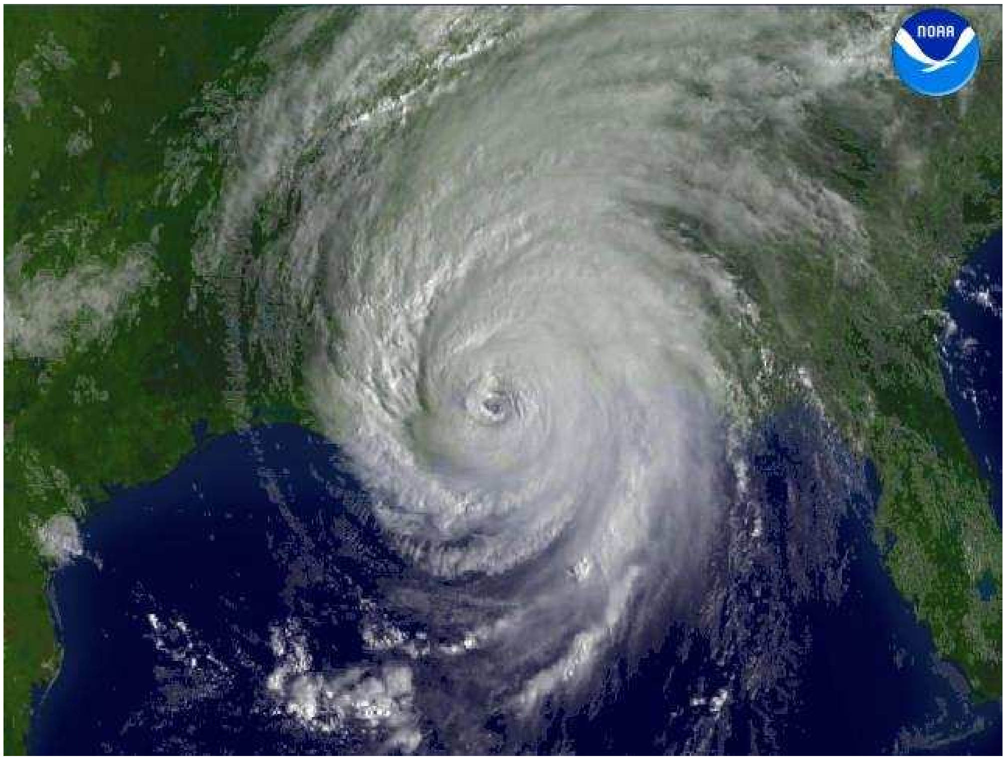
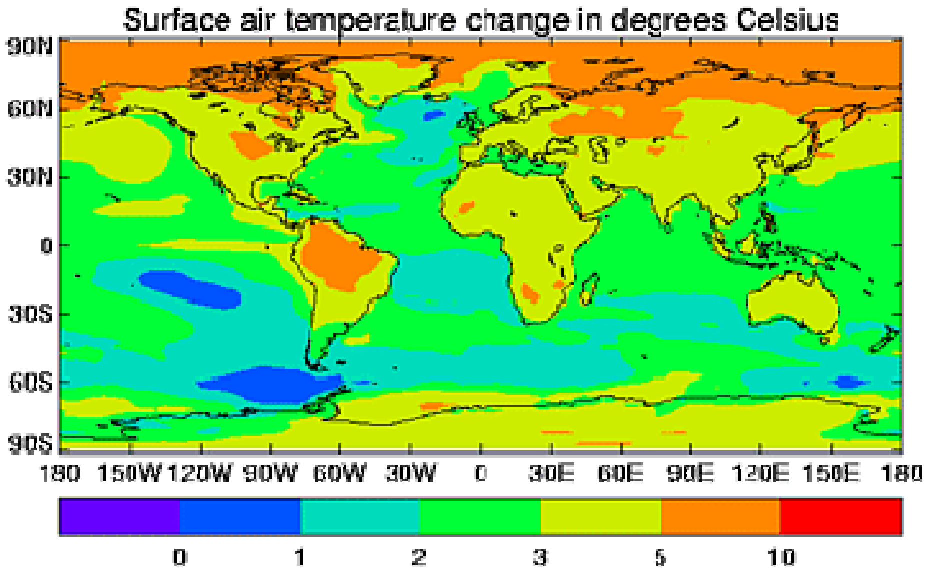
Weather and Climate





1. Introduction to the Atmosphere

2. Baroclinic Instability and Midlatitude Dynamics

3. Midlatiude Storm Tracks and the North Atlantic Osciallation

4. Equatorial Waves and Tropical Dynamics

5. El Nino Southern Oscillation

1. Introduction to the Atmosphere

Weather and Climate

Weather: events associated with atmospheric flows with length scales of hundreds of metres and more, and time scales of a few days or less.

Irregular, unpredictable (chaotic) behaviour characteristic of nonlinear evolution equations.

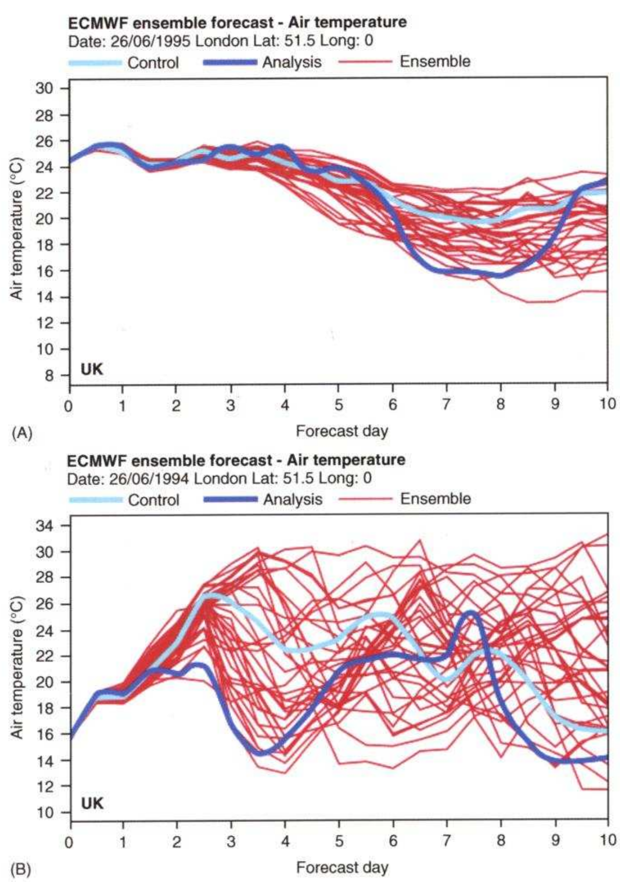


Figure 4 Forecasts given by the ECMWF operational ensemble prediction system (33 members at the time of running) temperature in London started from (A) 26 June 1995 and (B) 26 June 1994. (Courtesy of Thomas Petroliagis, 1995, pe communication.)

