Energy Balance of the Atmosphere

Atmospheric motions are generated by geographic variations in heating of the surface caused by meridional gradients of insolation, albedo variations, and other factors. By transporting energy, winds generally act to offset the effects of these heating variations on atmospheric temperature. The local energy balance of an atmospheric column of unit horizontal area includes the effects of radiation, sensible heat exchange with the surface, condensation heating, and the horizontal flux of energy in the atmosphere.

$$\frac{\partial E_a}{\partial t} = R_a + \mathrm{LP} + \mathrm{SH} - \Delta F_a \qquad (1)$$

Where $\partial E_a/\partial t$ is the time rate of change of the energy content of an atmospheric column of unit horizontal area extending from the surface to the top of the atmosphere, ***R,*** is the net radiative heating of the atmospheric column, ***LP*** is the heating of the atmospheric column by latent heat release during precipitation, SH is the sensible heat transfer from the surface to the atmosphere, and ***AFa*** is the horizontal divergence of energy out of the column by transport in the atmosphere.

The net radiative heating of the atmosphere is the difference between the net radiative heating at the top of the atmosphere and the net radiation at the ground.

$$R_a = R_{\mathrm{TOA}} - R_s \qquad (2)$$

The storage of energy in the atmosphere is negligible, particularly when averaged over a year, so that the atmospheric energy balance is the sum of radiative heating, sensible heating, and latent heating, balanced against the export of energy by atmospheric motions.

$$R_a + \mathrm{LP} + \mathrm{SH} = \Delta F_a \qquad (3)$$

The annually and zonally averaged net effect of radiative transfer on the atmosphere is a cooling of about -90 W m^{-2}, which is nearly independent of latitude (Fig.1). The radiative cooling corresponds to an atmospheric temperature decrease of about **1.5˚C** per day. The energy lost from the atmosphere in one week through radiative transfer equals about **2.5%** of the atmosphere's total energy content. If only the atmosphere's thermal capacity were considered, this cooling rate, acting alone, would bring the global mean surface air temperature to below freezing in about **2** weeks. Under normal circumstances the radiative cooling is balanced in the global mean by condensation heating and sensible heat transfer from the surface.

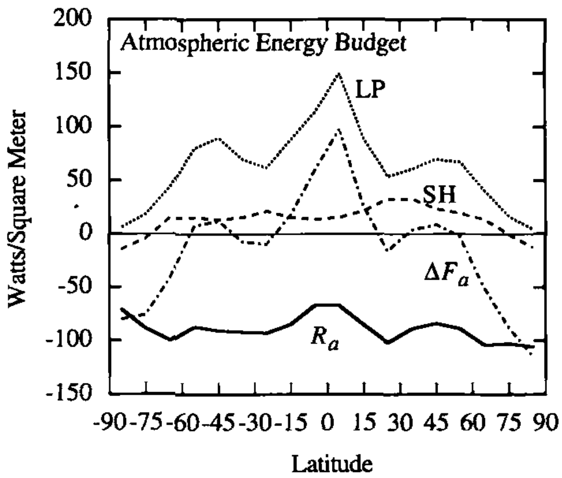

Fig.:1 Distribution with latitude of the components of the atmospheric energy balance averaged over longitude and over the annual cycle. Units are W m-'. [Data from Sellers (1965). Used with permission from the University of Chicago Press.]

Heating of the atmosphere by the transfer of sensible heat from the surface is relatively small. The largest contribution to balancing the radiative loss from the atmosphere is the release of latent heat of vaporization during precipitation. In contrast to the radiative cooling, the condensation heating has a very distinctive structure with latitude corresponding to that of precipitation. It peaks at about 150 W m^{-2} near the equator, drops to near 50 W m^{-2} in the subtropics, peaks again near 80 W m^{-2} in midlatitudes, and then decreases sharply to near zero at the poles. The latitudinal structure of the precipitation is reflected in the latitudinal structure of the atmospheric energy flux divergence. Atmospheric motions export close to 100 W m^{-2} from the equatorial region, have a relatively small net effect on the energy balance between about 20 and 60 degrees of latitude, and import about 100 W m^{-2} into the polar regions. The poleward transport of energy by the atmosphere has a broad, flat maximum between the equatorial and polar regions. This poleward transport of energy by the atmosphere is one of the important climatic effects of the general circulation of the atmosphere.

