Large-Scale Atmospheric Dynamics for Atmospheric Chemists

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1. Introduction

1.1. Motivation and Key Concepts

Atmospheric dynamics refers to the motion and thermodynamic state (temperature, pressure, density) of the atmosphere. Atmospheric composition is affected by atmospheric dynamics in several ways. First, atmospheric temperature affects many chemical reaction rates, as well as the phase of certain chemical species—notably but not exclusively H_2O. Second, atmospheric winds move molecules from place to place. To understand the observed distribution of chemical species in the atmosphere and possible changes in this distribution, it is therefore necessary to understand not only atmospheric chemistry, but also those aspects of atmospheric dynamics pertaining to temperature and to chemical transport. [Many chemical reaction rates are also affected by atmospheric pressure, but because of the very close relationship between pressure and altitude that follows from the *hydrostatic relation* (technical terms in italics are defined in the Glossary in section 12), one can regard pressure as a vertical coordinate. In that case, pressure is given rather than being something one needs to determine.]

Chemical production/loss and transport are closely coupled. The evolution equation for the mixing ratio χ of a chemical species can be written

$$\frac{\partial \chi}{\partial t} = - \mathbf{v}\cdot\nabla\chi + S \tag{1}$$

where t is time, $\mathbf{v}$ is velocity, ∇ is the three-dimensional spatial gradient operator, and S represents the net (rate of) chemical change, i.e., production minus loss. Strictly speaking, eq 1 should include molecular diffusion, but molecular diffusion plays a negligible role in atmospheric transport below about 100 km altitude compared to the bulk motion (including turbulence) which is represented by $\mathbf{v}$. (For this reason, below 100 km altitude the gravitational sedimentation of gases is unimportant and the atmosphere is well-mixed.) The term $\partial\chi/\partial t$ refers to the time rate of change of χ at a fixed location; this term can be nonzero because of transport, $-\mathbf{v}\cdot\nabla\chi$, or because of chemistry, S. Note that $-\mathbf{v}\cdot\nabla\chi > 0$ if the winds blow from higher to lower values of χ, in which case the winds replace lower by higher values of χ at a given location and act as a local source of χ. If $\nabla\chi = 0$, then molecules of χ are moved by the winds, but this is of no consequence since there are no spatial

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contrasts in χ. In a steady-state balance (such as a climatological average), the time derivative in eq 1 vanishes and there is a balance between transport and net chemical change. In fact, in some sense one can say that net chemical change is controlled by transport; without transport (i.e., if $\mathbf{v} = 0$), production and loss would have to balance for a species to exist in the atmosphere and $S = 0$ in a steady state (or in a climatological average).

A simple example makes this point clear. Imagine a spatially localized chemical source, such as a smokestack, with a steady wind (no turbulence) and suppose that the pollutant decays at a constant rate in time. The wind blows the pollutant away from the source, leading to a plume which decays chemically. The stronger the wind, the further the pollutant travels before it decays and the more extended the plume. At any point along the plume, eq 1 implies that the transport $-\mathbf{v}\cdot\nabla\chi$ (which is positive) must balance the loss S (which is negative). The decrease in the value of χ along the plume can be attributed to the chemical loss. However, the spatial distribution of chemical loss depends on the wind direction and speed; in the absence of a local source, chemistry can only destroy molecules of χ (and hence produce other molecules) that have first been brought to a given location by transport.

Often the term S in eq 1 acts to relax χ toward some photochemical steady state χ_{pc}, known as "photochemical equilibrium" in atmospheric chemistry. If the photochemical relaxation rate is relatively fast compared to dynamics (e.g., ozone in the upper stratosphere), then $\chi \approx \chi_{pc}$ and transport of χ is unimportant. If, however, the photochemical relaxation rate is relatively slow (e.g., ozone in the lower stratosphere), then the value of χ is pulled away from photochemical equilibrium by transport. In both cases, S is controlled by dynamical transport of χ, but only in the latter case is this of significance for the distribution of χ itself. However, even in the former case where χ is photochemically controlled, transport may be important through its effect on other species which are chemically linked to χ. For example, in the upper stratosphere the ozone concentration depends on the concentration of long-lived species such as N_2O and H_2O which affect the NO_x and HO_x catalytic ozone-destruction cycles. And, of course, dynamics is important for photochemically controlled species through the effect of temperature on reaction rates.

