Aspects of Climate Change

1. **Forcing of the atmosphere and ocean circulation**
2. **Dynamics of the atmosphere**
3. **Dynamics of the ocean**

1. Forcing of the atmosphere and ocean circulation

The Earth

The Earth is almost perfect sphere.

Mean **radius** a = 6370 km

gravity field g = 9.81 m/s^2

rotation period τ = 24 h (this

corresponds to an **angular velocity** of $\Omega = 2\pi / \tau = 7.27 \times 10^5 \text{ s}^{-1}$).

The Atmosphere and Ocean

The atmosphere and ocean are thin films of fluid on the spherical Earth under the influence of:

(i) gravity
(ii) Earth's rotation
(iii) heating by solar radiation

Atmospheric constituents

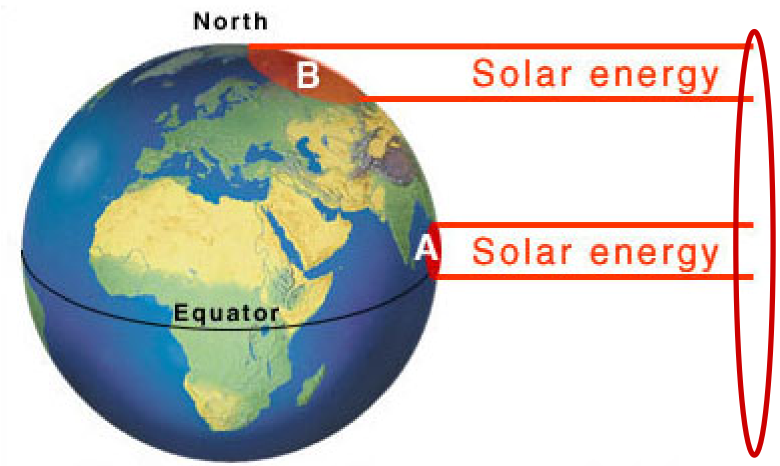
Atmosphere is mixture of ideal gases: N_2 and O_2 largest by volume, also CO_2, H_2O & O_3 .

Ideal gas: $p/\rho = RT$ (R gas constant, T temperature).

Atmospheric forcing

Forcing of atmosphere from Sun; interactions with land and ocean also important.

Incident solar flux, or power / unit area, of solar energy (the so-called solar constant) is $F = 1370$ W m^{-2}.



Power intercepted in tube of cross-sectional area πa^2, where a is Earth's radius.

Hence total solar energy received / unit time is $F \pi a^2$.

Albedo

Assume **albedo** of Earth is $\alpha = 0.3$

i.e., 30% of the incoming solar radiation is reflected back to space without being absorbed.

Final *incoming* power is

$$(1-\alpha)F\pi a^2 \qquad (1.1)$$

Black body

Assume Earth emits as **black body** at uniform absolute temperature T.

Stefan-Boltzmann law: power emitted / unit area = σT^4

(where σ is Stefan-Boltzmann constant)

Power emitted in all directions from a total surface area of $4\pi a^2$.

Final *outgoing* power is

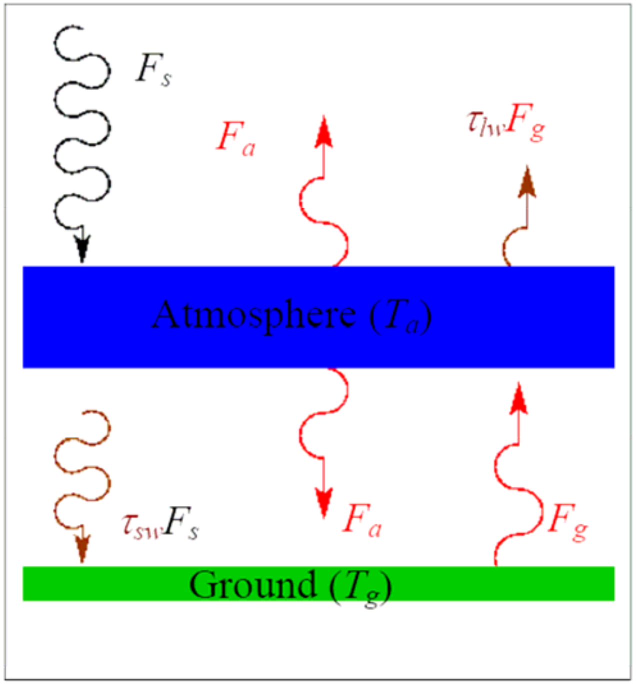
$$4\pi a^2 \sigma T^4 \qquad (1.2)$$

Power budget

By equating (1.1) and (1.2) and using standard values find that:

T =255K, but observed value is ~288K.

Greenhouse effect



Atmosphere temperature T_a, transmits fraction τ_{sw} shortwave and τ_{lw} longwave radiation, absorbs remainder.

From (1.1) mean incoming flux (power / unit area) $F_s = \frac{1}{4}(1-\alpha)F$.

Ground emits as black body, $F_g = \sigma T_g^4$

Atmosphere (not black body) emits[1] $F_a = (1-\tau_{lw})\sigma T_a^4$.

[1] Kirchhoff's Law

Top of atmosphere $F_s = F_a + \tau_{lw} F_g$ and ground $F_g = F_a + \tau_{sw} F_s$.

If τ_{sw}=0.9 and τ_{lw}=0.2, find T_g=286K.

Greenhouse effect: greater temperature from greater transmission for shortwave vs longwave radiation.