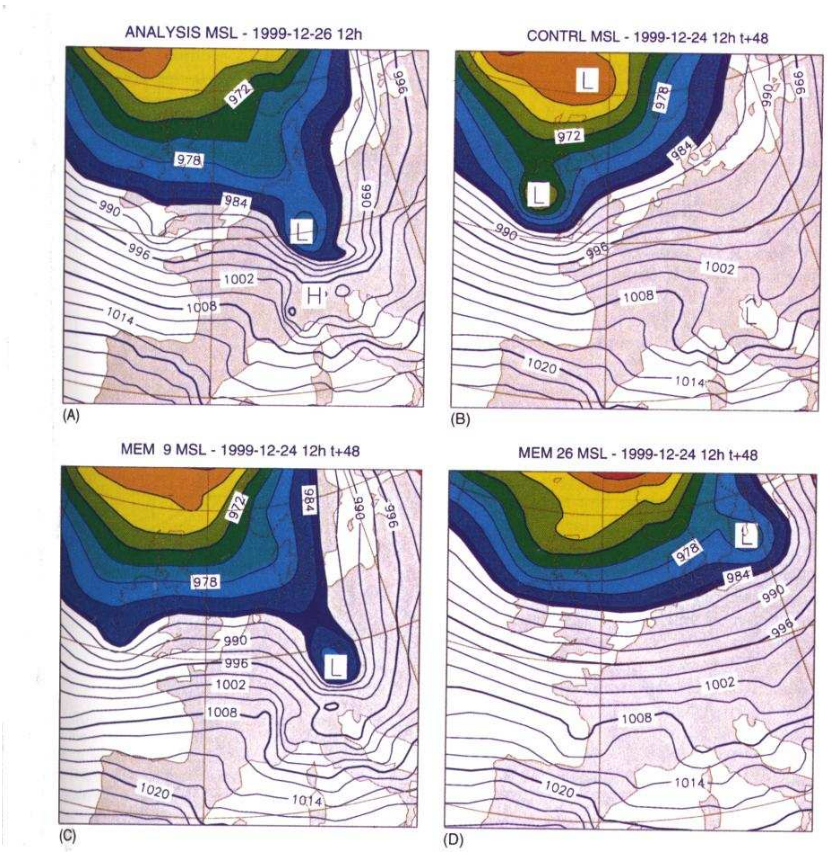
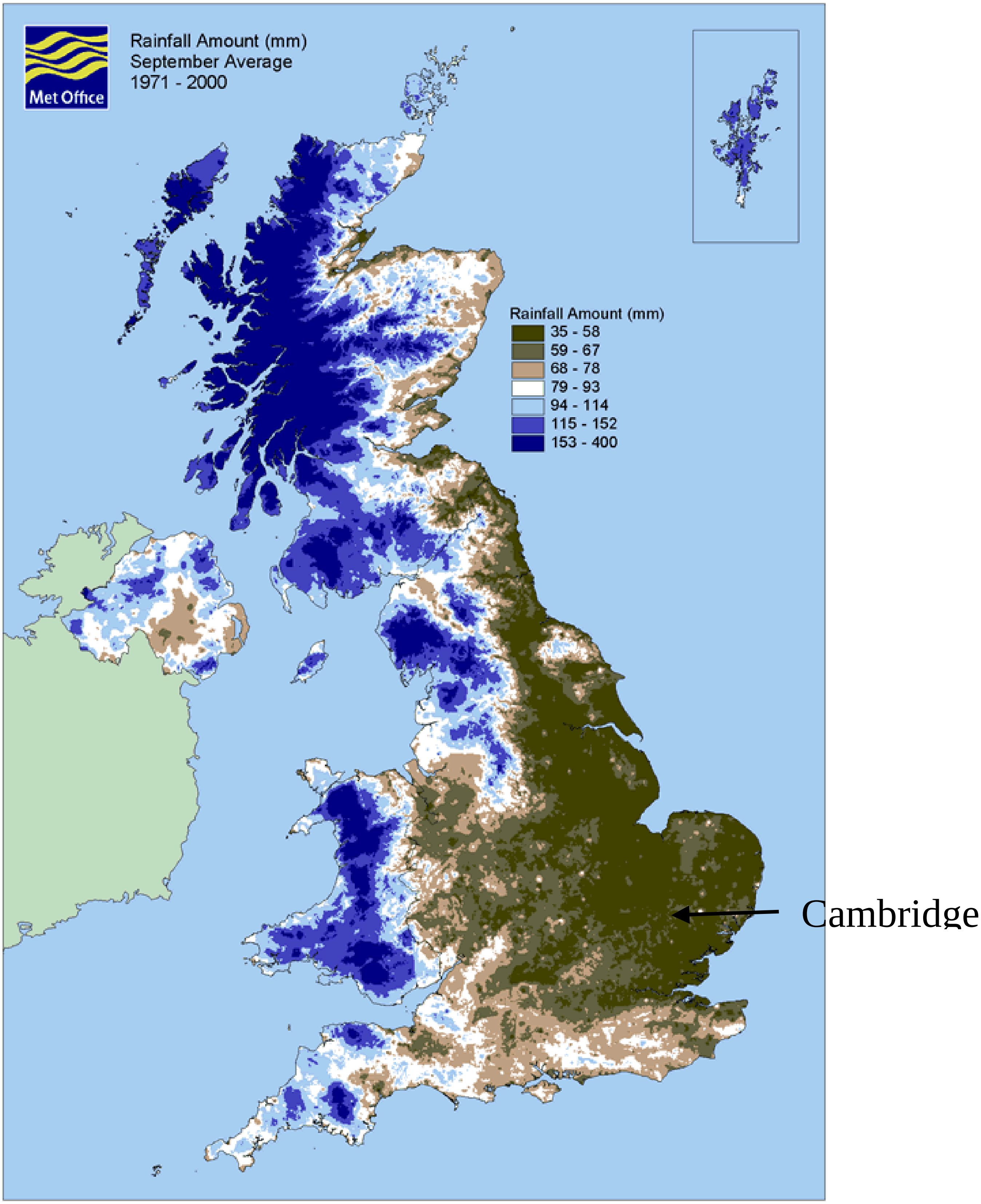


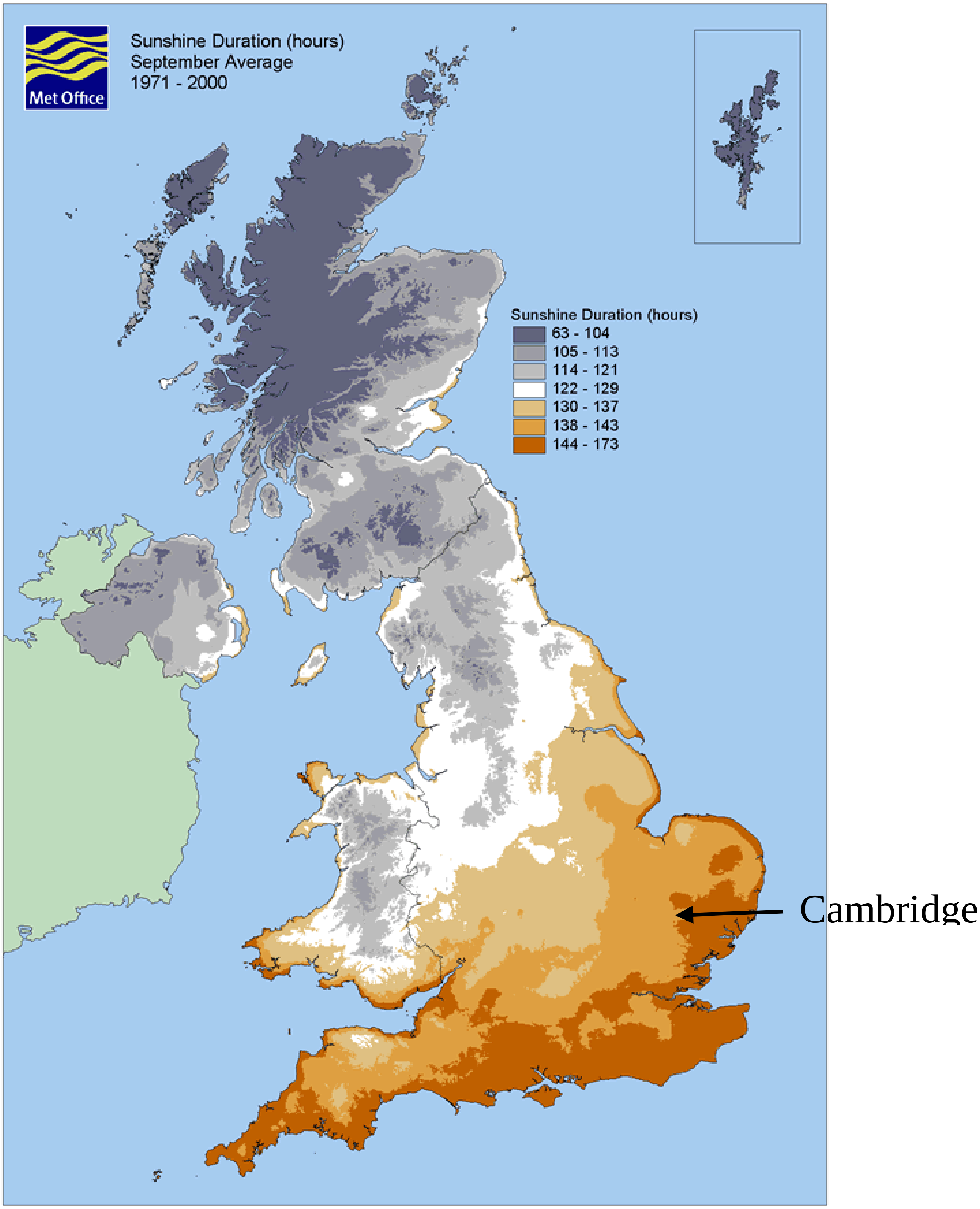
Figure 3 (A) Mean sea-level pressure field at 1200 UTC on 26 December 1999 and (B–D) 2-day mean sea-level pressure forecasts started at 1200 UTC on 24 December 1999 given by three members of the ECMWF ensemble prediction system. Contour interval is 3 hPa, with shading for values lower than 984 hPa.