Atmospheric Motions and the Meridional Transport of Energy

Motions in the atmosphere can be associated with many physical phenomena, which have a wide variety of space and time scales. Small-scale phenomena such as turbulence and organized mesoscale phenomena such as thunderstorms are effective primarily at transporting momentum, moisture, and energy vertically. Only very large-scale phenomena such as extratropical cyclones, planetary-scale waves, and slow meridional circulations that extend over thousands of kilometers are effective at transporting momentum, heat, and moisture horizontally between the tropics and the polar regions. The upward flux of energy and moisture in the boundary layer and the poleward flux of energy by planetary-scale circulations in the atmosphere have equal importance for climate. These phenomena have characteristic spatial scales that differ by nearly 10 orders of magnitude: from millimeters to 10 thousand kilometers.

Wind Components on a Spherical Earth

Wind velocities in the atmosphere are measured in terms of a local Cartesian coordinate system inscribed on a sphere. At each latitude (ϕ) and longitude ***(λ)***on a sphere of radius a, the zonal and meridional components of horizontal velocity are defined in the following way (Fig. 2):

$$u = a\cos\phi \, \frac{D\lambda}{Dt} = \text{zonal or eastward wind speed}$$

$$v = a \, \frac{D\phi}{Dt} = \text{meridional or northward wind speed} \qquad (3)$$

Here ***D/Dt*** represents the material derivative-the temporal tendency that is experienced by an air parcel moving with the flow. The vertical component of velocity can be measured in terms of the rate of change of altitude, or the rate of change of pressure following the motion of air parcels.

$$w = \frac{Dz}{Dt} = \text{rate of change of altitude following an air parcel}$$

$$\omega = \frac{Dp}{Dt} = \text{rate of change of pressure following an air parcel} \qquad (4)$$

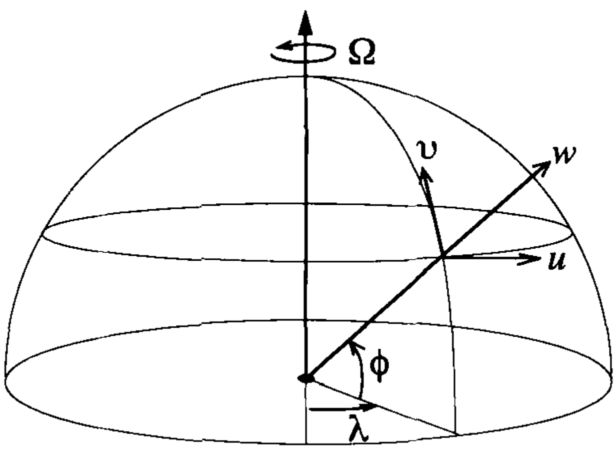

Fig. 2: Diagram showing local Cartesian coordinates on **a sphere and the zonal *(u),* meridional *(u),*** and **vertical *(w)* components of the local vector wind velocity.**

The vertical velocity and the pressure velocity are related to each other through an approximate equation, which is valid if a hydrostatic balance is maintained.

$$\omega \cong -\rho g w \qquad (5)$$

The Zonal-Mean Circulation

In describing the circulations of the atmosphere it is convenient to consider the zonal average, which is the average over longitude, **(λ)**, at a particular latitude and pressure, and is represented with square brackets.\

$$[x] = \frac{1}{2\pi} \int_{0}^{2\pi} x \, d\lambda \qquad (6)$$

Because of the relatively rapid rotation of Earth, and because diurnally averaged insolation is independent of longitude, averaging around a latitude circle captures a physically meaningful subset of the climate. For climatological purposes, we are normally interested in averages over a period of time, ***At,*** that is long enough to average out most weather variations. This time interval may correspond to a particular month, season, or year, or it may be an average over an ensemble of many months, seasons, or years.

$$\overline{x} = \frac{1}{\Delta t} \int_{0}^{\Delta t} x \, dt \qquad (7)$$

Climatological zonal averages are usually obtained by averaging over both longitude and time.