Radiative transfer

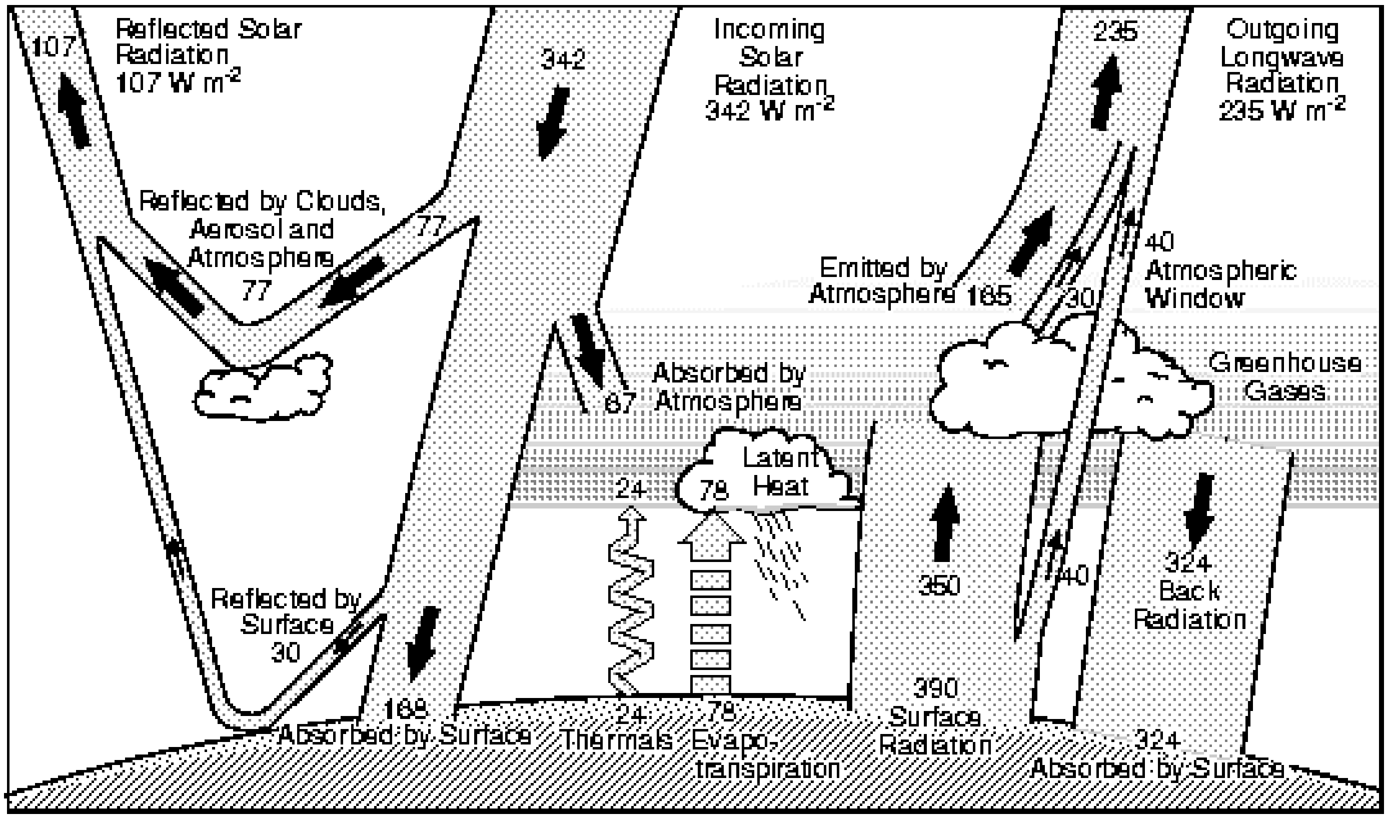
Black-body emission

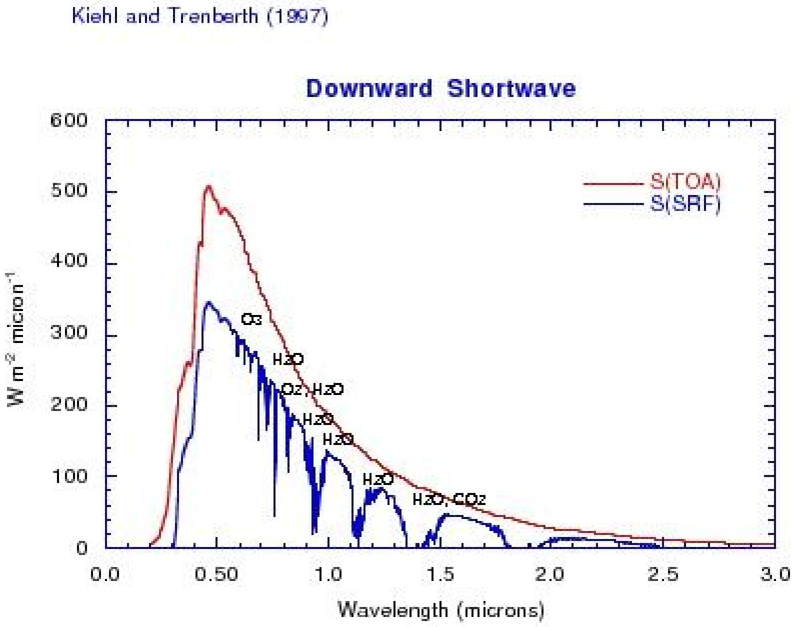
6000K (solar) UV, visible & IR wavelengths 0.1 to 4μm (shortwave);

288K (Earth) IR wavelengths 4 to 100 μm (longwave).

