Zonal Mean Climatology

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Introduction

The middle atmosphere is the region of the atmosphere between approximately 10 and 100 km: above the troposphere (dominated by convection and synoptic-scale weather systems) and below the thermosphere (where ionization and molecular diffusion become important physical processes). This article discusses aspects of the planetary-scale 'climatological' temperature and circulation features of the middle atmosphere, i.e., those features that vary slowly from month to month during the seasonal cycle and occur regularly each year. The predominant climatological features are most simply understood in terms of longitudinal (or zonal) averages, and much of the analysis is presented in terms of latitude–height diagrams of zonal average statistics. While transient large-scale departures from zonal symmetry ('planetary waves') are an important aspect of the winter stratosphere, these are discussed mainly in terms of their calculated net influence on the zonal mean flow. The focus here is on structure and seasonality in the middle atmosphere, based on long records (>20 years) of observations, and on understanding the large-scale balances between temperatures, winds, and the global-scale forcing mechanisms. Important inferences about mechanisms controlling middle atmosphere circulation can be gained from recognizing the large differences in climatological structure between the Northern and Southern hemispheres, and these inter-hemispheric differences are highlighted here. The tropical middle atmosphere exhibits behavior distinctive from that at midlatitudes, and a brief overview of tropical climatology is discussed separately.

Atmospheric Vertical Structure

The Earth's atmosphere is conventionally divided into layers on the basis of the vertical structure of the temperature field (**Figure 1**). The lowest layer of the atmosphere, termed the troposphere, exhibits decreasing temperature with height up to a minimum at the tropopause (near 8–16 km, depending on latitude). The middle atmosphere is the region between approximately 10 and 100 km, consisting of the stratosphere (where temperature increases with height, mainly owing to ozone absorption of solar ultraviolet radiation), and the mesosphere (where temperature decreases with height). The temperature maximum near 50 km is termed the stratopause, and the minimum near 85 km is the mesopause. The temperature increases rapidly with height above 100 km in the thermosphere; in contrast to lower altitudes, molecular diffusion, ionization, and ion-drag are important in the thermosphere, so important physical processes are fundamentally different above ~100 km. The density of the atmosphere decreases exponentially with height (with an approximate 7 km e-folding scale), so that the

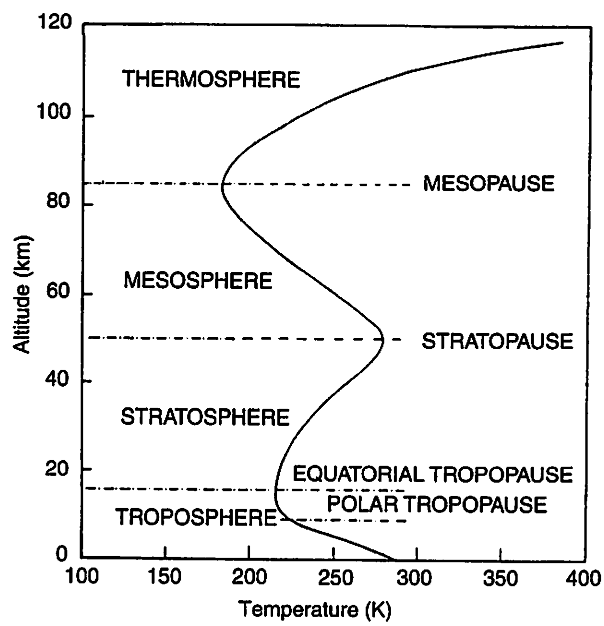

Figure 1 Vertical structure of atmospheric temperature and distinct layers.

middle atmosphere contains only ~10% of the total mass of the atmosphere (with only 0.1% above the stratopause).

Observations

Observations of the middle atmosphere are derived from balloon or rocket measurements, ground-based remote sensing (lidar or radar techniques), and satellite remote sensing. Routine balloon observations of the lower-to-middle stratosphere have been made since the late 1950s, with data mostly over land areas of the Northern Hemisphere. Global temperature measurements of the stratosphere have been available since ~1979 from a series of operational meteorological satellites. Shorter records of middle atmosphere wind and constituent measurements are available from research satellites, particularly data from the NASA Upper Atmosphere Research Satellite (UARS) (1991–2002). Global meteorological analyses now routinely extend from the surface into the middle stratosphere, optimally combining balloon and satellite measurements with numerical forecasts to produce daily analyses of winds and temperatures at stratospheric levels. Daily observations at mesospheric levels (> 50 km) are not routinely available at present. The climatological data shown here are from a variety of sources, but for the most part span the time frame of the 1990s (UARS period).

