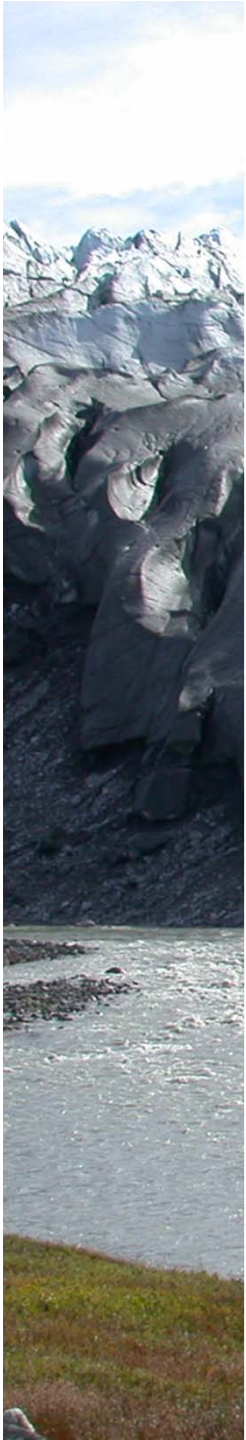
NR = zonal mean net radiation

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)

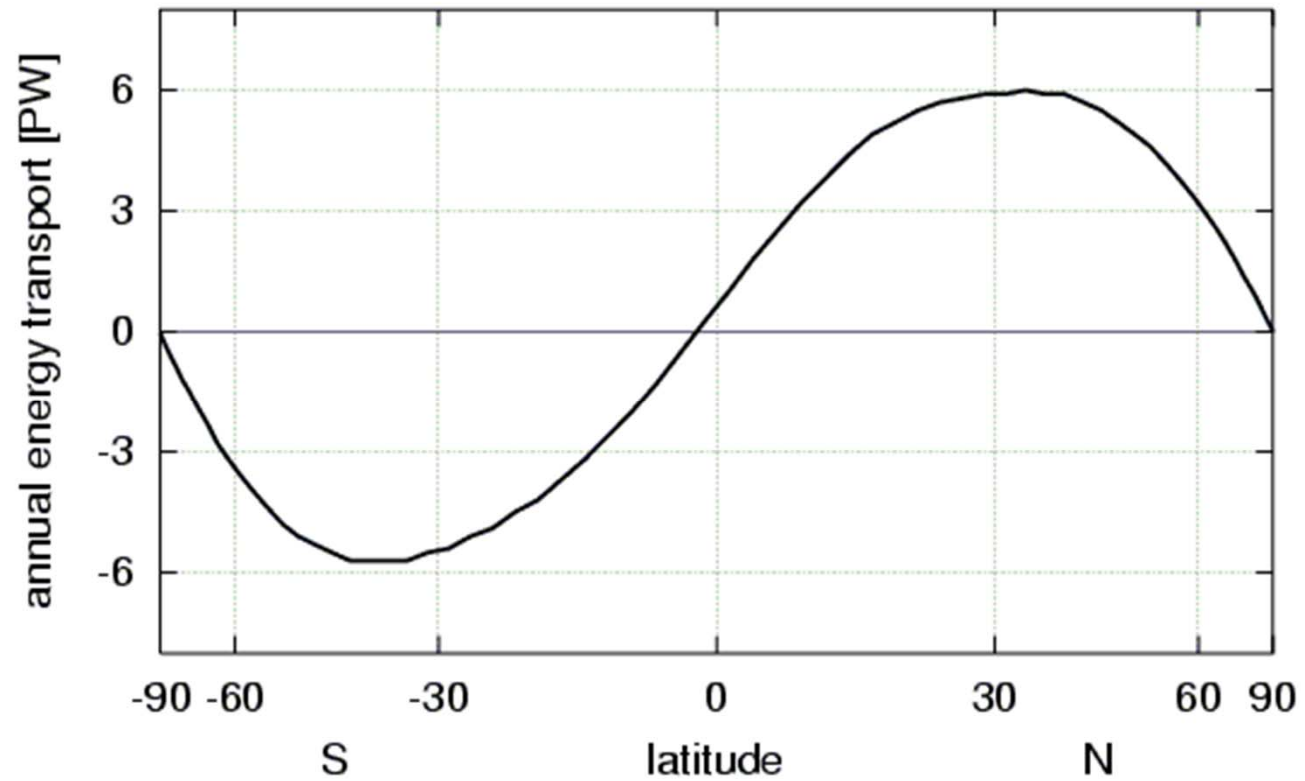
It follows that:

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

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 $[PW]$. $1 PW \equiv 10^{15} W$.