There is a useful analogy here with the interaction between radiation and dynamics. The first law of thermodynamics (again neglecting molecular diffusion) can be written

$$c_v \frac{\partial T}{\partial t} + c_v \mathbf{v}\cdot\nabla T + \frac{p}{\rho}\nabla\cdot\mathbf{v} = Q \qquad (2)$$

where c_v is the specific heat of air at constant volume, T is temperature, p is pressure, ρ is density, and Q is *diabatic* heating or cooling, which we take to be solely radiative for the purposes of the present discussion. The third term on the left-hand side of eq 2 is the heating or cooling due to *adiabatic* contraction ($\nabla\cdot\mathbf{v} < 0$) or expansion ($\nabla\cdot\mathbf{v} > 0$). In the absence of motion and in a steady-state balance, Q must vanish—this is radiative equilibrium, in which the heating due to absorption of incoming solar radiation balances the cooling due to net emission of infrared radiation. Departures from radiative equilibrium ($Q \neq 0$) can therefore be attributed to dynamics. In fact, often radiative forcing is represented in terms of Newtonian cooling, with $Q = -r(T - T_{rad})$ where r is the radiative relaxation rate and T_{rad} is the radiative equilibrium temperature (which depends on the spatial distribution of radiatively active species). In this case it is clear that $T = T_{rad}$ in steady state in the absence of motion and that dynamics generally leads to $T - T_{rad} \neq 0$ and hence $Q \neq 0$. If the relaxation rate r is relatively fast compared to dynamics, then a given dynamical forcing leads to only a small radiative imbalance in T and $T \approx T_{rad}$ to a good approximation. However, if r is relatively slow—which is true nearly everywhere in the atmosphere—then temperature can be significantly affected by dynamics.

To say that dynamics affects chemistry through temperature and through transport is not quite the full story, of course, because of feedbacks. The distribution of radiatively active species determines T_{rad}, which affects T (since dynamical processes control the departure from radiative equilibrium, $T - T_{rad}$) and thereby $\mathbf{v}$. Thus, chemistry, radiation, and dynamics are coupled; in fact, some of the most interesting problems in climate science concern such couplings and feedbacks. There are also significant nonlocal couplings; for example, stratospheric ozone abundance (which is controlled in large part by dynamics) controls the actinic fluxes entering the troposphere and so has a direct impact on tropospheric chemistry. Thus, the atmospheric chemistry of a particular region can rarely be understood as an isolated chemical system; rather, it is intimately tied to dynamics and is thereby coupled to the rest of the atmosphere.

There is an important difference between dynamics on one hand and radiation and chemistry on the other, which is crucial for understanding cause-and-effect. The governing equations of atmospheric radiation and chemistry generally do not possess *chaotic* solutions (although there are exceptions in the case of chemistry), and radiative and photochemical equilibrium are well-defined steady solutions of these equations, to which time-dependent solutions generally converge (in the absence of motion). This is, however, far from the case for dynamics. Solutions of the dynamical equations—even of very simple models of atmospheric dynamics—are generally chaotic and do not approach a steady solution even under constant forcing. This is, after all, why the weather varies from day to day and why each year is different from the previous year (quite apart from long-term forcings on climate such as anthropogenic effects). Thus, "unforced" (which is to say, internal) variability in the atmosphere is due principally to dynamics (including the dynamics of the ocean). (Changes in the atmosphere associated with a volcanic eruption would be an example of "forced" variability.) However, dynamical variability has consequences for atmospheric chemistry, and eq 1 will generally possess chaotic solutions for χ once variations in $\mathbf{v}$ and in T (the latter affecting S) are included. Thus, while internal variability may be dynamical in origin, it can certainly be seen in changes in chemical composition and the dynamical variability may itself depend on chemistry through nonlinear feedbacks.

1.2. Purpose and Scope of This Paper

The purpose of this review is to present an introduction to large-scale atmospheric dynamics, which is aimed specifically at atmospheric chemists but is hopefully of use to other chemists as well. The goal is to describe some of the basic features of the atmospheric dynamical state, with an emphasis on temperature and transport, and to explain those features in simple terms. There are many good textbooks available which provide a detailed physical description of the atmosphere and which are felt to be accessible to atmospheric chemists. However, dynamics textbooks are generally aimed at dynamics students and thus tend to be quite formal in nature. Therefore, while a physical description is certainly provided here, the emphasis is on explanation.