Zonally Averaged Equations of Motion

A simplified perspective of the middle atmosphere general circulation is provided by examining the seasonal variation of the longitudinally (or zonally) averaged flow and the balance of terms in the corresponding governing equations of motion. It is straightforward to separate the governing equations (for momentum, thermodynamic energy, and mass conservation) into zonal mean and eddy components, and a useful form of the equations is provided by the quasi-geostrophic transformed Eulerian-mean eqns [1]–[3].

$$\frac{\partial \bar{u}}{\partial t} - 2\Omega \sin \phi \overline{v^*} = \bar{G} \qquad [1]$$

$$\frac{\partial \bar{T}}{\partial t} + \overline{w^*}\left(\frac{HN^2}{R}\right) = \bar{Q} \qquad [2]$$

$$\frac{1}{a \cos \phi} \frac{\partial}{\partial \phi}(\overline{v^*} \cos \phi) + \frac{1}{\rho} \frac{\partial}{\partial z}(\rho \overline{w^*}) = 0 \qquad [3]$$

Here $\bar{u}$ and $\bar{T}$ are the zonal mean wind and temperature fields, and $(\overline{v^*}, \overline{w^*})$ are components of the mean circulation in the latitude–height plane. The latitude–height coordinates are denoted by (ϕ, z), ρ is the background atmospheric density (proportional to $e^{-z/H}$), H is an atmospheric scale height (7 km), Ω is the Earth's angular velocity (7.3×10^{-5} rad s^{-1}), a is the radius of the Earth (6.37×10^6 m), R the gas constant for dry air (287 J kg^{-1} K^{-1}), and N is the buoyancy frequency. Changes in the zonal mean flow are driven by eddy forcing ($\bar{G}$) of the zonal (or angular) momentum, together with radiative forcing ($\bar{Q}$) of the thermodynamic field. The zonal mean momentum and thermodynamic fields are coupled by geostrophic thermal wind balance (eqn [4]).

$$2\Omega \sin \phi \frac{\partial \bar{u}}{\partial z} + \frac{R}{H} \frac{1}{a} \frac{\partial \bar{T}}{\partial \phi} = 0 \qquad [4]$$

The meridional circulation $(\overline{v^*}, \overline{w^*})$ is driven by the $\bar{G}$ and $\bar{Q}$ forcing terms, and acts to redistribute their effects nonlocally in a manner that maintains thermal wind balance.

The wave forcing term ($\bar{G}$) is often partitioned into components associated with large-scale waves (which are resolved in meteorological analyses), and components attributable to smaller or unresolved scales (in particular, gravity waves). While the forcing from large-scale waves can be estimated directly from stratospheric analyses, the global climatology of gravity wave forcing is poorly quantified at present. The amplitude of gravity waves grows exponentially with height (owing to decreasing atmospheric density), and nonlinear effects at high altitudes (due to convective or dynamic instabilities) cause a substantial forcing to the background flow. Numerical models suggest that gravity wave forcing is a dominant effect in the mesosphere, and also probably important in the lower stratosphere.

Observed Structure and Seasonality

Wind and Temperature

The seasonal variation of zonal mean temperature and wind fields are shown in **Figures 2** and **3**. The stratospheric circulation is characterized by a cold pole and strong westerly (from west to east) winds in the winter hemisphere. This cyclonic circulation is referred to as the stratospheric polar vortex, and the wind maximum is known as the polar night jet. The polar temperatures are substantially colder and the jet is more intense during Southern Hemisphere winter. The autumn and winter westerly jets extend throughout the mesosphere, exhibiting an equatorward tilt with height. The seasonal sequence shows that the polar night jet forms initially in the upper stratosphere and mesosphere, and descends with time.