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).

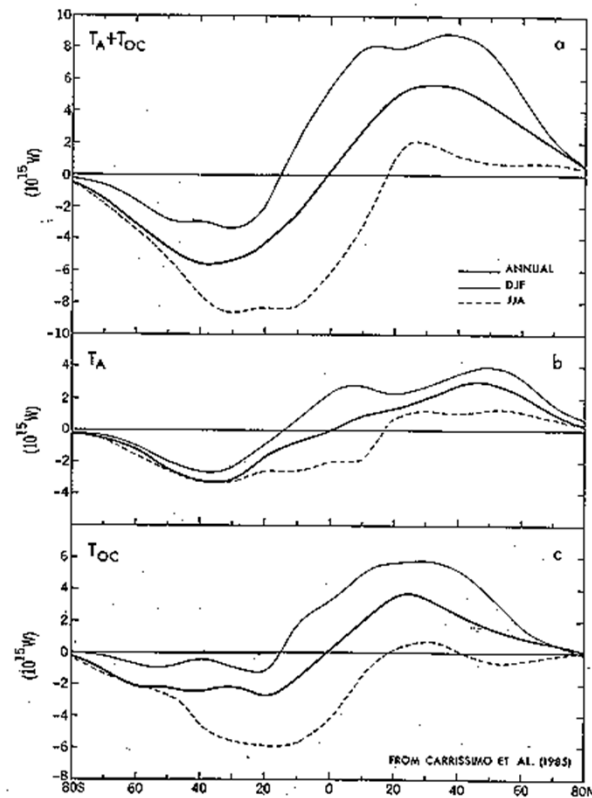


FIGURE 13.19 Meridional profiles of the northward transport of energy in the total system (a), the atmosphere (b), and the oceans (c) for annual, DJF, and JJA mean conditions in units of 10^{15} W (based on data from Carrisano *et al.*, 1985).



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).