Atmospheric dynamicists generally express their understanding in terms of partial differential equations, which are applied to a variety of approximate models. As there is no single accepted way of formulating dynamical models, the result is a zoo of models and notation which can be an impediment to understanding by the nondynamicist (or even by the dynamicist!). Thus, in this paper equations are kept to a minimum and are used only where they succinctly summarize a point. Rather, the emphasis is placed wherever possible on explanations in terms of fundamental physical principles—which can be expressed in terms of equations if necessary but need not be. Similarly, technical terms are kept to a minimum; where they are unavoidable, the term is defined in the Glossary in section 12. It is hoped that the Glossary may be of value in itself.

There is no particular time scope for the paper, which is pedagogical in nature rather than being a literature review. In addition to describing topics for which understanding is well established, topics of current interest in chemistry–climate coupling are discussed and significant areas where dynamical understanding remains lacking are identified. It is felt to be important for the atmospheric chemist to know which aspects of atmospheric dynamics are well understood and which are not.

Finally, the emphasis herein is squarely focused on large-scale dynamics, with an eye to its relevance for climate chemistry, including ozone. Many significant dynamical processes that are important for atmospheric chemistry are given scant or no treatment, including clouds, boundary-layer and surface processes, and so-called "mesoscale" processes (e.g., convection and fronts). These are important not only for regional air quality but also for climate chemistry but are beyond the scope of this review.

2. Radiative–Convective Description

The atmosphere is heated by absorption of ultraviolet through near-infrared radiation from the Sun and of infrared radiation from the Earth's surface (taken here to include the oceans). It is also heated from below by fluxes of both sensible and latent heat from the Earth's surface. (Sensible heat is that associated with the internal energy of air; latent heat is that associated with the latent heat release by condensation of evaporated moisture.) In fact, since most absorbed solar radiation is absorbed at the Earth's surface, these surface fluxes are a dominant part of the energy balance of the atmosphere. In the absence of motion, the atmosphere can only cool by radiative emission. The radiative balance that one then infers from the distribution of radiatively active trace species and the surface fluxes has temperature maxima at the surface and at around 50 km altitude, the latter arising from ozone heating.

In a compressible fluid such as the atmosphere, high temperatures underlying low temperatures does not necessarily imply convective instability (such as we encounter when heating a pot of soup on the stove) because as air rises its pressure decreases and it cools adiabatically. Taking account of the adiabatic heating and cooling associated with vertical motion, the condition for convective instability is that the entropy decrease with altitude. For dry air, this corresponds to the temperature decreasing with altitude more rapidly than g/c_p (where g is the gravitational acceleration and c_p the specific heat of dry air at constant pressure) or about 10 K/km—known as the (dry) adiabatic *lapse rate*. The radiative equilibrium temperature distribution generally does not satisfy the condition for convective instability but it does so in the lowest few kilometers of the atmosphere when surface fluxes are included.

Thus, the lowest part of the atmosphere is convectively unstable, and rapid vertical overturning ensues. However, atmospheric convection is invariably moist convection: as warm air rises and cools adiabatically, under most conditions the water vapor within the air reaches saturation and begins to condense. This produces clouds, either of liquid water—or, if the convection reaches low enough pressures (which is to say, high enough altitudes), of water ice—and rain. The latent heat release associated with condensation warms the air and causes the convection to penetrate still further. If the phase change of water is included in the calculation of entropy, then the condition for moist convection (provided the air is already saturated) is that the temperature decreases with altitude more rapidly than the moist abiabatic lapse rate—typically around 6.5 K/km.

Convection leads to intense vertical mixing (as with a heated pot of soup on the stove) and thus tends to an equilibrated state of uniform *moist entropy*. To quantify thermodynamics, atmospheric dynamicists use not entropy itself but rather a particular function of entropy known as *potential temperature*, usually denoted by θ. Uniform moist entropy then corresponds to uniform moist (or "equivalent") potential temperature. The mixing is constrained in its vertical extent by conservation of (moist) static energy—the first law of thermodynamics. Above the region of convective overturning, the atmosphere remains in