Different components vary on different timescales (predictability individual clouds < 1 hour, midlatitude weather systems ~ days).

Atmospheric data averaged over ~month more regular.

Climate: state of atmosphere on longer time scales, several years or more. Characterised by probability distribution of variable atmospheric flow.


Met Office
Rainfall Amount (mm)
September Average
1971 - 2000
Rainfall Amount (mm)
35 - 58
59 - 67
68 - 78
79 - 93
94 - 114
115 - 152
153 - 400
Cambridge


Met Office
Sunshine Duration (hours)
September Average
1971 - 2000
Sunshine Duration (hours)
63 - 104
105 - 113
114 - 121
122 - 129
130 - 137
138 - 143
144 - 173
Cambridge

Atmospheric constituents

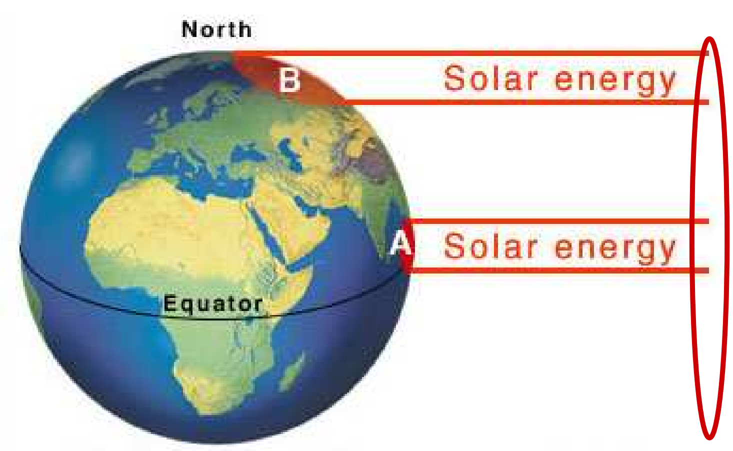
Atmosphere is mixture of ideal gases: N_2 and O_2 largest by volume, also CO_2, H_2O & O_3 .

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. So 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.

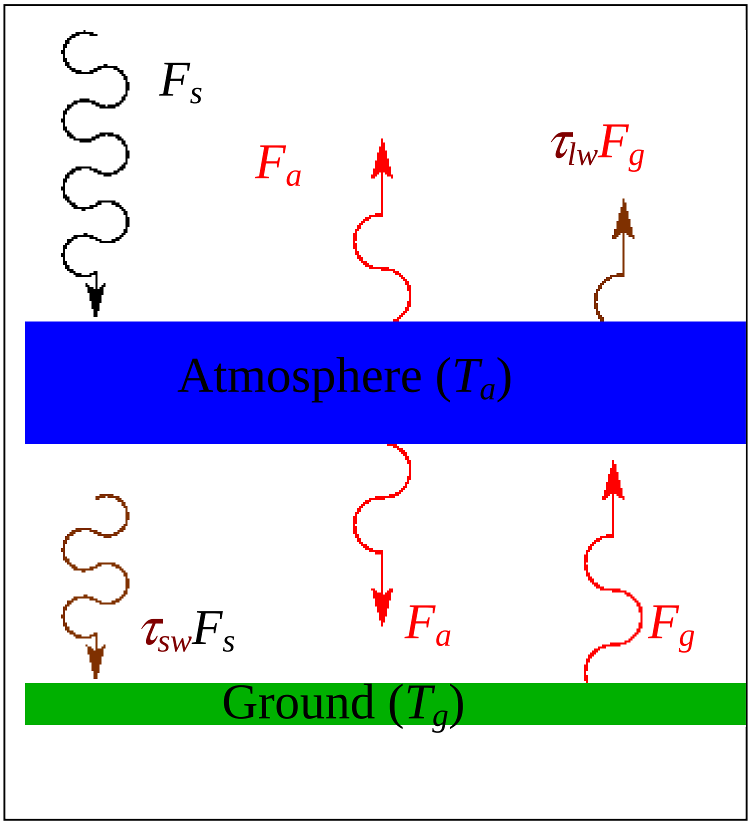
Stefan-Boltzmann law: power emitted / unit area $= \sigma T^4$, where σ is Stefan-Boltzmann constant.

Power emitted in all directions from a total surface area of $4\pi a^2$. So final outgoing power is

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

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 and absorbs remainder. From (1.1) mean incoming flux (power / unit area)

$$F_s = \frac{1}{4}(1-\alpha)F.$$

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

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

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.

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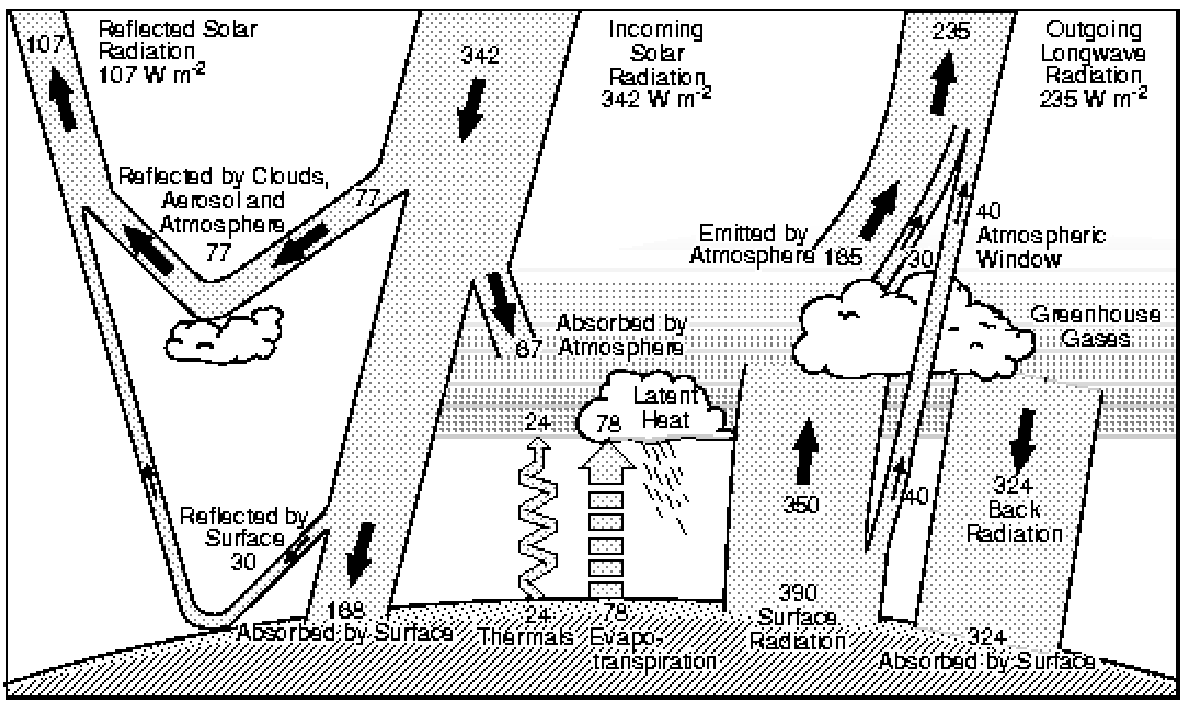
[1] Kirchhoff's Law

Radiative transfer

Solar shortwave radiation - scattered by atmospheric gases, or reflected clouds/ground back to space; absorbed by atmospheric molecules (H_2O & O_3)/clouds or ground: heats parts of atmosphere or ground.