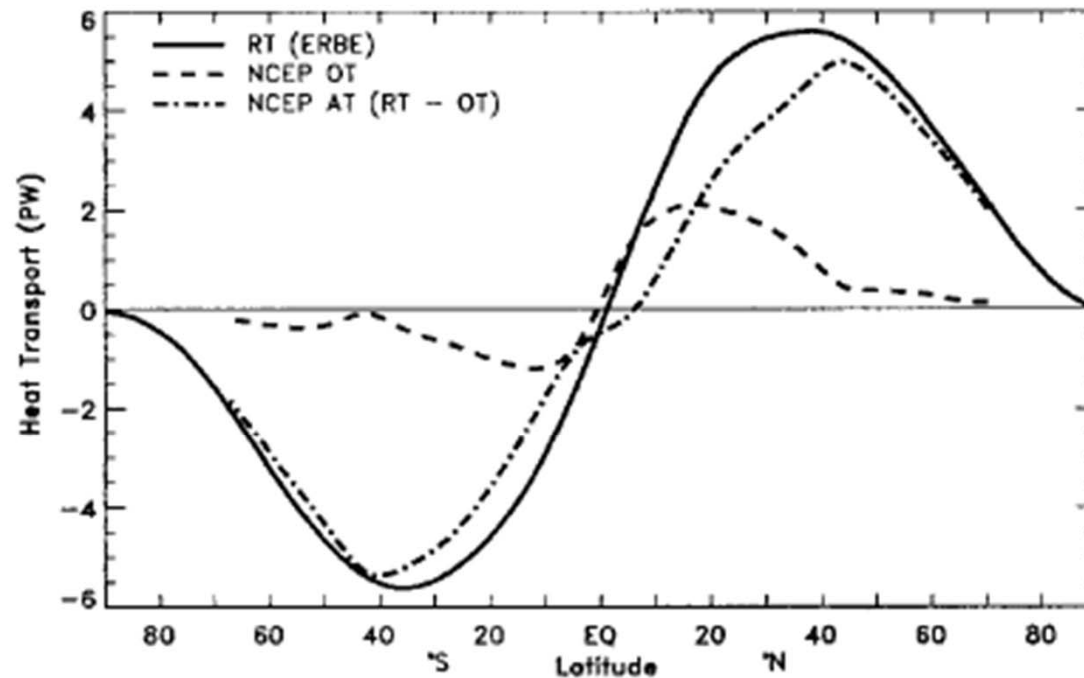


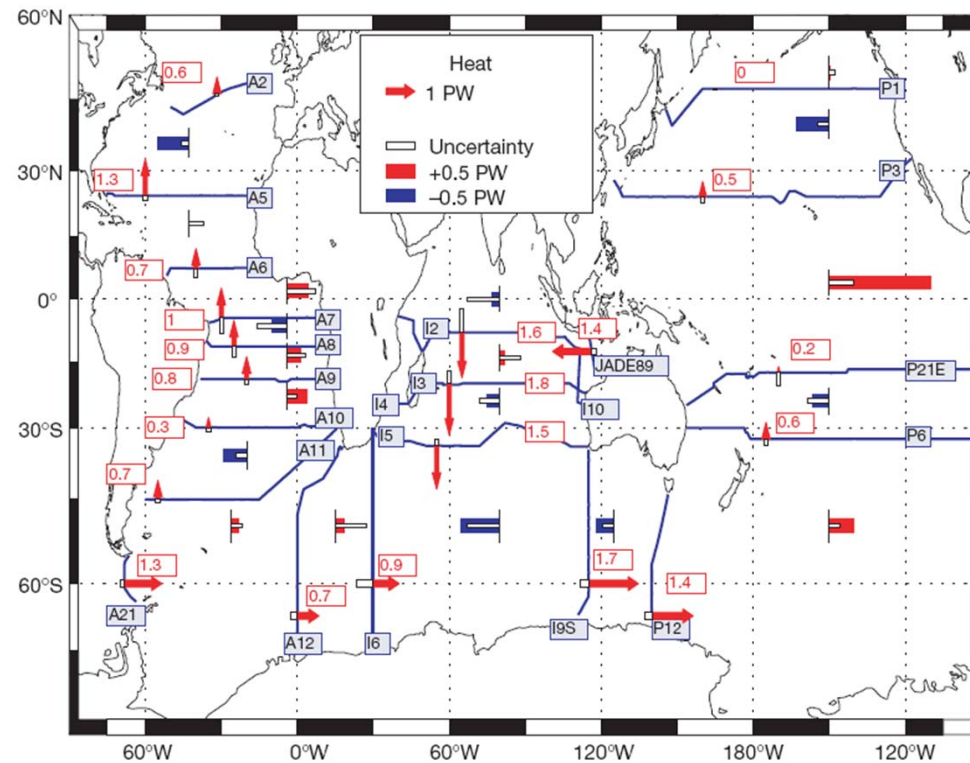
FIG. 7. The required total heat transport from the TOA radiation RT is compared with the derived estimate of the adjusted ocean heat transport OT (dashed) and implied atmospheric transport AT from NCEP reanalyses (PW).