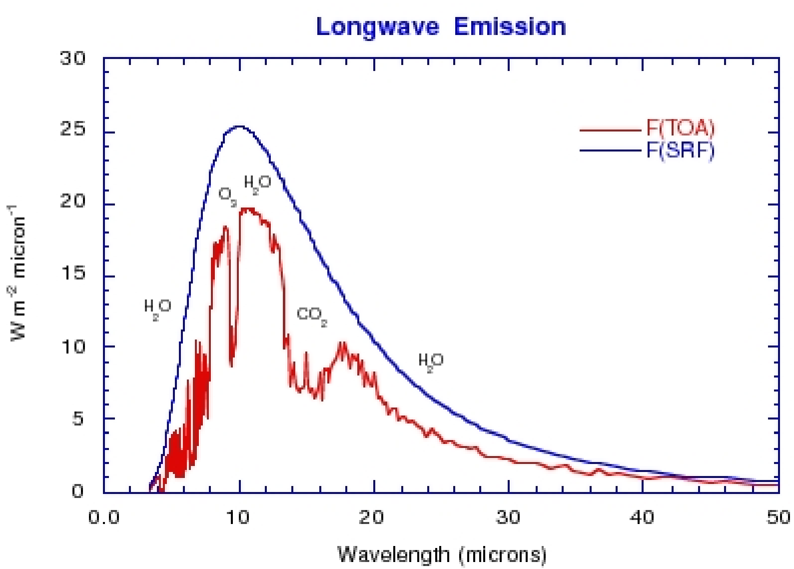
Global heat balance

Absorbing gases: ozone (O_3) in UV & visible; carbon dioxide (CO_2) & water vapour (H_2O) in IR





Shortwave radiation - scattered by atmospheric gases, or reflected clouds/ground back to space; absorbed by (& heats) atmospheric gases (H_2O, O_3)/clouds/ground



Longwave radiation - emitted & absorbed by atmospheric gases (CO_2, H_2O, O_3), clouds/ground: heat transfer, or heat lost to space.

Radiative forcing

Greenhouse effect: difference between surface & top-of-atmosphere emission of longwave radiation. [8-12μm atmospheric window (absorption weak); 9.6μm O_3; 6.3μm & >16mm H_2O; CO_2 15μm.]

Radiative forcing (IPCC): change in net irradiance at tropopause (~10-15km). Radiative forcing due to doubling CO_2 estimated ~4 Wm^{-2}. Without feedbacks gives surface temperature rise of ~1.2°C.

Climate sensitivity (IPCC): equilibrium change in global mean surface temperature due to doubling CO_2 (takes into account feedbacks). Likely in range 2-4.5°C; best guess 3°C.

Aerosols

Aerosol: particles (0.1-10μm diameter). Sulphate, fossil fuel organic & black carbon, biomass burning, mineral dust, sea salt.

Direct effect: scatter (negative RF) and absorb short(long)wave radiation (positive or negative RF)

Indirect effect: alter cloud microphysics (hence radiative properties), amount (cloud condensation nuclei) & lifetime. Potential for large impact from small change (negative RF).

Volcanoes and solar variability

Solar output increased gradually over industrial era (in addition to 11-year cycle) causing small +RF.

Explosive volcanic eruptions can lead to short-lived (few yrs) –RF from sulphate aerosol in stratosphere[2].