The distribution **of** the zonal mean **of** the eastward component **of** wind, **[u]**, through latitude and height is one **of** the best known characterizations of the global atmospheric circulation, and is often called the zonal-mean wind (Fig. 3). In meteorology, winds are called westerly when they flow from west to east and easterly when they flow from east to west. The zonal-mean wind is westerly through most of the troposphere, and peaks at speeds in excess of 30 m s^{-1} in the subtropical jet stream, which is centered near 30 degrees of latitude and at an altitude of about 12 km. The subtropical jet stream is strongest in the winter season. The zonal winds at the surface are westerly at most latitudes between 30 and 70 degrees, but in the belt between 30''N and 30's zonal-mean easterly surface winds prevail.

The zonal-average meridional and vertical components of wind are much weaker than the zonal wind. Maximum mean meridional winds are only about 1 m s^{-1}, and mean vertical wind speeds are typically a hundred times smaller than the mean meridional wind. The mean meridional circulation (MMC), which is composed of the zonal-mean meridional and vertical velocities, can be described by a mass streamfunction, which is defined by calculating the northward mass flux above a particular pressure level, p.

$$\Psi_M = \frac{2\pi a \cos\phi}{g} \int_0^p [v]\, dp \qquad (8)$$

The mass flow between any two streamlines of the mean meridional streamfunction is equal to the difference in the streamfunction values. The conservation of mass for the zonal-mean flow implies a relationship between the mass streamfunction for the mean meridional circulation and the mean meridional velocity and pressure velocity

$$[v] = \frac{g}{2\pi a \cos\phi} \frac{\partial \Psi_M}{\partial p} \qquad (9)$$

$$[\omega] = \frac{-g}{2\pi a^2 \cos\phi} \frac{\partial \Psi_M}{\partial \phi} \qquad (10)$$

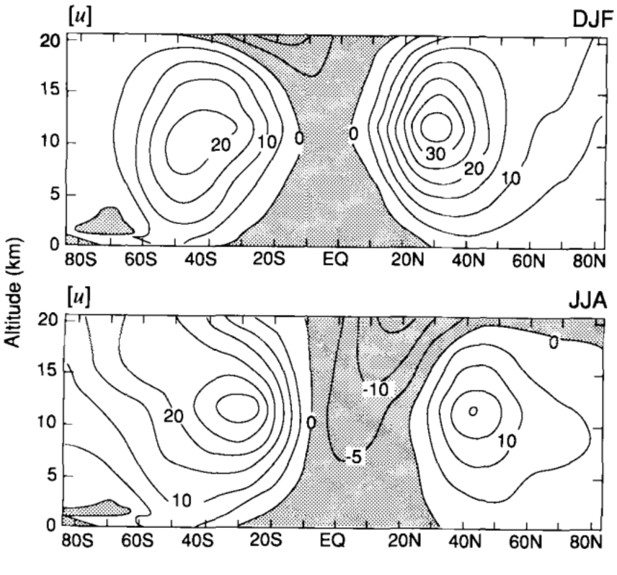

Fig. 3: Latitude-height cross section of zonal-average wind speed for DJF and JJA. Contour interval is ***5* m s-l;** easterly values **are** shaded. [Data from Oort **(1983).]**

The mean meridional velocity thus depends on the rate at which the streamfunction changes with pressure, and the zonal-average pressure velocity depends **on** the rate at which the streamfunction changes with latitude.

The mean meridional circulation is dominated in the solstitial seasons by a single circulation cell in which air rises near the equator, flows toward the winter hemisphere at upper levels, and sinks in the subtropical latitudes of the winter hemisphere (Fig. 6.5). This mean meridional circulation cell is often called the ***Hadley cell*** after George Hadley, who in 1735 proposed it as an explanation for the tradewinds. The mean meridional winds near the surface bring air back toward the equator. The rising branch is displaced slightly into the summer hemisphere. The mean meridional circulations for the equinoctal seasons and for the annual average consist of two smaller cells of about equal strength located on opposite sides of the equator.

In midlatitudes weaker cells called ***Ferrel cells*** circulate in the opposite direction to the Hadley cell. In these midlatitude mean meridional circulation cells, rising occurs in cold air and sinking in warmer air. These cells are therefore thermodynamically indirect, in that they transport energy from a cold area to a warm area. The mean meridional circulation is a small component of the total flow in midlatitudes, and the Ferrel cells are a byproduct of the very strong poleward transport of energy by eddy circulations. Eddies are the deviations from the time or zonal average, and are a key component of the general circulation of the atmosphere.