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



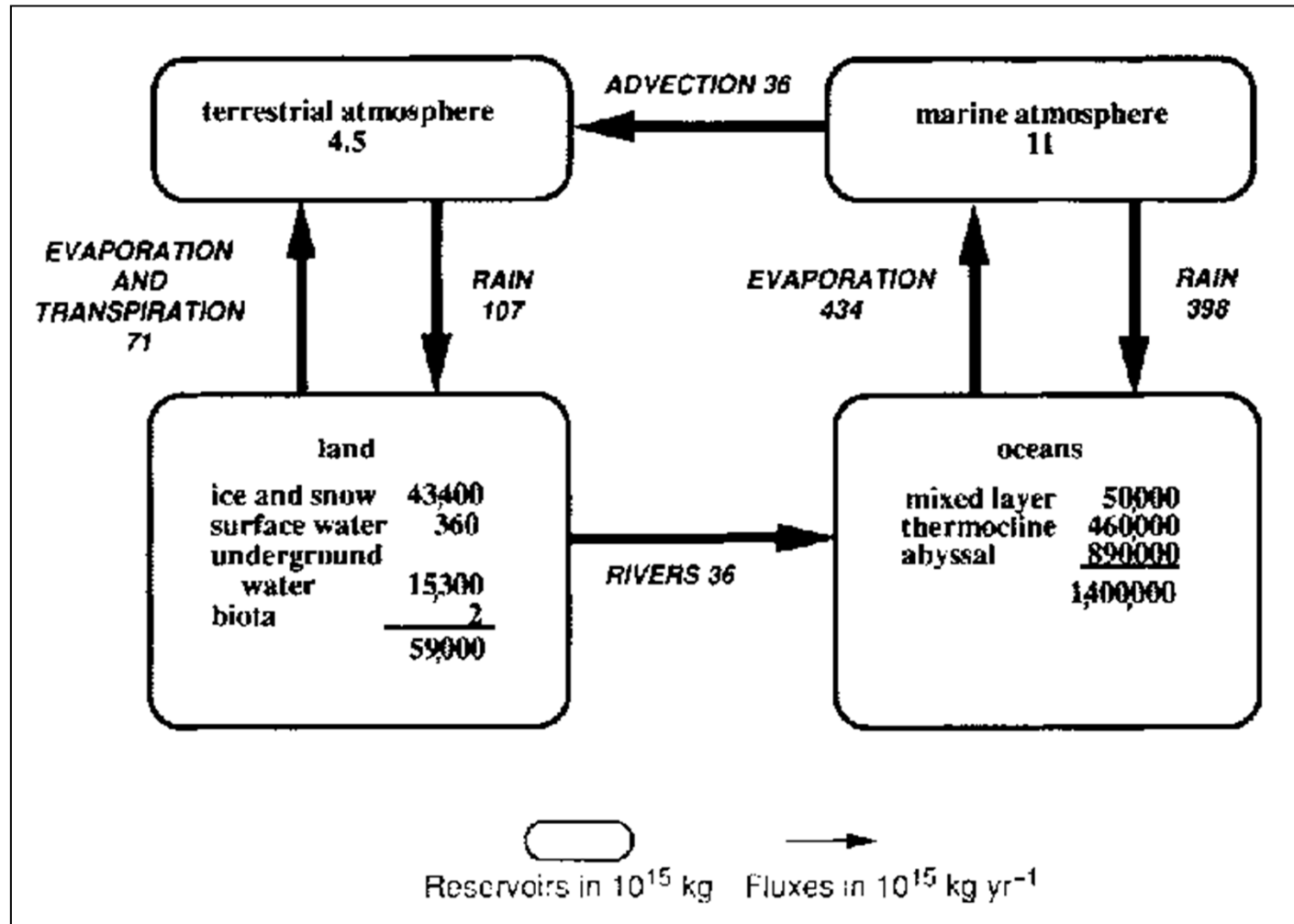
Ocean heat transport

Heat transport in the ocean can be estimated using an inverse model in which the flow is adjusted such that energy, mass and tracers are conserved and that the model reproduces measured horizontal and vertical transects of temperature, salinity .

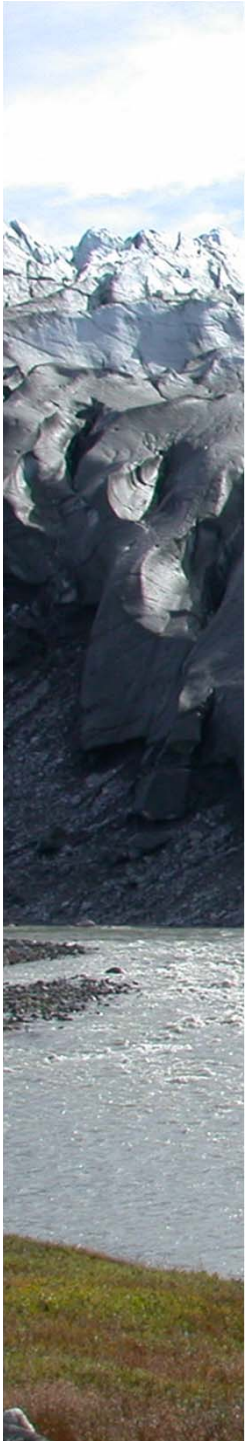


Ganachaud, A., and C. Wunsch (2000), Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data, *Nature*, **408**, 453-457.