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



Schematic of global heat balance[2]

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[2] Earth's Annual Global Mean Energy Budget, Kiehl, J. T. and Trenberth, K. E., 1997, Bull. Amer. Meteor. Soc., 78, 197-208

Hydrostatic balance

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.

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

Result of hydrostatic balance and ideal gas law is that typically pressure and density fall exponentially with height. For atmosphere with temperature T_0

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

Here $\frac{RT_0}{g}$ is '**scale height**' (~7 km if $T_0 = 240K$)

p decreases upwards – can be used as a vertical co-ordinate (measure of total mass above a certain height).

Density stratification

Gravity produces **density stratification**. A small portion of air displaced upwards (*downwards*) from its equilibrium position is negatively (*positively*) buoyant compared to its surroundings and will fall (*rise*) back towards equilibrium 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**.

For ideal gas, $U = c_V T$, 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** - 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 lapse rate exceeds a certain value (dry adiabatic lapse rate if little water vapour).

If temperature of surroundings falls more quickly with height, a rising parcel would be warmer than surroundings and continue to rise under own buoyancy: instability.

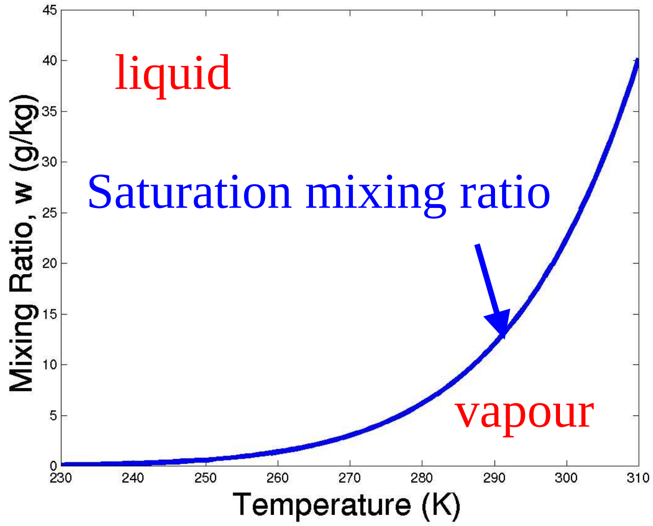
Convection carries heat up and thus reduces lapse rate until equilibrium value.

Latent heat

Latent heating/cooling provides important contribution to heat transfer.

E.g., evaporation of droplet of sea-water and condensation of resulting water vapour into droplet at another location in atmosphere transfers heat from ocean to atmosphere.

Water vapour



Amount of **water vapour** in air parcel given by **mixing ratio**[3] (mass water vapour / mass dry air) or **partial pressure** (pressure exerted by only water vapour part of air parcel).

Limited by **saturation mixing ratio** (when condensation and evaporation are in equilibrium).

Amount of water vapour relative to saturation value is **relative humidity** (at 100%, water droplets condense out forming clouds).

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[3] Mixing ratios up to 25 g/kg in humid tropical air masses.

Moist convection

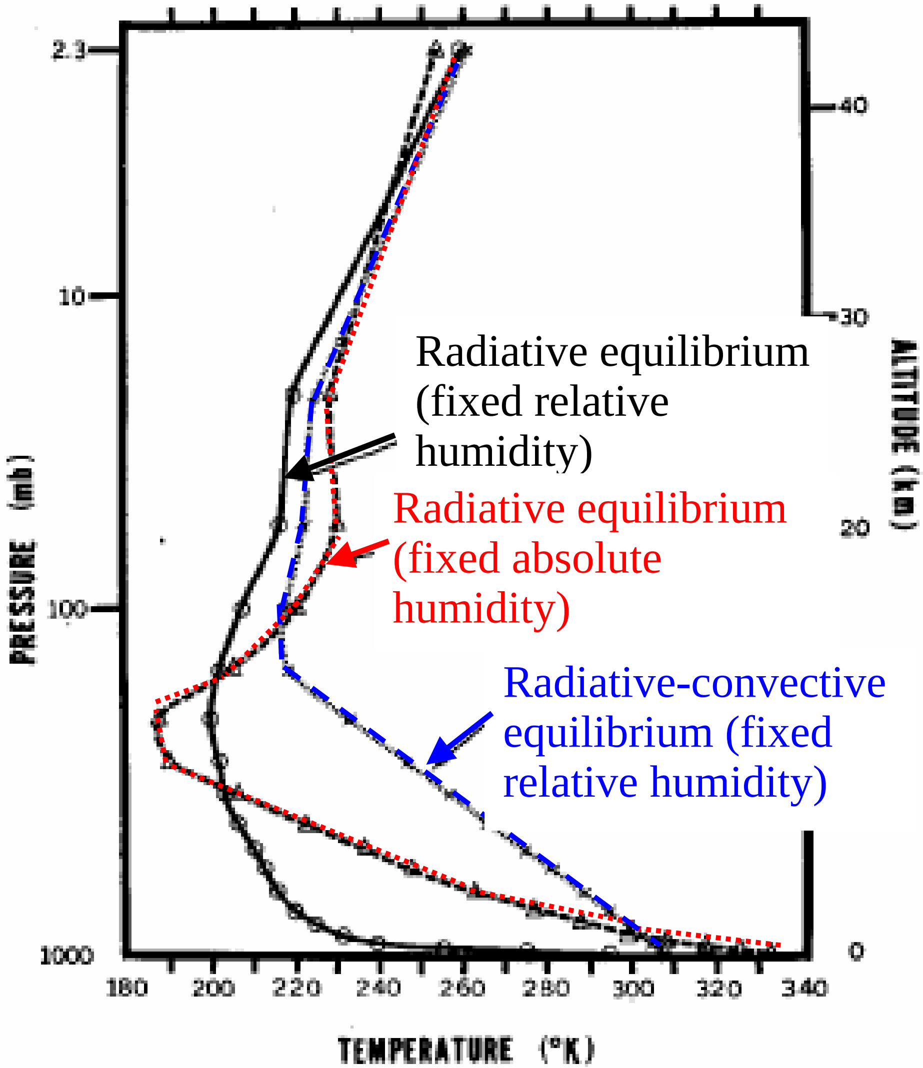
As a parcel rises adiabatically, p falls, so T falls, saturation mixing ratio decreases, water vapour condenses, latent heat released. **Moist adiabatic lapse rate** less than for dry air (more easily exceeded). For descending air, saturation mixing ratio increases so dry adiabatic lapse rate relevant.

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.



Fluid motion acts to reduce temperature gradients.

Radiative-convective equilibrium calculations[4]