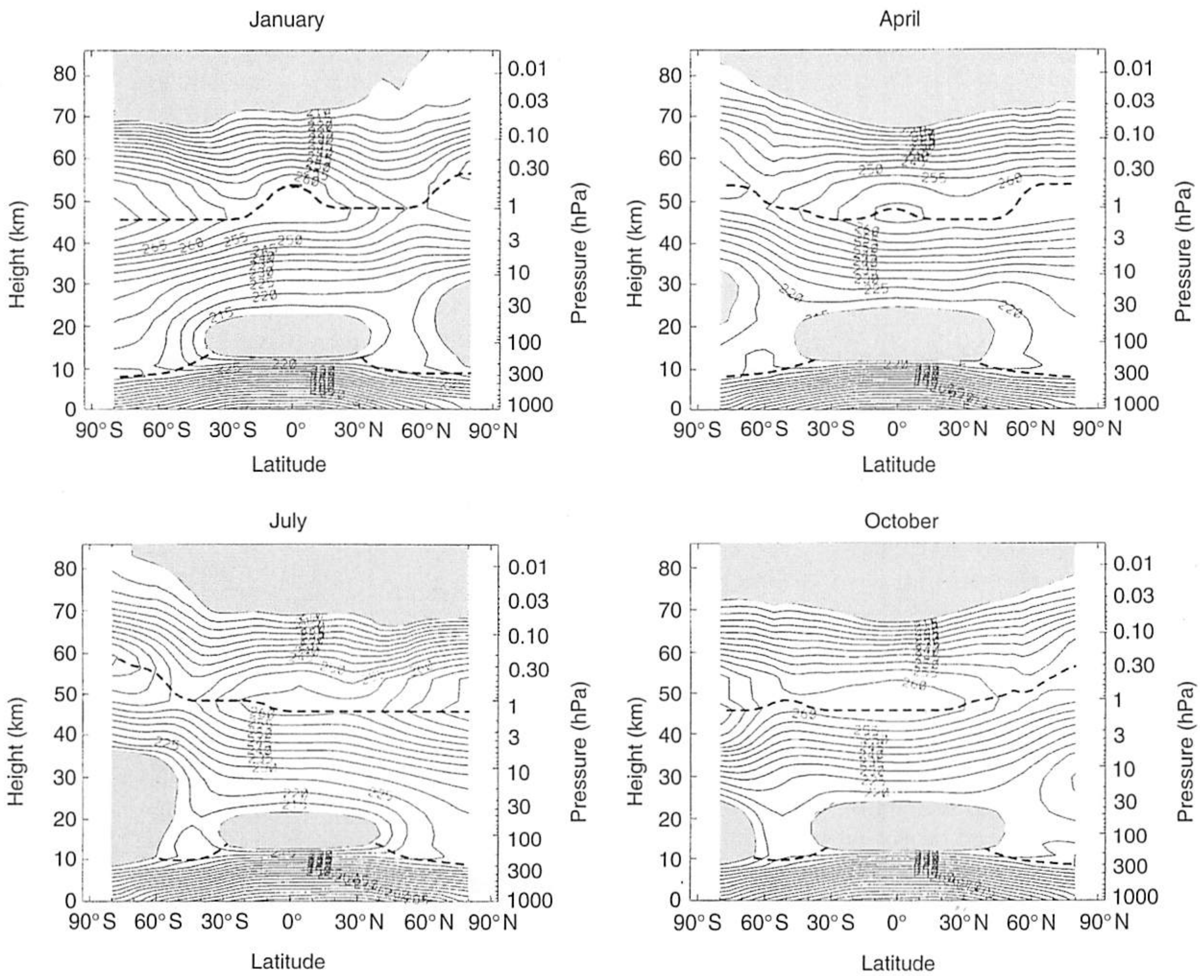

Figure 2 Latitude–height cross-sections of zonal mean temperature (K) in January, April, July, and October. Upper and lower dashed lines denote the local stratopause and tropopause for each month. Temperatures below 210 K are shaded.

The summer hemisphere circulation is characterized by a reversed equator-to-pole temperature gradient and easterly winds. These easterlies span the entire stratosphere and mesosphere, with the jet maximum tilting poleward with increasing altitude. There is an overall symmetry between the strength of the Northern Hemisphere and Southern Hemisphere summer easterlies, except near the tropical stratopause where the January jet is stronger than that in July (this is related to the tropical semiannual oscillation, discussed below).

Heating Rates and Meridional Circulation

Radiative heating and cooling variations provide a first-order driver of the stratospheric circulation and its seasonal variation. Heating of the middle atmosphere is due mainly to ozone absorption of solar ultraviolet radiation, while infrared cooling is due to emission by carbon dioxide, ozone, and water vapor molecules. The latitudinal distribution of the solar heating is strongly dependent on season (maximum in the summer hemisphere), whereas the infrared cooling is mostly dependent on temperature and has less latitudinal structure. The net radiative forcing (heating plus cooling: the $\bar{Q}$ in eqn [2]) is thus a strong function of latitude during solstice seasons (**Figure 4**), with heating in the summer and cooling in the winter hemisphere.