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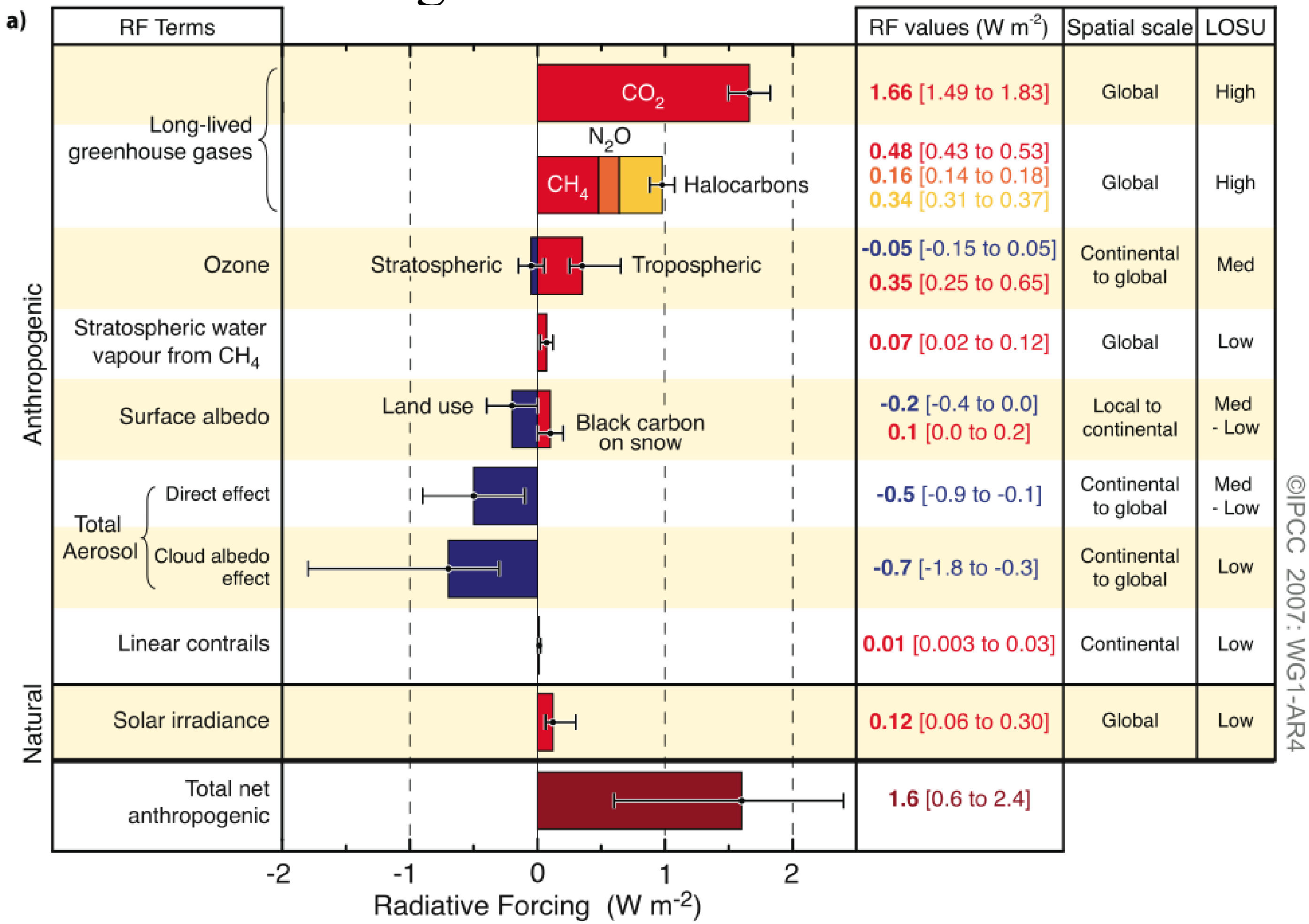
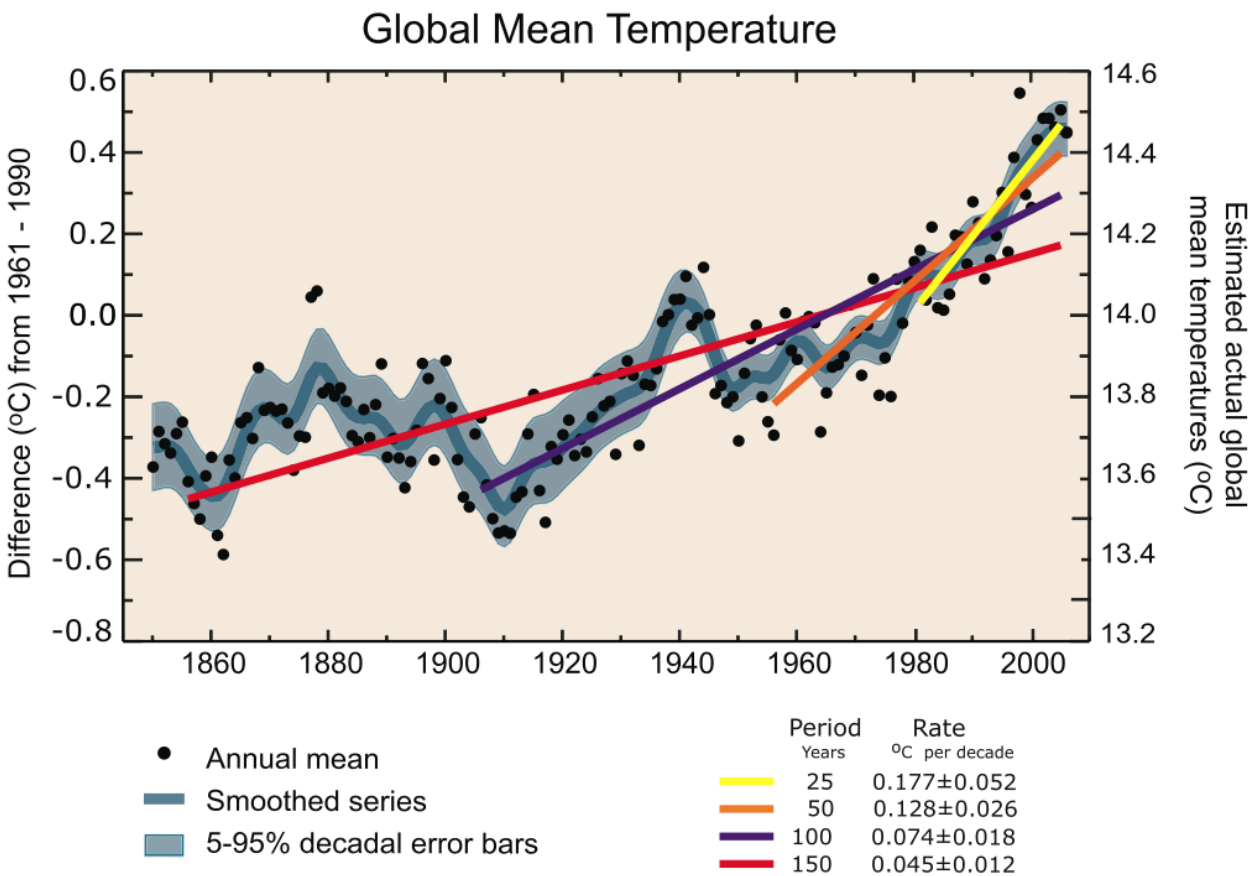
[2] Last major eruption Mt Pinatubo (1991)

Estimate of radiative forcing (IPCC)