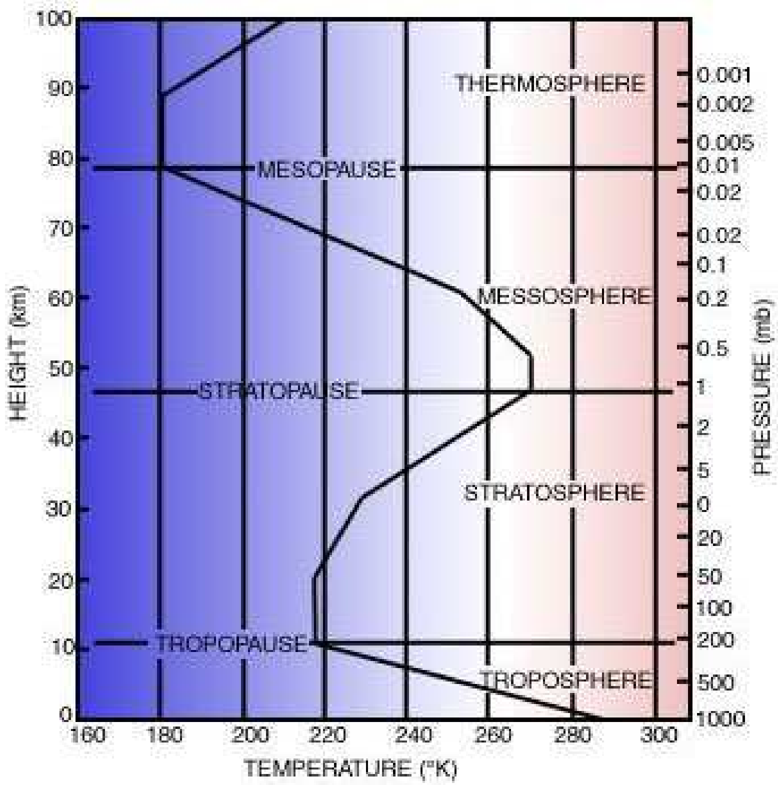
Convective region **troposphere**

Radiative region **stratosphere**

Top of convective region **tropopause**

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

Vertical temperature variation



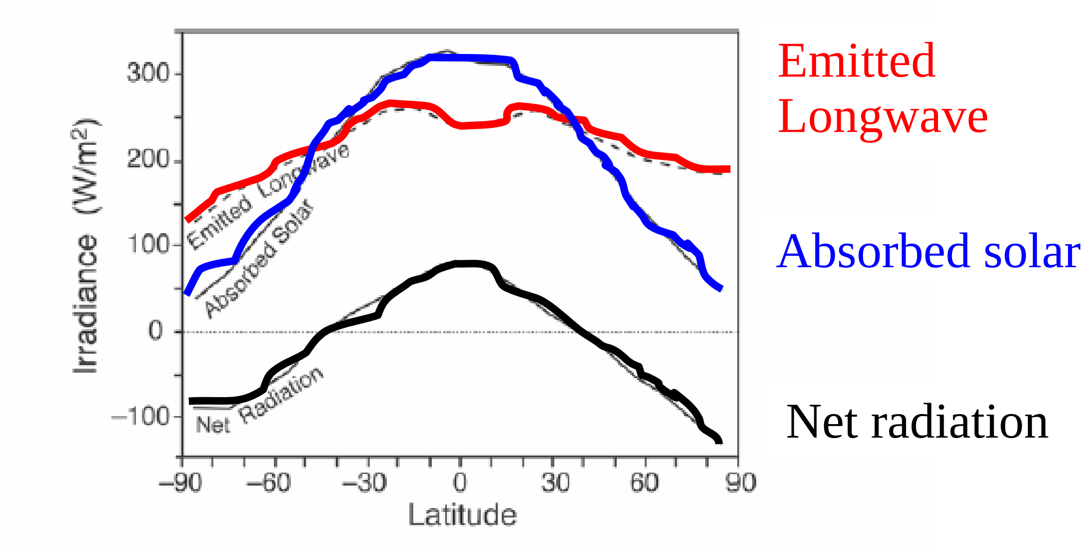
Hydrostatic balance: pressure proportional to mass of air above. ~90% of mass in troposphere and ~10% in stratosphere.

Vertical temperature profile of the ICAO Standard Atmosphere

Ground to ~15km altitude, temperature decreases with height: troposphere (bounded above by tropopause).

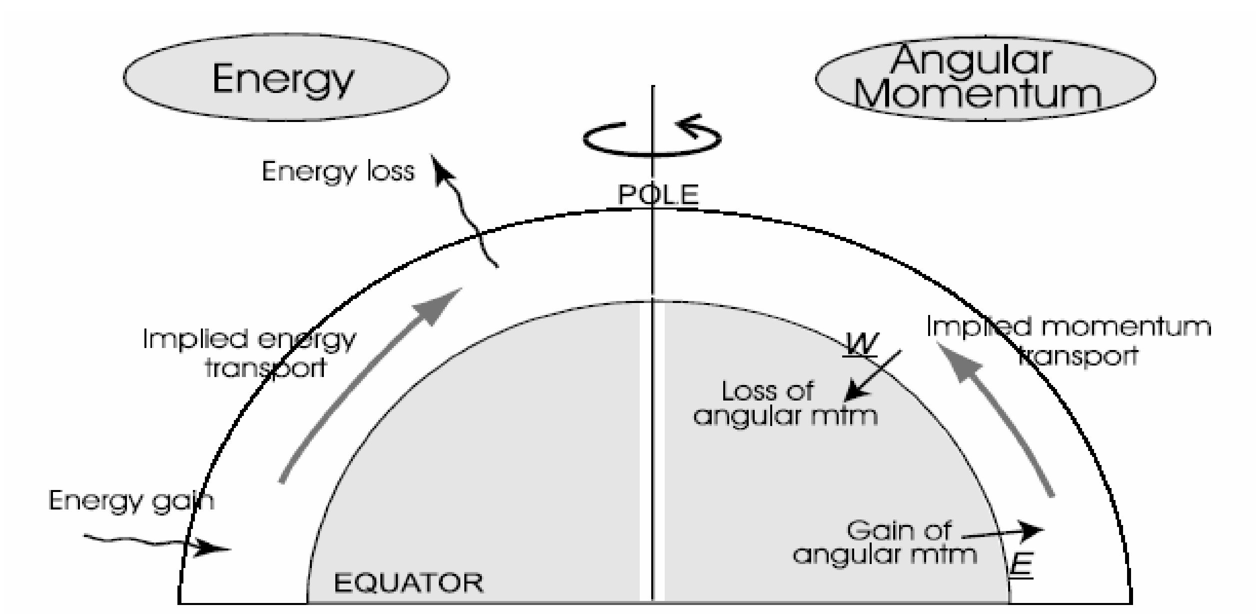
Tropopause to ~50km, temperature increases with height: stratosphere. (bounded above by stratopause).

Stratopause to ~85-90km, temperature decreases: mesosphere.