The mean meridional circulation $(\overline{v^*}, \overline{w^*})$ can be derived from the net radiative heating rates using the thermodynamic balance (eqn [2]) coupled with mass continuity (eqn [3]). Climatological estimates of the circulation for January and July are shown in **Figure 5**. Mean upward motion is observed in tropical latitudes throughout the year, with strong flow toward the winter hemisphere and sinking over the winter polar regions. This overturning circulation is referred to as the Brewer–Dobson circulation, and the strong annual cycle seen in **Figure 5** plays a primary role in mass and constituent transport in the middle atmosphere. In addition to the large seasonal cycle, the

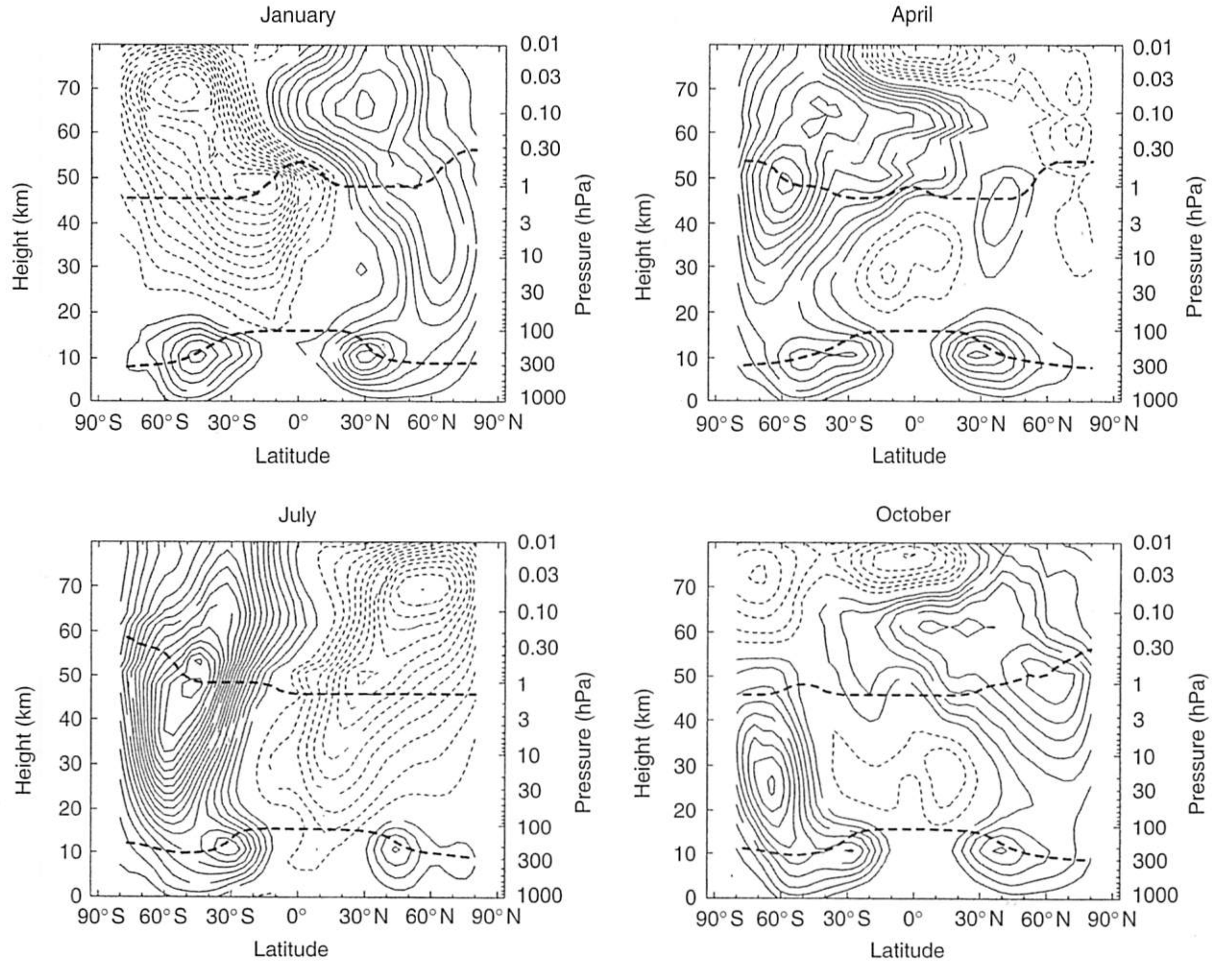

Figure 3 Latitude–height cross-sections of zonal mean zonal wind ($m\,s^{-1}$) for January, April, July, and October. Contour interval is $5\,m\,s^{-1}$, with zero contours omitted. Solid lines are westerly winds, and dashed lines are easterly winds. Dark dashed lines denote the tropopause and stratopause for each month, as in **Figure 2**.