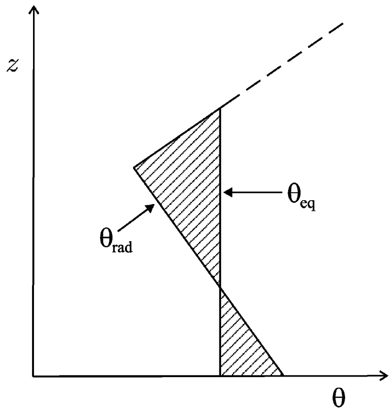

Figure 1. Schematic of the process of convective adjustment, where z is altitude and θ is understood to represent equivalent potential temperature (including effects of moisture). The vertical profile of radiative equilibrium potential temperature, denoted θ_{rad}, decreases with altitude up to a certain level and is therefore convectively unstable. Vertical mixing acts to create a layer of uniform θ; the depth of this layer is limited by conservation of energy. The resulting stable profile is shown as θ_{eq} and coincides with θ_{rad} above the well-mixed region. The hatched regions are not of equal area because of the decrease of density with altitude.

radiative equilibrium. A schematic of this process, known as moist convective adjustment, is depicted in Figure 1. The reality is not quite so simple because radiative processes come into play in determining the equilibrated state, including those associated with clouds. A more realistic calculation of the temperature (not potential temperature) adjustment, reproducing the classical results of ref 1, is shown in Figure 2. Nevertheless, to a first approximation the expected radiative–convective equilibrium state of the atmosphere consists of a convectively adjusted layer with a lapse rate at the moist adiabat (known as the troposphere), overlain by a layer in radiative equilibrium with increasing temperature with altitude (the stratosphere), overlain by another layer in radiative equilibrium but with decreasing temperature with altitude (the mesosphere). This is indeed essentially the case, as shown in Figure 3, with the cold point tropopause occurring at around 16 km altitude in the global and annual mean temperature profile. Deviations from this picture will be discussed in later sections.

Because of the considerable thermal inertia of the Earth's surface (both land and ocean), the highest surface temperatures are found in the tropics where the greatest amount of total sunlight is received throughout the year. Thus, the surface heating is strongest in the tropics, so convection reaches the greatest altitudes there, and one expects the tropopause to be highest in the tropics. This is also the case: the tropopause altitude varies from roughly 16–17 km in the tropics to 7–8 km in the extratropics, as shown in Figure 4. (That the global mean temperature profile has a tropopause located at about the same altitude as the tropical tropopause reflects the fact that the temperature increase with altitude in the lower stratosphere is much weaker than the temperature decrease with altitude in the troposphere.) The high altitude of the tropical tropopause

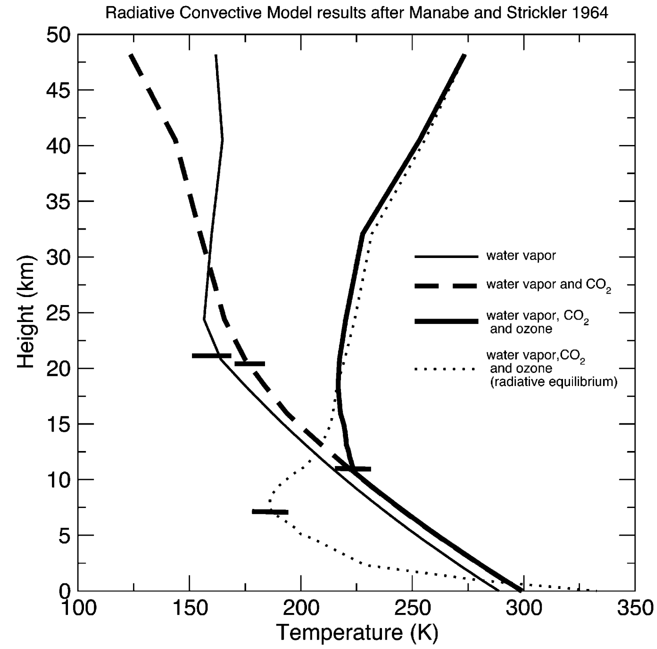

Figure 2. Global and annual mean radiative equilibrium temperature profile (dotted) for an atmosphere containing water vapor, CO_2, and O_3 and the corresponding radiative–convective equilibrium (thick solid) computed using the simple radiative–convective model described in ref 2. Also shown are the radiative–convective equilibrium profiles obtained with O_3 removed (dashed) and with O_3 and CO_2 removed (thin solid). The location of the *lapse-rate tropopause* is indicated for each profile as a thick black horizontal bar. The figure shows that water vapor is the dominant greenhouse gas in the troposphere with CO_2 providing an incremental warming effect of about 10 K (and cooling in the stratosphere); also that O_3, while having essentially no effect in the lower troposphere, is responsible for the existence of the stratosphere and for the temperature minimum (as compared with the water vapor plus CO_2 radiative–convective equilibrium). Above 10 km altitude, ozone has a very significant effect on atmospheric temperature. Figure courtesy of Piers Forster, University of Reading.