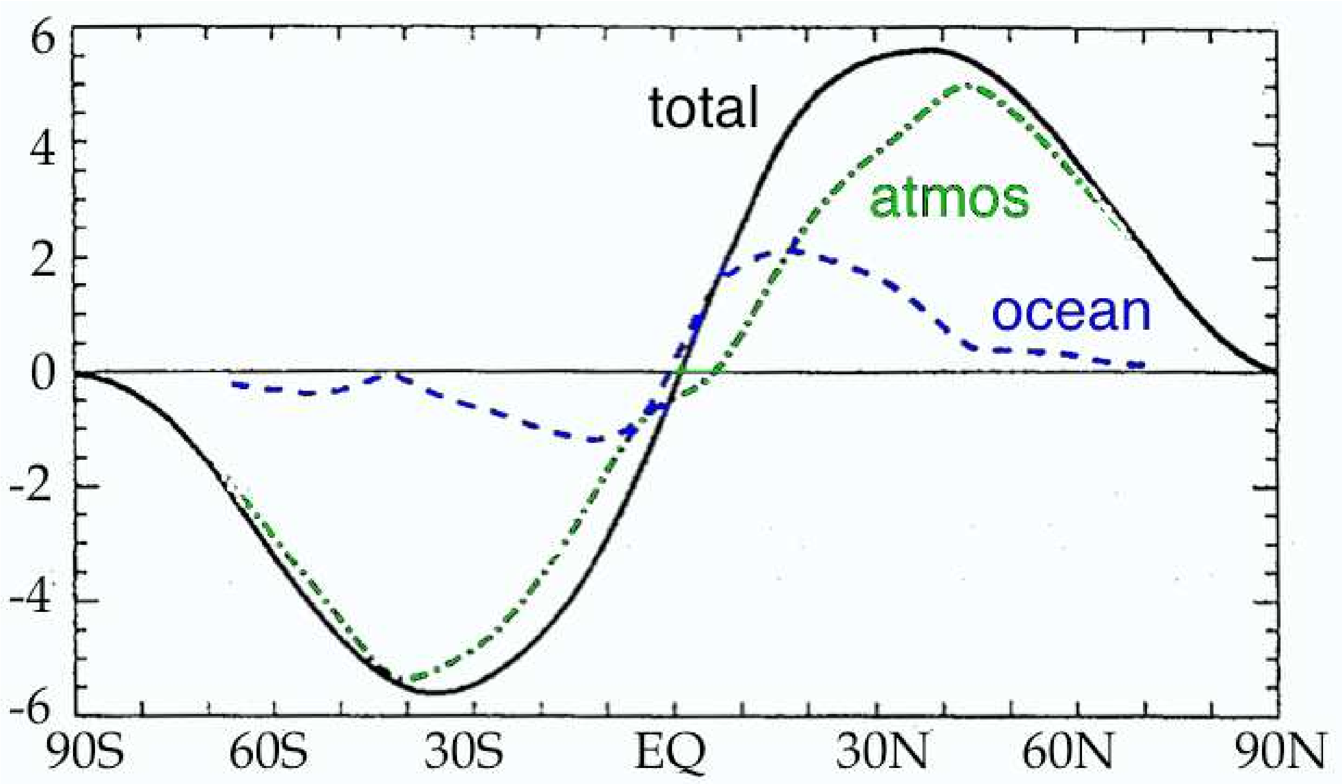


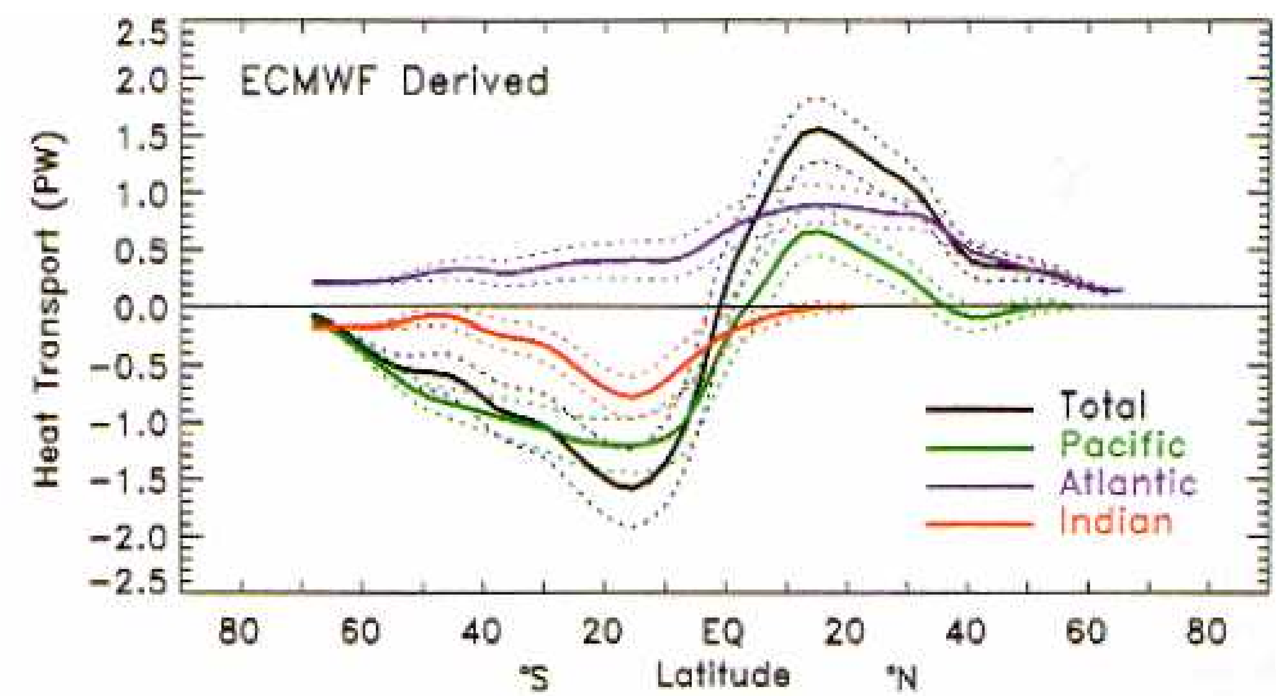
Tropical (*polar*) regions receive more (*less*) energy than emit back to space; implies transport of energy from equator to pole in atmosphere and/or ocean.

Surface flow easterly (*westerly*) in tropics (*midlaitudes*); transport of westerly angular momentum from low to mid-latitudes.