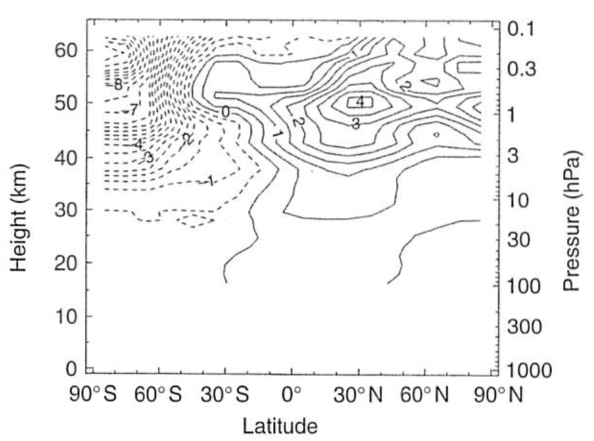

Figure 4 Latitude–height cross-section of climatological radiative heating ($K\,d^{-1}$) in July. Positive values denote heating, and negative (dashed) contours show cooling.

Brewer–Dobson circulation is somewhat stronger during Northern Hemisphere winter; note the stronger downward polar flow in January compared to July in **Figure 5**. This asymmetry is due to interhemispheric differences in the strength of tropospheric planetary wave forcing in midlatitudes.

Large-scale Wave Forcing

The rectified effect of large-scale waves that propagate from the lower atmosphere is a primary mechanism for forcing the middle atmosphere circulation, and observed wave characteristics explain many of the Northern Hemisphere–Southern Hemisphere asymmetries discussed above. Note that while the global radiative heating acts mainly as a restorative force (toward radiative equilibrium), wave driving acts in the opposite direction and tends to drive the middle atmosphere away from radiative equilibrium.

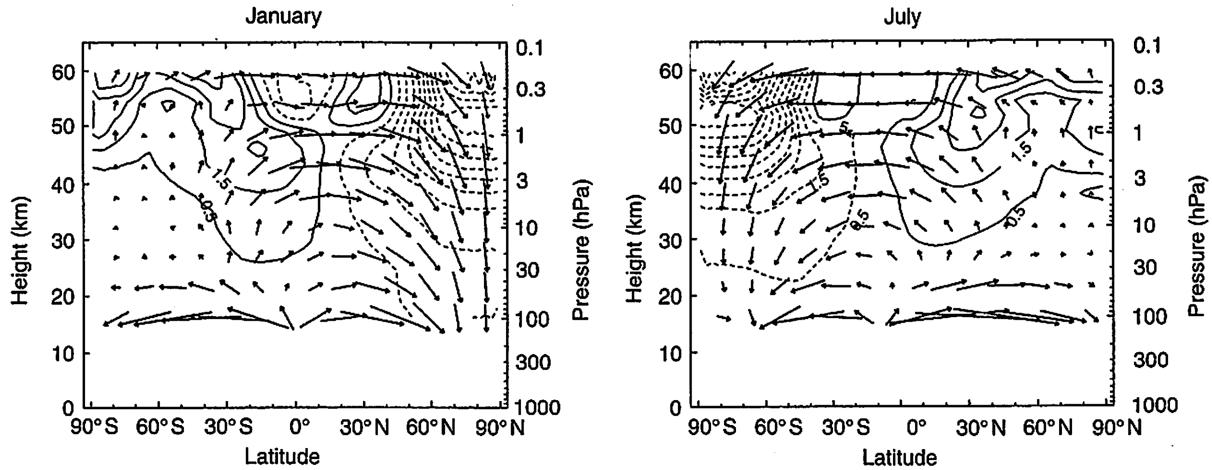

Figure 5 Cross-sections of the mean meridional circulation in January and July. Vectors show components and direction of the local ($\overline{v^*}$, $\overline{w^*}$) circulation, and contours show the magnitude of $\overline{w^*}$, with contours of $\pm$0.5, 1.5, 2.5, ... (mm s^{-1}).