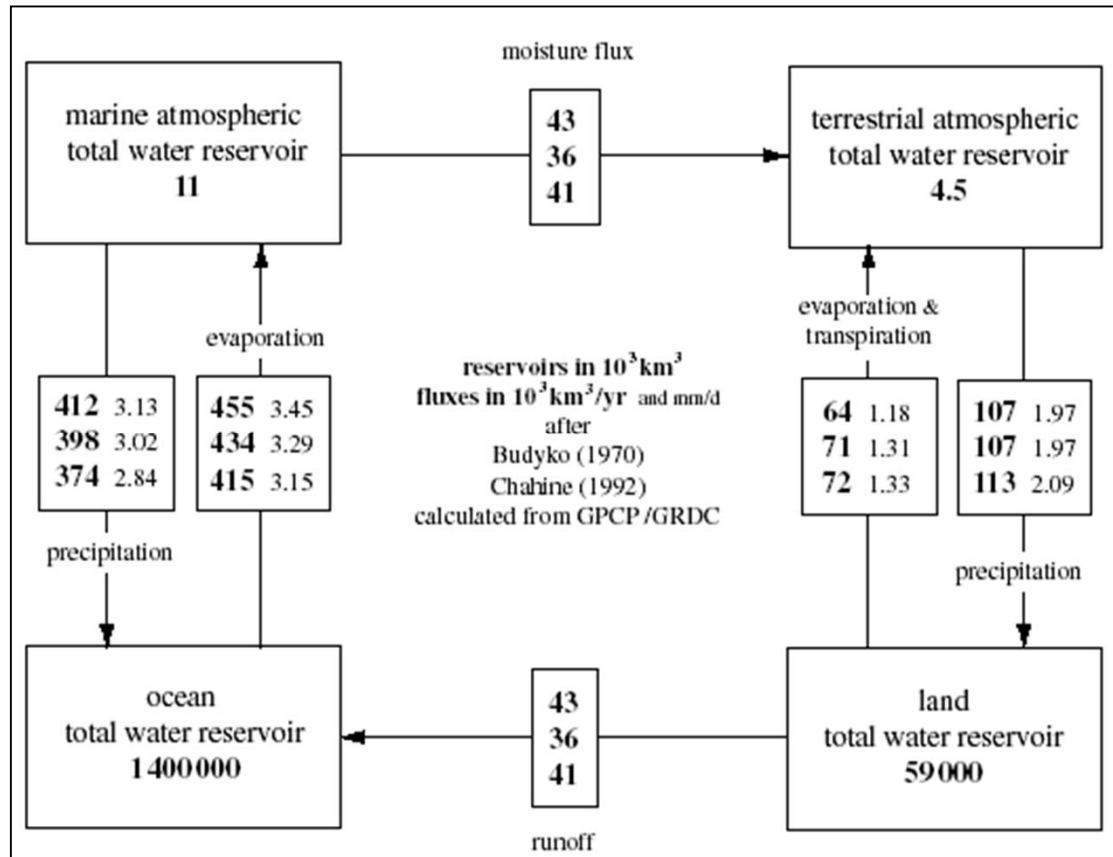
Global water balance (1)



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 $4 \pi r_E^2 = 5.1 \cdot 10^{14} \text{ m}^2 = 5.1 \cdot 10^8 \text{ km}^2$ 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 \equiv 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):

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

where

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

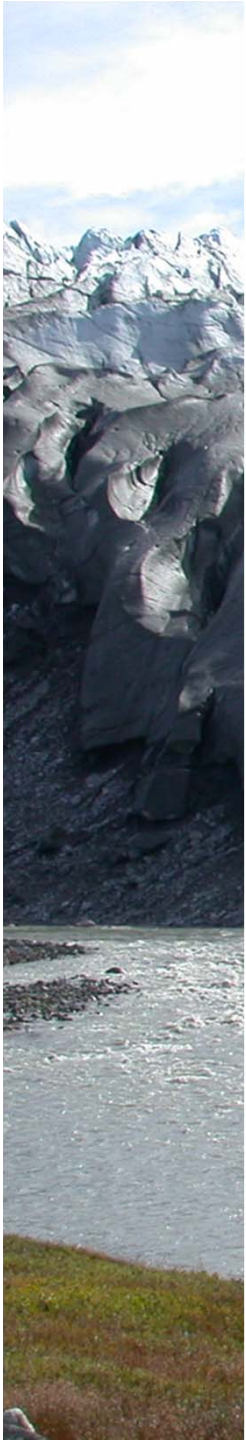
$\Delta = de_{\text{sat}}/dT$, gradient of the saturation water vapour pressure