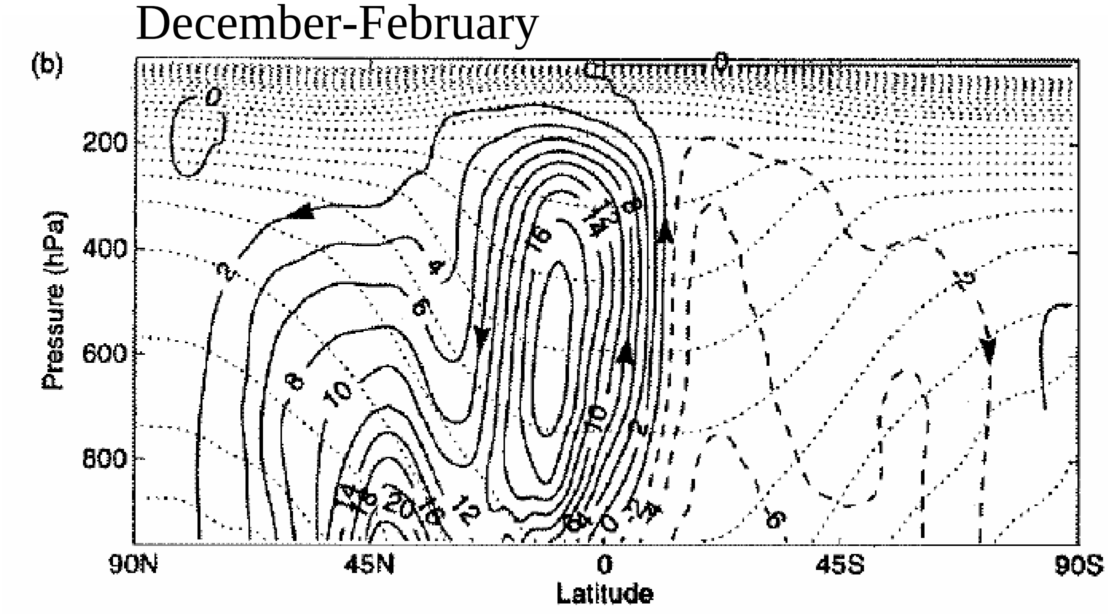
Heat transport in atmosphere and oceans



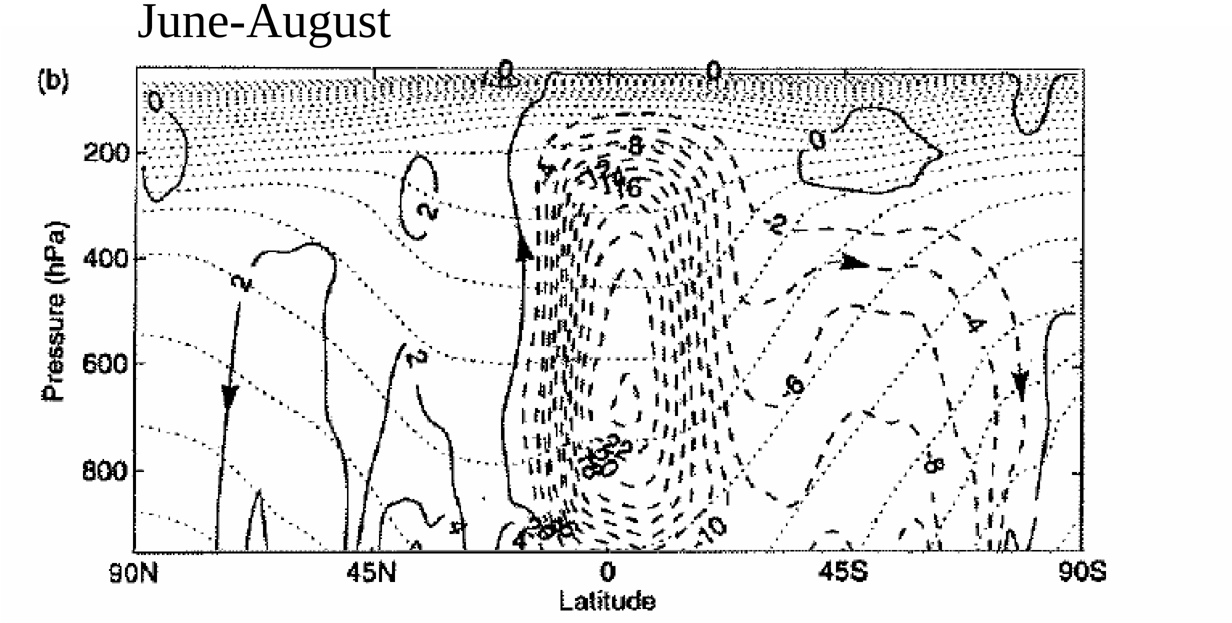


Mean meridional circulation

December-February



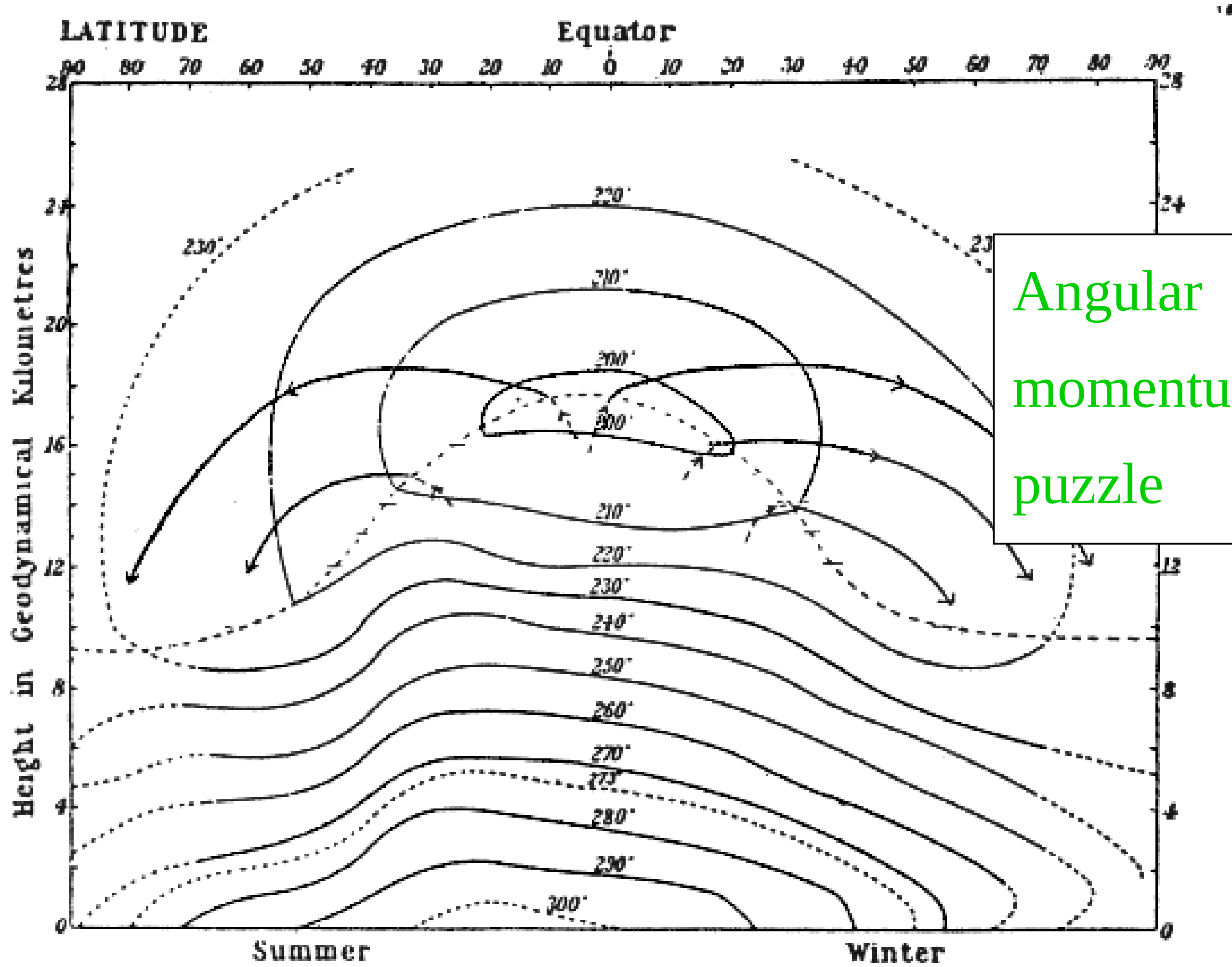
June-August



Residual mean circulation[5]

[5] Average circulation in the height-latitude plane has many definitions – here consider residual mean circulation from D.J. Karoly et al, QJRMS,123, 519-526

Stratospheric mean-meridional circulation



Isotherms over the Globe

FIG. 5. A supply of dry air is maintained by a slow mean circulation from the equatorial tropopause.

Angular momentum puzzle

Brewer (1949) noted water vapour in lower stratosphere less than expected and deduced a 'world circulation' with air entering stratosphere in tropics, moving polewards and descending in extratropics[6]: **Brewer-Dobson circulation.**

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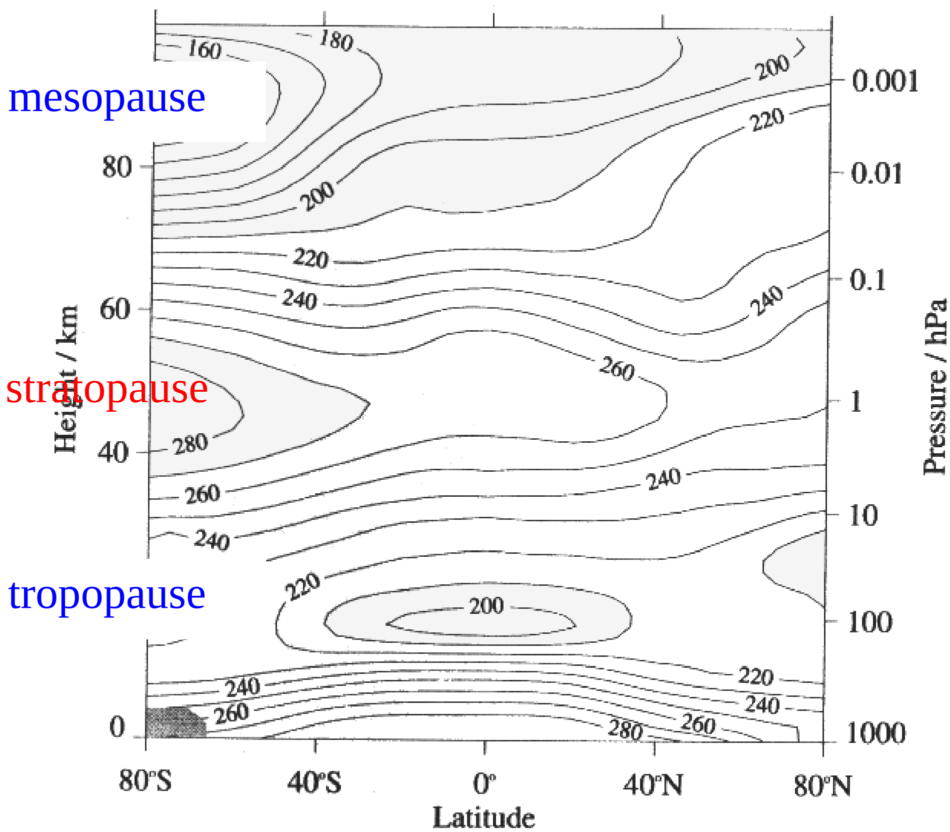
[6] A.W. Brewer, Quart. J. Roy. Meteor. Soc., 75, 351-363, 1949.

Latitudinal temperature variation



Zonal-mean temperatures (K) for January

Equatorial tropopause: higher & colder than extratropical (height: moist convection in equatorial, also cyclones/anti-cyclones at midlatitudes; temperature: dynamically driven rising motion at equator).

Summer stratopause: lower & warmer than winter (absorption of solar radiation by ozone).

Summer mesopause is extremely cold

Coriolis force

For large-scale atmospheric motions (~100 km) Coriolis forces (from rotation of Earth) significant: tend to deflect a moving portion of air to the right (*left*) of its motion in the Northern (*Southern*) Hemisphere.

Geostrophic balance between Coriolis forces and horizontal pressure gradient forces leads to wind motions that circulate along isobars (surfaces of constant pressure) at a given height.