Radiative forcing of climate between 1750 and 2005



Climate change over the past century

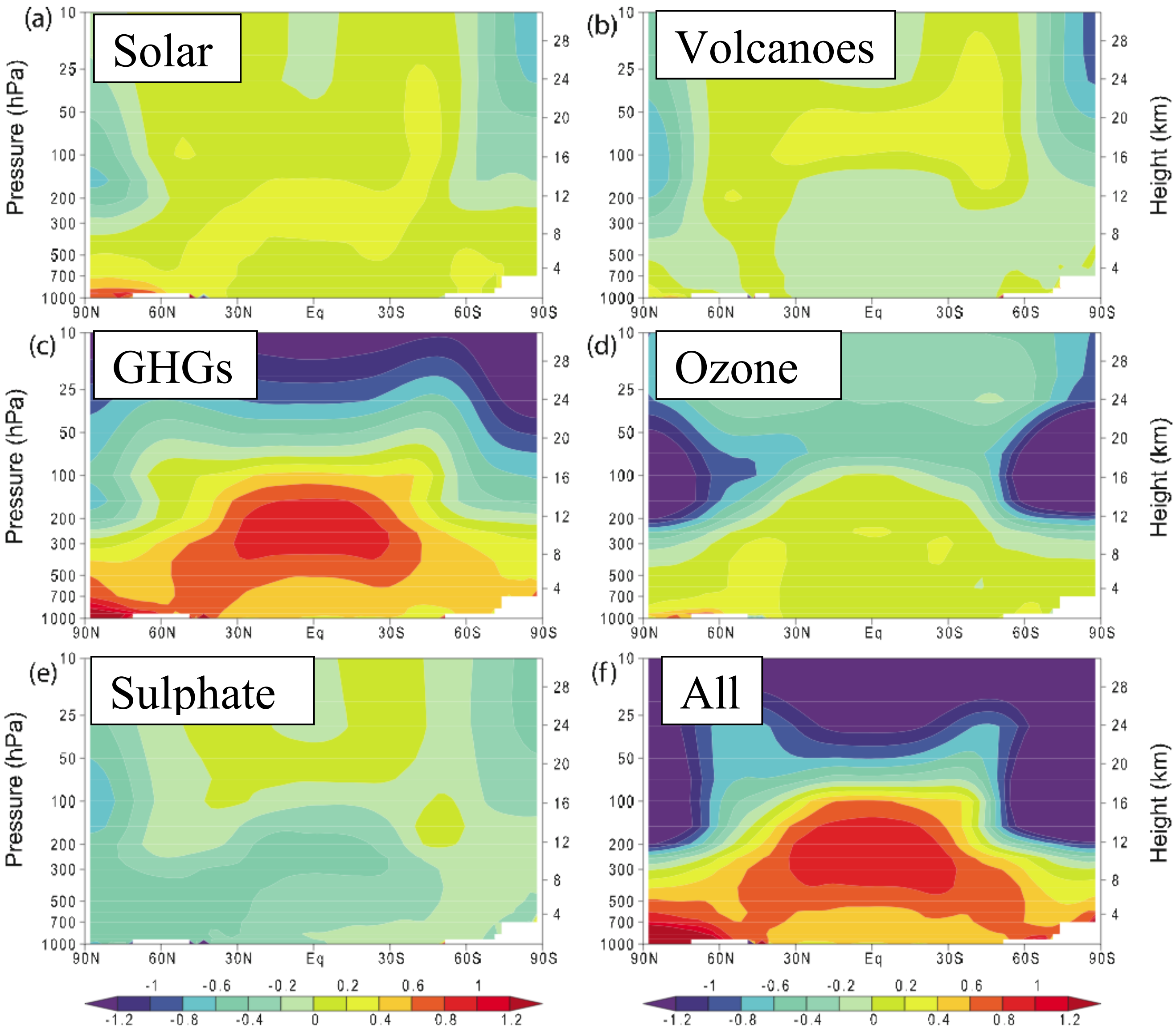


Earth's surface has warmed 0.76 ± 0.19K since 1850.

[IPCC fig 3.3].

Tropospheric humidity increasing by 1.2%/decade [fig 3.20]

Attribution



Optimal fingerprinting. Zonal mean atmospheric temperature change from 1890 to 1999 (K/century) [fig 9.1]

Physics of atmosphere

Each portion of atmosphere approx in **hydrostatic balance** (usually valid on scales > few km)

i.e., weight supported by pressure difference pressure between lower & upper surfaces.

$$g\rho = -\frac{\partial p}{\partial z},$$ ρ is density, p is pressure.

Result of hydrostatic balance & ideal gas law: typically pressure and density fall exponentially with height.

For atmosphere with temperature T_0

$$\frac{gp}{RT_0} = -\frac{\partial p}{\partial z}, \qquad p = p_0\, \mathrm{e}^{-gz/RT_0} \qquad (1.3)$$

$\dfrac{RT_0}{g}$ is '**scale height**' (~7 km if $T_0 = 240K$); p decreases upwards.

Density stratification

Gravity produces **density stratification**. An air parcel displaced upwards (*downwards*) from its equilibrium position is negatively (*positively*) buoyant & will fall (*rise*) back under gravity. Buoyancy acts as restoring force; atmosphere is **stably stratified**.

Thermodynamics

First law of thermodynamics: the increase in internal energy of a system δU equals heat supplied plus work done on the system, i.e. $\delta U = T\delta S - p\delta V$, where S is the **entropy** and V .is the volume.

For a unit mass of ideal gas with $V = 1/\rho$, then have $U = c_V T$, where c_V is specific heat at constant volume ($c_p = c_V + R$ is specific heat at constant pressure). Hence,

$$\delta S = c_p \frac{\delta T}{T} - R\frac{\delta p}{p} \tag{1.4}$$

Potential temperature

Adiabatic process: no gain/loss heat, $\delta S = 0$.

Cylinder of air, temperature T & pressure p, compressed adiabatically until temperature θ & pressure p_0.

Integrating (1.4) gives $\theta = T(p_0 / p)^{\kappa}$ where $\kappa = R / c_p$.

θ is **potential temperature**

θ is conserved (as is entropy) in adiabatic motion. For stable atmosphere, θ increases upwards, ('**isentropic co-ordinate**').

Lapse rate

For adiabatically rising parcel, entropy (and θ) constant as height changes. Hence from (1.3) and (1.4),

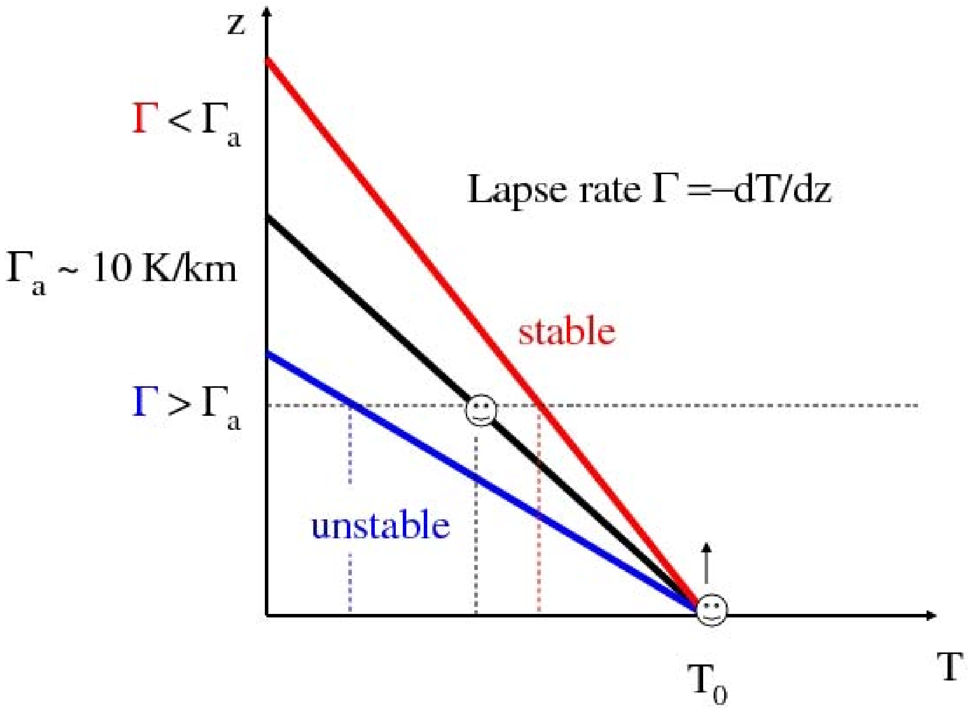
$$-\left(\frac{dT}{dz}\right)_{parcel} = \frac{RT}{c_p p}\left(\frac{dp}{dz}\right)_{parcel} = \frac{g}{c_p} \equiv \Gamma_a .$$