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

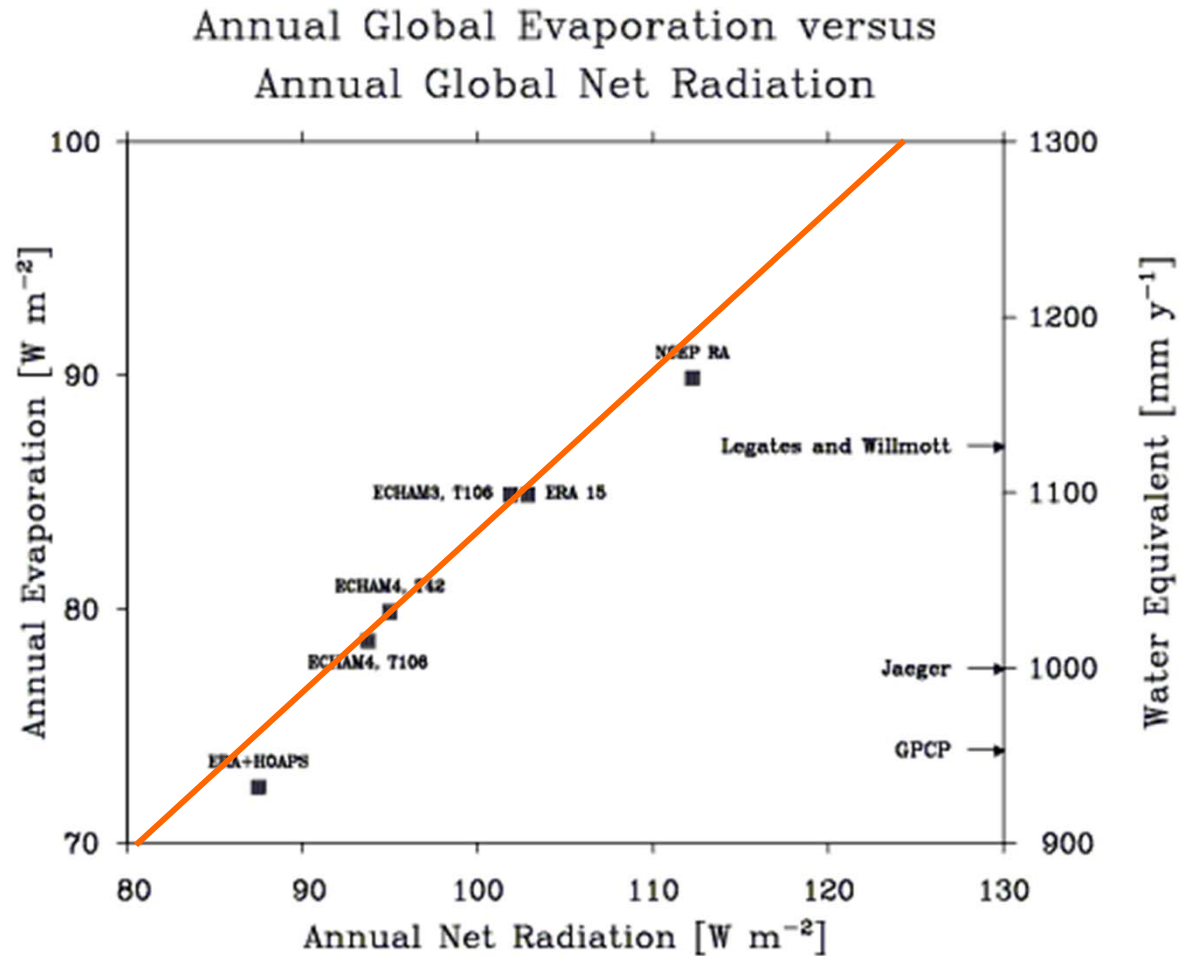
$\varepsilon = M_w/M = 0.622$, ratio of molecular weights of water and dry air.