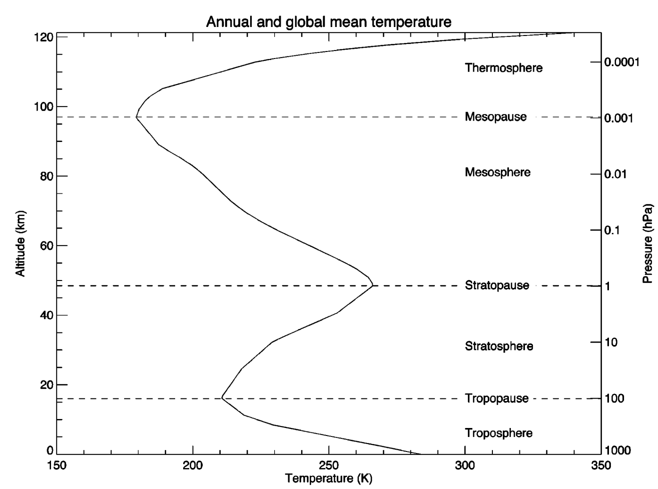

Figure 3. Global and annual mean temperature profile from the COSPAR International Reference Atmosphere (CIRA) data set.[3] The tropopause is marked as the temperature minimum (cold point tropopause).

means that despite the higher surface temperatures in the tropics, tropopause temperatures there are lower than in extratropical latitudes, reaching below

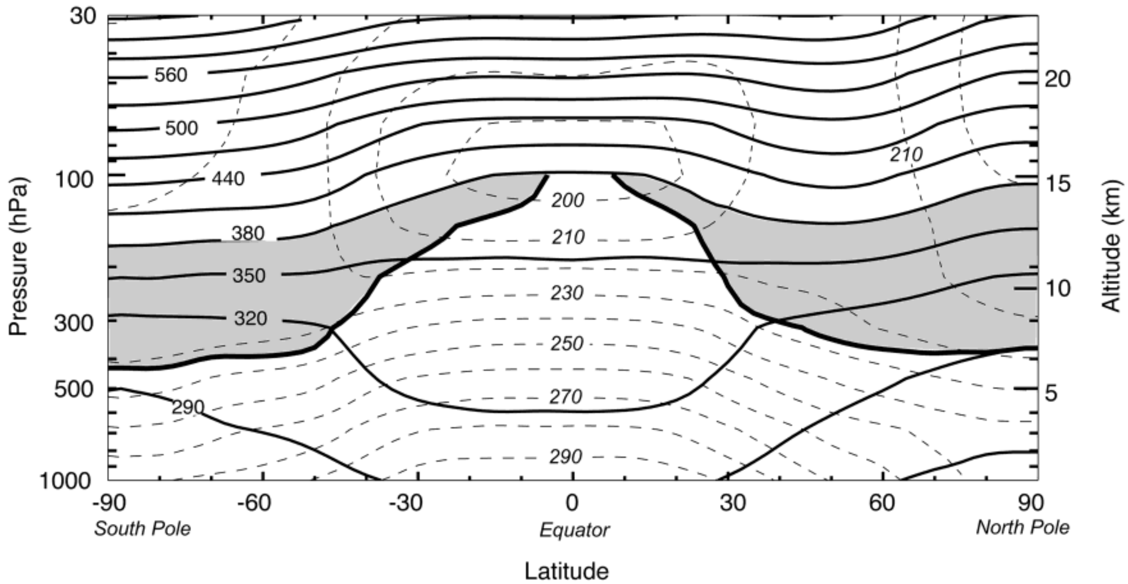

Figure 4. Annual and zonal mean distribution of potential temperature (solid) and temperature (dashed), in Kelvin. The thick line denotes the lapse-rate tropopause. Features to note are the weak stratification in the troposphere (and strong lapse rate, close to moist adiabatic), the strong stratification in the stratosphere, and the temperature minimum at the tropical tropopause. The shaded regions denote the "lowermost stratosphere", consisting of that part of the stratosphere that is connected to the troposphere along isentropic surfaces. (Reprinted with permission from ref 4. Copyright 1995 American Geophysical Union.)