Γ_a, **adiabatic lapse rate**, is rate of decrease of temperature with height following the adiabatic parcel as it rises.

Dry adiabatic lapse rate ~10 K km^{-1}.

Convection

Convection occurs only if atmospheric lapse rate exceeds certain value (e.g. dry adiabatic lapse rate).



If background temperature falls more quickly with height, a rising parcel is warmer than surroundings & continues to rise under own buoyancy: instability.

Convection carries heat up & thus reduces lapse rate until equilibrium value.

Latent heat

Latent heating/cooling can transfer heat (e.g., evaporation of droplet of sea-water & condensation into droplet at another location in atmosphere transfers heat from ocean to atmosphere).

Moist convection

As a parcel rises adiabatically, p falls, so T falls, water vapour condenses, latent heat released.

Moist adiabatic lapse rate less than for dry air (more easily exceeded).

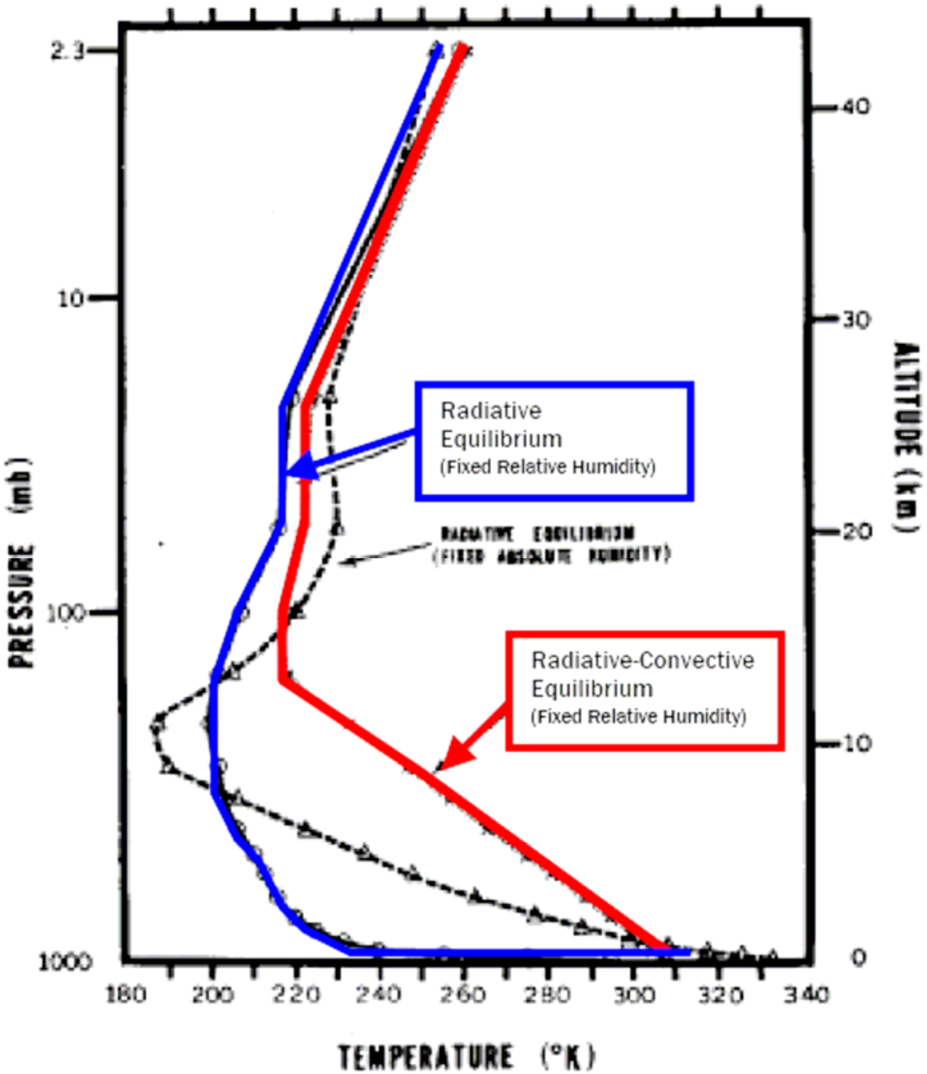
(For descending air, dry adiabatic lapse rate is relevant.)

Radiative-convective model

1-D radiative equilibrium calculation predicts temperature sharply decreasing with height at lower boundary, implying convectively unstable.

Radiative-convective calculation adjusts temperature gradient to neutral stability where necessary (takes account of moisture).

Manabe & Wetherald (1967): radiative-convective calculation with fixed relative humidity. Simplest possible model including combined effect of fluid dynamics and other physical processes.