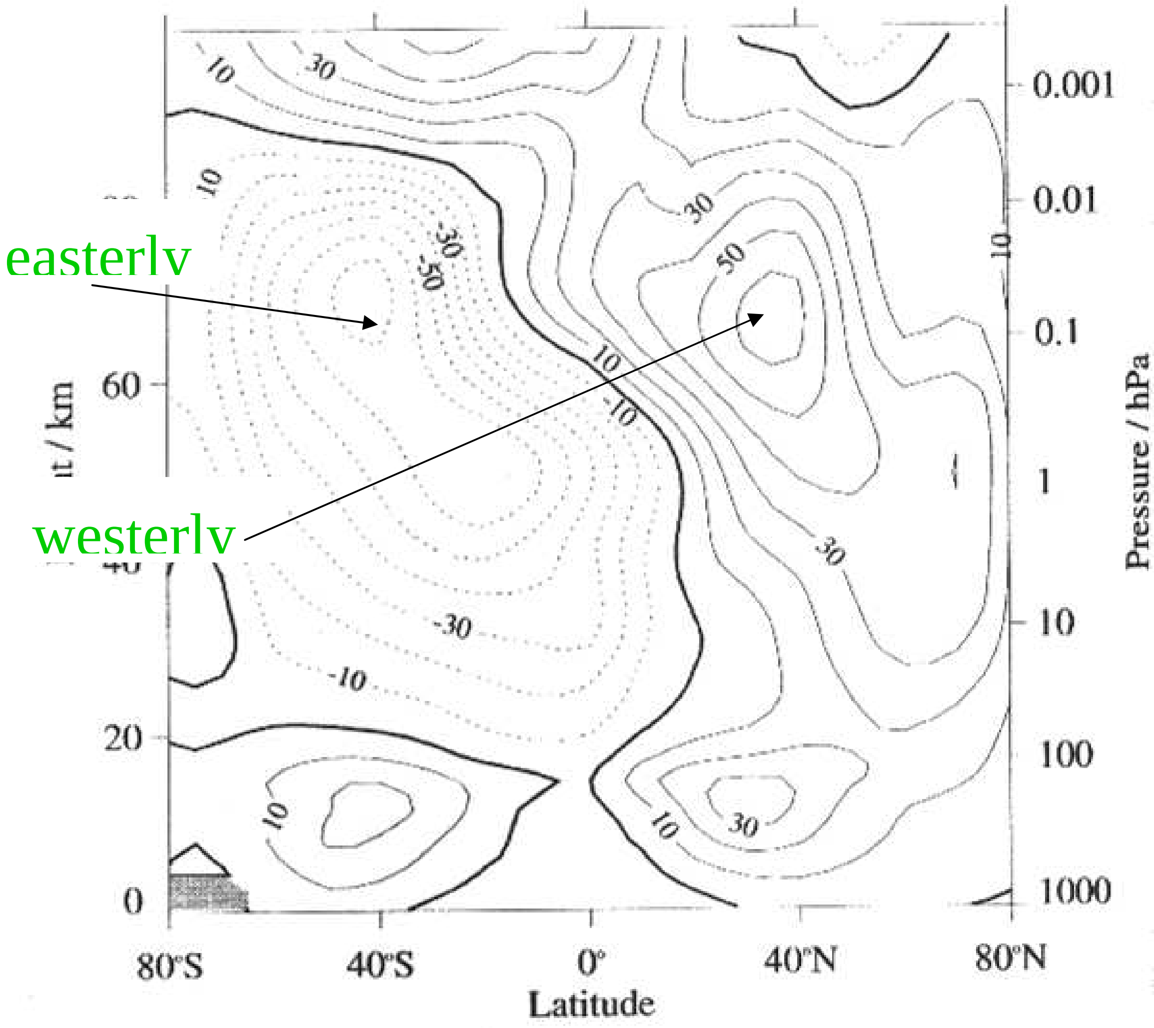
$$fv = \frac{1}{\rho}\frac{\partial p}{\partial x} \qquad -fu = \frac{1}{\rho}\frac{\partial p}{\partial y} \qquad (1.5)$$

Anticlockwise (*clockwise*) rotation around low (*high*) pressure regions in Northern Hemisphere and vice versa in Southern Hemisphere.

With ideal gas and hydrostatic balance gives **thermal wind equations:**

$$f\frac{\partial v}{\partial z} \approx \frac{g}{T}\frac{\partial T}{\partial x} \qquad -f\frac{\partial u}{\partial z} \approx \frac{g}{T}\frac{\partial T}{\partial y} \qquad (1.6)$$

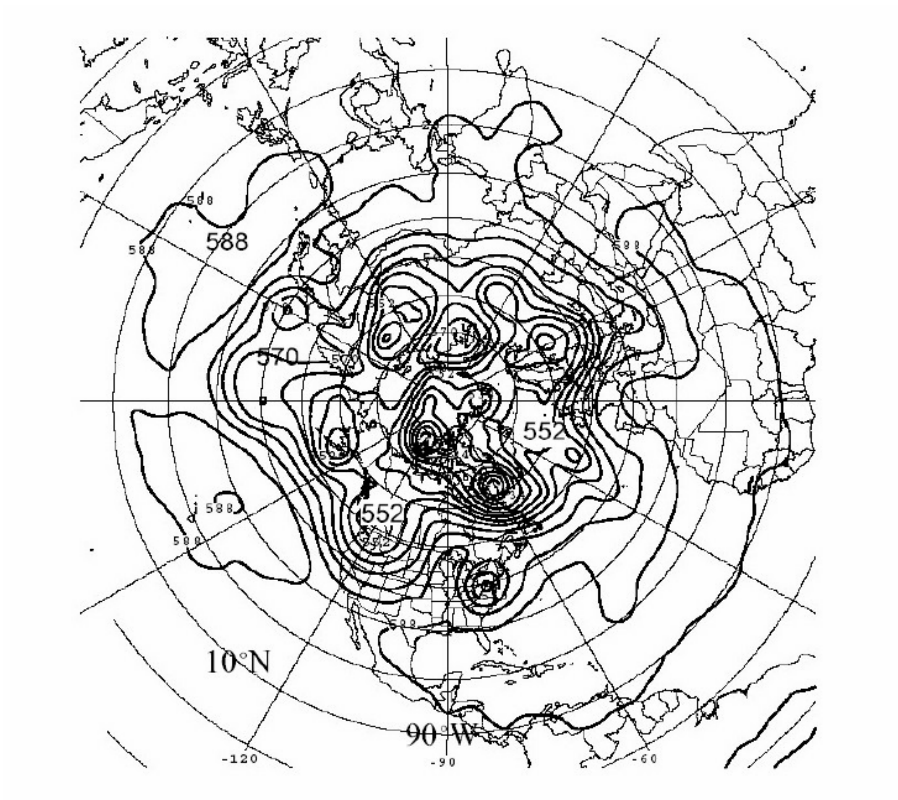
Height-latitude structure of wind field



Zonal-mean zonal wind (m/s) for January

Troposphere: mean zonal winds eastward (westerly) at midlatitudes with two prominent 'jet streams', and westwards (easterly) at low latitudes. Stratosphere & mesosphere: mean zonal winds westerly (*easterly*) in winter (*summer*).

Eddies



Typical 500hPa height analysis: 12 GMT on June 21st, 2003

Eddies (longitudinally varying part of flow) play vital role in transporting heat, moisture, chemical species in latitude/height plane.

Gravity and Rossby waves

Stably stratified atmosphere can support fluid-dynamical **gravity waves** in which the fluid pressure, density, temperature and velocity fluctuate together.

Rossby waves - associated with many large-scale disturbances in troposphere and stratosphere.

Scales of fluid flow in the atmosphere

Microscale: $10^0 - 10^3\,\mathrm{m}$, $10^0 - 10^3\,\mathrm{s}$ (<1 km, <1 hour)
3D turbulence in planetary boundary layer, convective clouds, intermittent patches.

Mesoscale: $10^4 - 10^5\,\mathrm{m}$, $10^2 - 10^4\,\mathrm{s}$ (< 10 km,<1 day)
Gravity waves (e.g. mountains, convection), organised convection (e.g. squall lines, fronts), gravity currents (e.g. sea breeze).

Synoptic: $\sim 10^6\,\mathrm{m}$, $10^5 - 10^6\,\mathrm{s}$ (~1000 km, ~week)
Hurricanes, cyclones, tropical waves (intraseasonal oscillations)

Planetary: $\sim 10^7\,\mathrm{m}$, $> 10^6\,\mathrm{s}$ (~10,000 km ~months)
Rossby waves, monsson circulations, Walker circulation.

Important interaction between scales – small scale processes have systematic effect on large scales.