200 K (Figure 4). In fact, the tropical tropopause is one of the coldest places in the entire atmosphere.

The above considerations have important implications for atmospheric chemistry. Namely, there is a layer in the very lowest part of the atmosphere (the troposphere), of between 7 and 17 km depth, within which the temperature decreases strongly with altitude, vertical mixing is relatively rapid (time scale of several days), and there are abundant clouds and rain. Ice clouds (known as cirrus) occur at the low temperatures and pressures characteristic of the tropopause. Clouds and aerosols offer sites for heterogeneous reactions, and rain washes out soluble species. It can be expected that molecules will, typically, not be isolated from the Earth's surface for more than several days at a time. In contrast, the region above (known as the middle atmosphere) is convectively stable and relatively quiescent. Within this region potential temperature increases with altitude and the atmosphere is therefore stratified by quasi-horizontal surfaces of constant potential temperature, known as isentropic (constant entropy) surfaces (Figure 4). Adiabatic motion is, by definition, confined to isentropic surfaces (although these surfaces themselves undulate). It follows that molecules, once in the middle atmosphere, cannot easily reach the Earth's surface. In fact, they can be isolated from the Earth's surface for years.

3. Meridional Temperature Structure and Zonal Winds

As noted above, the highest surface temperatures are found in the tropics. Within the troposphere the lapse rate is relatively uniform, although as is evident in Figure 4 there are certainly variations between the tropics and the extratropics. (In the extratropics, the lapse rate is established not so much through moist convective adjustment as through a process called baroclinic instability, which will be discussed in section 5.) Thus, the highest tropospheric temperatures at any given altitude are also found in the tropics. As Figure 4 illustrates, there is an intermediate range of altitudes from about 7−17 km wherein the tropics lie within the troposphere and the extratropics lie within the stratosphere; in the upper few kilometers of this region, the lowest temperatures at a given altitude are found in the tropics. In the middle atmosphere, out of the reach of convection (and baroclinic instability) so that air parcels do not come into regular contact with the Earth's surface, the thermal inertia is set by the radiative relaxation time which is at most a few weeks. Thus, radiative equilibrium temperatures in the middle atmosphere follow the Sun, with the highest temperatures in the tropics at equinox and over the summer pole at solstice. The observed meridional−vertical distribution of temperature at solstice (Figure 5a) exhibits all these features, except in the upper mesosphere where the meridional temperature gradient is actually the reverse of that expected on radiative grounds. This rather peculiar phenomenon is the result of a wave-driven circulation and will be discussed in section 7.

A fluid normally cannot maintain a horizontal temperature gradient because it implies a horizontal pressure gradient which would lead to downgradient motion. However, a rotating fluid can maintain a horizontal temperature gradient if it is balanced by the Coriolis force associated with motion normal to the gradient. The actual relationship turns out to be between the vertical gradient (or shear) of the horizontal wind and the horizontal temperature gradient. This can be seen as follows. The Coriolis force associated with a zonal flow (i.e., a flow along latitude circles) can be thought of as a correction to