Radiative-convective equilibrium calculations[3]

Convective region

troposphere

Radiative region

stratosphere

[3] S. Munabe and R.T. Wetherald, Journal of the Atmospheric Science, 24, 241–259

Oceans

71% of Earth's surface covered by water.

Average depth of 3.7 km, but sometimes exceeds 6km.

Heat capacity of upper 3m ocean ≈ entire atmosphere.

Ocean stores 50 times more carbon than atmosphere and takes up roughly 1/3 of carbon released into atmosphere by human activities.

Changes in sea surface temperature (SST) can affect atmosphere (e.g. hurricanes, El Niño).

Forcing of ocean

Forcing of the ocean from:

(i) wind stress

(ii) heat flux

(iii) freshwater flux

(iv) tides

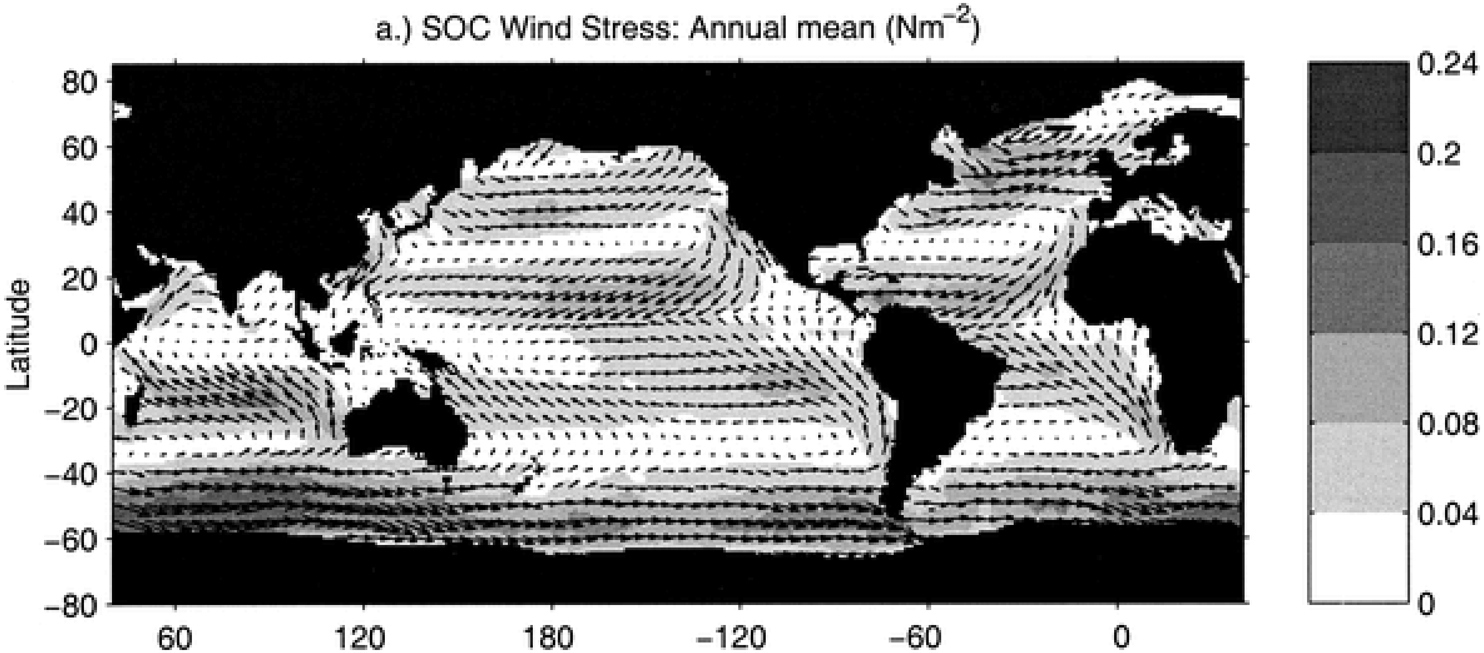
Wind stress: winds exert stress on surface[4]: drives ocean currents.

Highly variable; much uncertainty.

$\tau_s = \rho_a U^2 C_D$ ρ_a is density of air; U^2 = wind speed at 10m,

C_D = drag coefficient (function of wind speed, atmospheric stability & sea state).

C_D= 0.00115 when $|U| < 11$m/s and $(4.9+0.065|U|)\times10^{-4}$ otherwise.