Climatological estimates of wave forcing in the troposphere and stratosphere for January and July are shown in **Figure 6**. These are derived from daily meteorological analyses that span 0 to 50 km and involve covariance estimates of eddy heat and momentum fluxes. The vectors show the magnitude and direction of planetary wave propagation in the latitude–height plane, and contours are derived estimates of wave forcing ($\bar{G}$ in eqn [1]). Strong tropospheric wave activity is observed in mid-latitudes throughout the year, with more intense maxima in the winter hemisphere. The wave statistics in **Figure 6** show upward and equatorward propagation of wave activity in the troposphere; this is a statistical signature of traveling baroclinic waves (synoptic-scale weather systems) that populate the troposphere. While the majority of lower-atmosphere wave activity is confined below the tropopause, some fraction is observed to propagate vertically into the stratosphere, but only in the winter hemisphere. This seasonal variation is due to the fact that vertical propagation of planetary (Rossby) waves occurs only for stratospheric westerly winds; the summer easterlies effectively shield the stratosphere from tropospheric forcing during summer. This effect produces a large annual cycle in wave forcing of the winter stratosphere. A further important feature of the climatology is that the Northern Hemisphere winter stratospheric wave activity is substantially larger than that in the Southern Hemisphere; this is primarily due to stronger orographic and thermal stationary planetary wave forcing in the Northern Hemisphere troposphere (stationary waves are small in the Southern Hemisphere troposphere). A consequence of the larger wave forcing in the

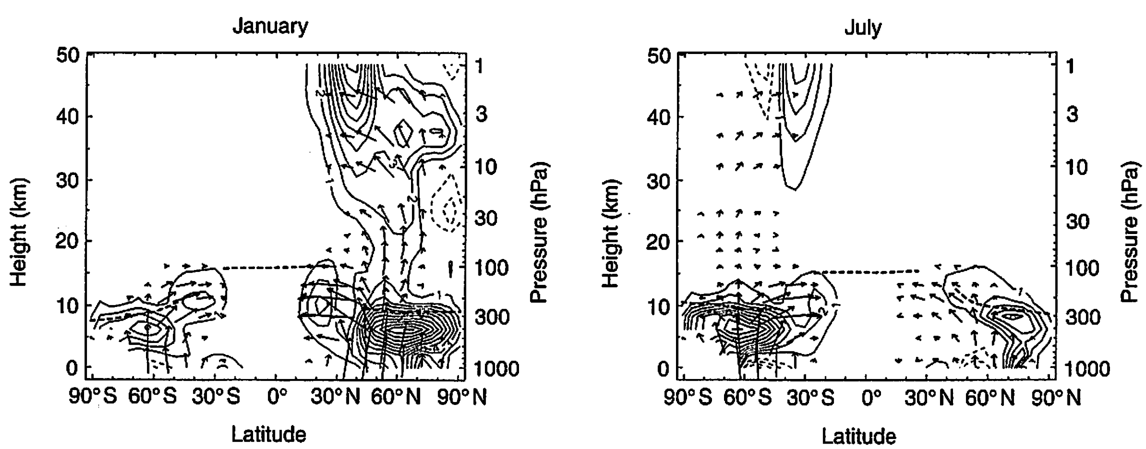

Figure 6 Climatological large-scale wave activity cross-sections for January and July. The vectors show the direction and magnitude of planetary wave propagation, and contours show the net forcing of the zonal mean flow ($\overline{G^*}$ in eqn [1]). Contours are $\pm$1, 2, 3,... m s^{-1} per day. In this figure solid contours denote negative (easterly) forcing. The heavy dashed line denotes the tropopause.

Northern Hemisphere is that the Northern Hemisphere stratospheric polar vortex is warmer and weaker (and further from radiative equilibrium) than that in the Southern Hemisphere; the observed Northern Hemisphere–Southern Hemisphere vortex asymmetry is a direct result of this difference in wave forcing. The stronger Northern Hemisphere wave driving also forces a more intense Brewer–Dobson circulation during Northern Hemisphere winter (see **Figure 5**), and gives rise to a seasonal cycle in tropical upwelling (stronger during Northern Hemisphere winter).