In sum:

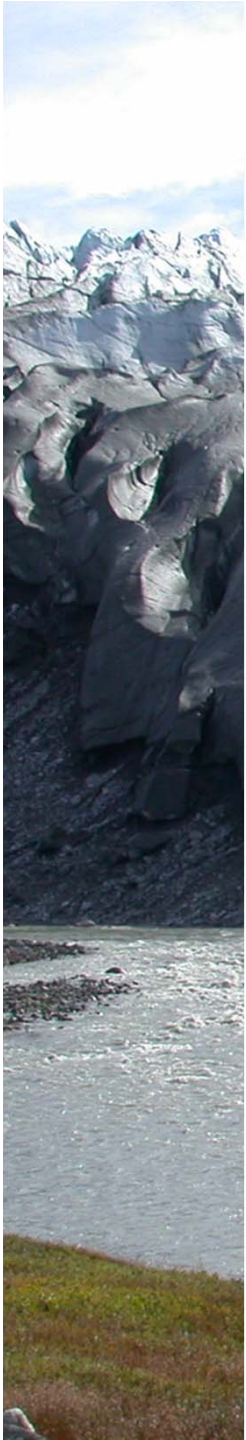
$$P \sim ET \sim NR$$



Global precipitation and the surface radiation budget (2)

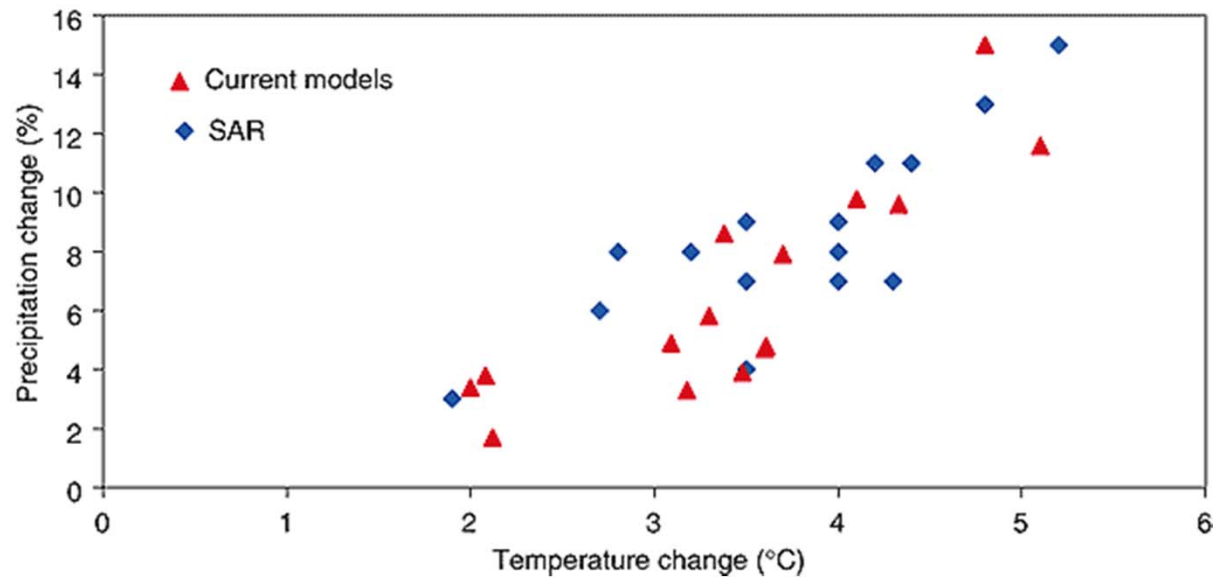


Ohmura (pers. comm.)



Global precipitation in a warmer climate

Global change in precipitation vs. climate sensitivity, the global equilibrium surface warming for a doubling of atmospheric CO₂ predicted from climate models.



IPCC Third Assessment Report (2001), Fig. 9.18

The hydrological cycle intensifies in a warmer world.

Radiation affects the hydrological cycle, and the water vapor in the atmosphere affects the radiation budget.