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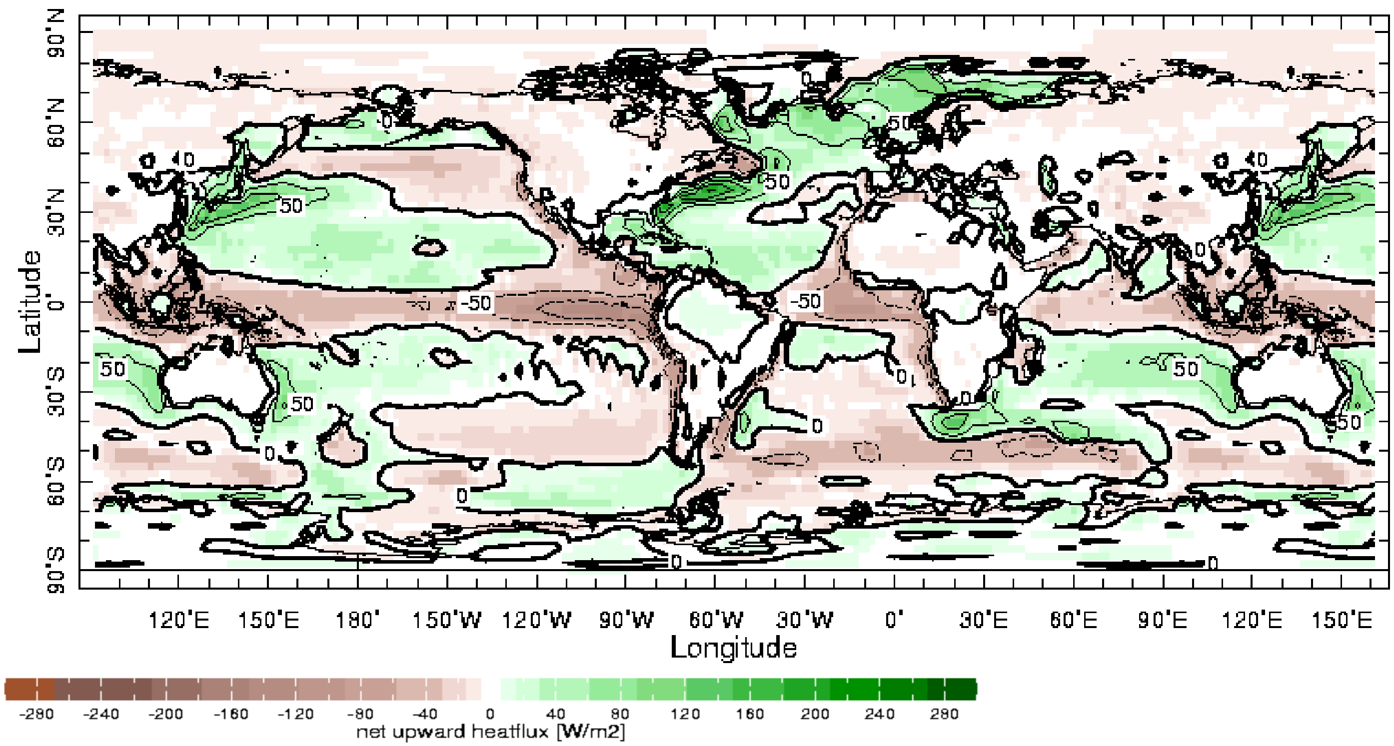
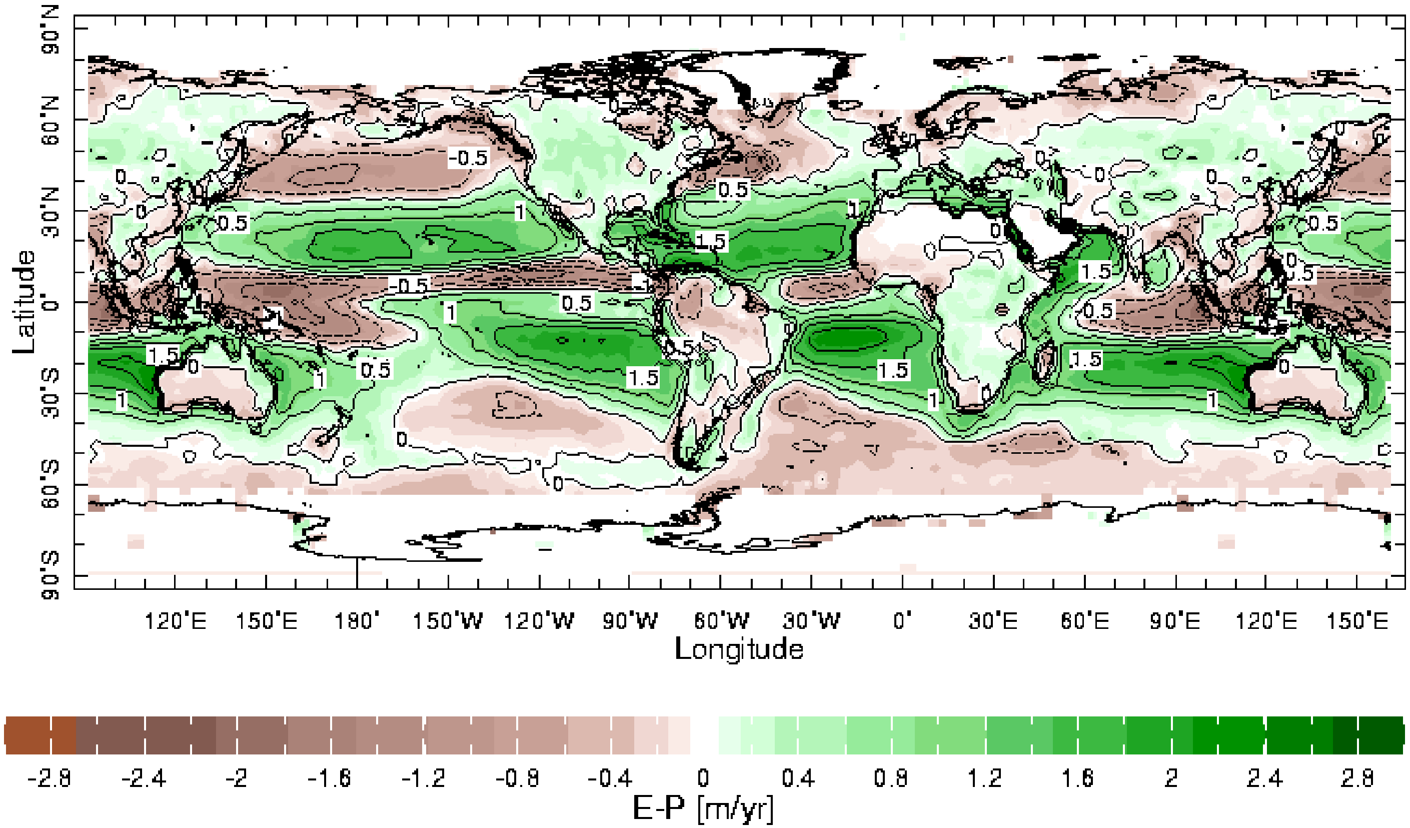
[4] through turbulent transfer of momentum across atmospheric boundary layer

Heat flux has 4 components: (i) sensible heat flux (air/sea temperature difference); (ii) latent heat flux (evapouration); (iii) incoming shortwave radiation from sun; (iv) longwave radiation from atmosphere & ocean



Freshwater flux from evaporation and precipitation



Tides are also an important force on the ocean.