Constituents and Transport

Important information on the transport and overall flow of mass in the middle atmosphere can be obtained by analysis of long-lived chemical constituents. One such constituent is methane, which is produced by biotic activity in the troposphere, transported into the stratosphere by the mean upward circulation in the tropics and chemically destroyed above 35 km. The photochemical lifetime in the upper stratosphere is on the order of months, so that the overall distribution is determined mainly by the circulation. January and July distributions of stratospheric methane derived from satellite measurements are shown in **Figure 7**. The superimposed vectors represent the corresponding mean meridional circulation for each month, taken from **Figure 5**. High values of methane are observed in the tropical lower stratosphere, resulting from the upward transport of air recently in the troposphere (with methane volume mixing ratios of ~1.7 ppm). Relatively high mixing ratios are observed in the tropics throughout the stratosphere, consistent with the derived upward tropical circulation (and demonstrating a degree of dynamical isolation of the tropics from mid-latitudes). Note that the tropical mixing ratio maximum is displaced into the summer subtropics in the upper stratosphere, consistent with the seasonally varying upward circulation. Lower methane mixing ratios are observed in the extratropics, resulting from the downward transport of 'older' photochemically aged air. This is particularly evident in polar winter latitudes, where the polar vortex inhibits meridional transports and effectively isolates polar regions (particularly for the more intense Southern Hemisphere vortex). The overall consistency of constituent and derived circulation fields promotes confidence in details of the circulation statistics.

Tropical Climatology

The observed global annual cycle in temperature and mean zonal winds (**Figures 2** and **3**) is roughly antisymmetric with respect to the Equator, and the annual harmonic has relatively small amplitude in the tropics. Instead, in the equatorial middle atmosphere above 35 km the seasonal variation is characterized mainly by a semiannual oscillation (SAO) of the zonal mean wind and temperature fields. Below 35 km, the seasonal variation is small, and variability is dominated by a long-term oscillation that is not directly tied to the seasonal cycle. This oscillation has an irregular period slightly longer than 2 years (averaging ~28 months), and is called the quasi-biennial oscillation (QBO). **Figure 8** shows the tropical zonal wind variability during 1992–96, when direct measurements of mesospheric winds were available from the UARS satellite. Although both the SAO and QBO are discussed in separate Encyclopedia articles, we include here a brief summary of observed characteristics.

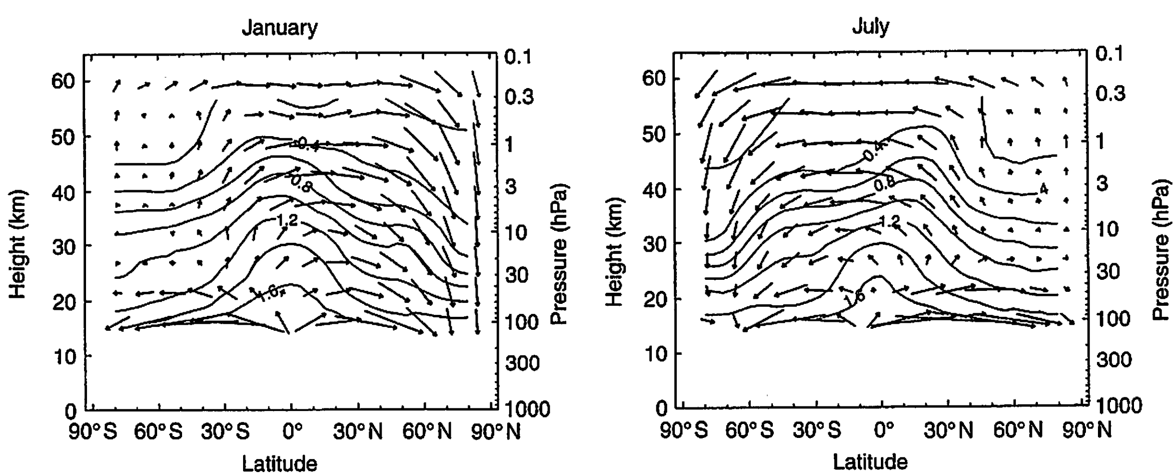

Figure 7 Contours show the mixing ratio of methane in the stratosphere (volume ppm), for climatological January and July observations. Arrows denote the mean Brewer–Dobson circulation for each month, taken from **Figure 5**.

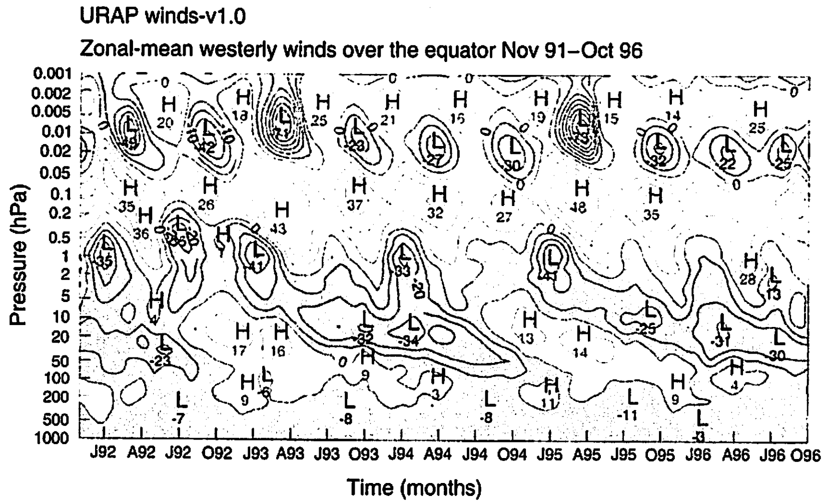

Figure 8 Height–time section of zonal wind variability at the equator during November 1991–October 1996. Contour interval is 10 m s^{-1}, with negative (easterly) winds shaded.

The Semiannual Oscillation (SAO)

The SAO in equatorial zonal wind is evident in **Figure** 8 as a regular, twice-yearly reversal in the wind direction near the stratopause (~50 km). There is also an SAO near the mesopause (~80 km), which is out of phase with the stratopause SAO. A relative minimum in amplitude occurs near ~65 km, but the SAO phase can be traced downward continuously over ~80–40 km. Harmonic analysis shows the SAO to be an equatorially centered phenomena with latitudinal extend of $\pm 30°$, but there is a strong coupling to the extratropical annual cycle that obscures this tropical symmetry for monthly 'snapshots' (see **Figure** 3). One feature of the stratopause SAO is that the amplitude during the first half of the calendar year is typically larger than that during the second half. This asymmetry is linked to the stronger extratropical planetary wave forcing and Brewer–Dobson circulation during Northern Hemisphere winter discussed above.

An intriguing aspect of the SAO (and QBO) winds is the existence of westerly winds at the Equator, which have greater angular momentum than that of the rotating Earth. This 'superrotation' cannot be explained by direct thermal forcing or symmetric circulations, but must arise from the rectified effects of wave forcing. Extensive numerical modeling of the SAO suggests it is forced by momentum transport associated with vertically propagating waves generated in the tropical lower atmosphere; wave interactions with the mean flow and coupling with the strong extratropical annual cycle lead to the seasonally phase-locked behavior of the SAO.

The Quasi-Biennial Oscillation (QBO)

The QBO dominates the variability of the equatorial stratosphere over altitudes of ~16–40 km, and is evident in **Figure** 8 as alternating easterly and westerly wind regimes of magnitude -30 to $+15$ m s^{-1}, with an approximate 2-year period. The period of the QBO in the available observational record (since about 1953) varies between ~2–3 years, with an average period of ~28 months. Downward propagation of successive wind regimes occurs at a rate of ~1 km per month; the westerly (positive) winds descend more rapidly and more regularly than the easterly winds (the easterly phase can often 'stall' for several months, as evident during late 1993 in **Figure** 8). The QBO is weakly linked to the seasonal cycle in that the onset of QBO westerlies in the upper stratosphere is tied to descending westerly phases of the stratopause SAO, and also transitions in the lower stratosphere tend to occur primarily during May–July. The QBO is equatorially centered, but has a much narrower meridional scale (about $\pm 15°$) than the SAO. The QBO is also the dominant signal in temperature in the tropical lower stratosphere, and modulates the strength of the Brewer–Dobson circulation and trace constituents in the tropical stratosphere. Although the dynamical QBO is primarily a tropical phenomenon, it affects the middle atmosphere circulation on a global scale, with substantial QBO signals observed in mid-latitudes and polar regions of both hemispheres. This is due to the modulation of extratropical planetary wave propagation by the alternating regimes of tropical easterly and westerly winds. Changes in wave propagation then affect the pattern of wave dissipation, and hence wave forcing of the global Brewer–Dobson circulation.

Like the SAO, the theoretical understanding of the QBO is that it is forced by momentum transfer by vertically propagating waves forced in the lower atmosphere, interacting with the mean flow. The dynamical theory of the QBO is explained in detail elsewhere.

See also

Middle Atmosphere: Planetary Waves; Polar Vortex; Quasi-Biennial Oscillation; Semiannual Oscillation. **Quasi-geostrophic Theory. Wave Mean-Flow Interaction.**

Further Reading

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Brasseur G and Solomon S (1986) *Aeronomy of the Middle Atmosphere*. Boston, MA: Reidel.

Labitzke KG and van Loon H (1999) *The Stratosphere: Phenomena, History and Relevance*. Berlin: Springer-Verlag.