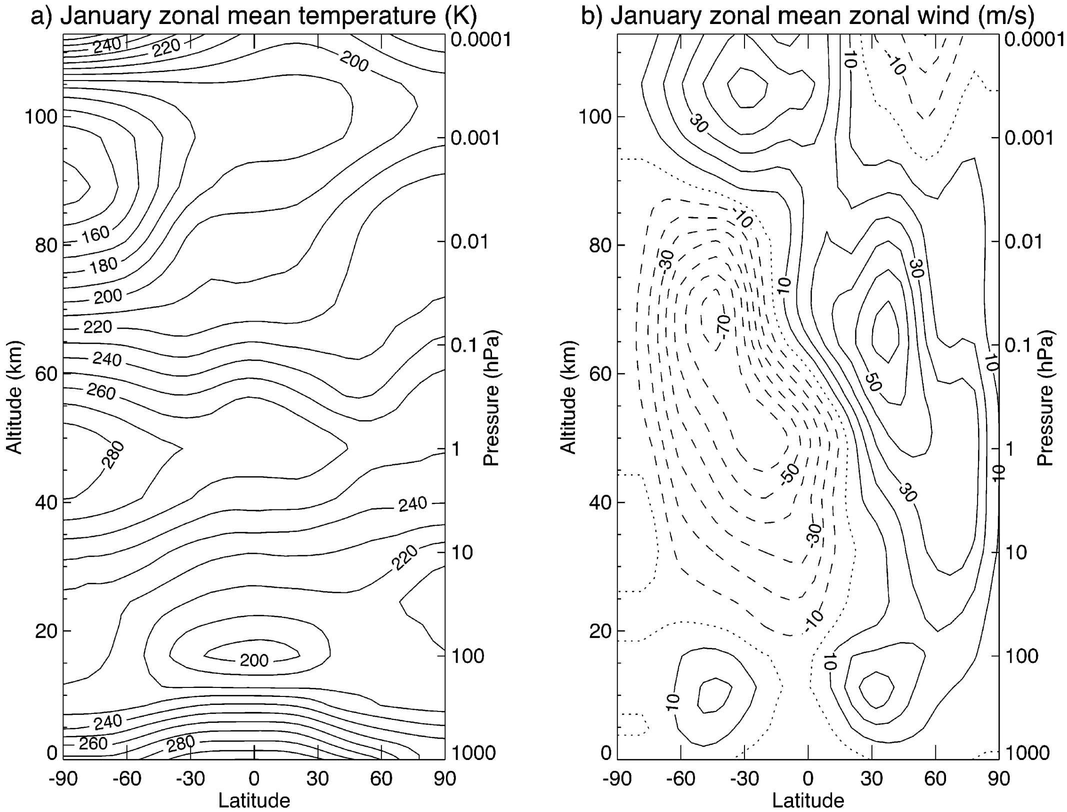

Figure 5. Zonal mean (a) temperature and (b) zonal wind as a function of latitude and height, for January, from the COSPAR International Reference Atmosphere (CIRA) data set.[3] July conditions are, to a first approximation, a mirror image of these.

the projection of the centrifugal force associated with the planetary rotation (which acts normal to the Earth's rotation axis) on the meridional direction. Super-rotating (or eastward) flow provides an excess centrifugal force, acting away from the Earth's axis of rotation and thus toward the equator, and needs to be balanced by a poleward pressure-gradient force; while retrograde (or westward) flow provides a reduced centrifugal force, acting toward the Earth's axis of rotation and thus toward the pole, and needs to be balanced by an equatorward pressure-gradient force. This relationship is known as "geostrophic balance" and in *pressure coordinates* takes the form

$$fu = -\frac{1}{a}\frac{\partial \Phi}{\partial \phi} \tag{3}$$

where u is the zonal wind, $f \equiv 2\Omega \sin \phi$ is the Coriolis parameter, Ω is the rotation rate of the Earth, ϕ is latitude, a is the radius of the Earth, and $\Phi \equiv gz$ is the geopotential height (g times the height z of the given pressure level). The factor of sin ϕ in the Coriolis parameter is a geometric factor, associated with the meridional projection of the correction to the centrifugal force. This projection is zero at the equator, which explains why horizontal temperature gradients are weak in the tropics.

The pressure-coordinate form of the hydrostatic relation is

$$\frac{\partial \Phi}{\partial p} = -\frac{1}{\rho} = -\frac{RT}{p} \tag{4}$$

the last equality following from the ideal gas law $p = \rho RT$, where R is the gas constant for air. If the atmosphere were isothermal, with a constant temperature T_0, then eq 4 implies that the pressure would decrease exponentially with altitude, with an e-folding scale of RT_0/g. The density would also decrease exponentially with altitude with the same e-folding scale, known as the density scale height. Variations in T lead to variations in the e-folding scale of pressure and density, but a typical value is around 7 km. The relation between pressure and altitude can be seen from the vertical axis labels on either side of Figure 3; roughly speaking, pressure decreases by a factor of 10 every 15 km. The fact that pressure decreases quasi-exponentially with altitude motivates the introduction of a "log-pressure" vertical coordinate $z_* \equiv -H \ln(p/p_{00})$, where H is a constant with dimensions of length and p_{00} is a constant reference pressure. In terms of z_*, eq 4 may be rewritten as

$$\frac{\partial \Phi}{\partial z_*} = \frac{RT}{H} \tag{5}$$

The coordinate z_* is an approximate altitude. If H is chosen to be the density scale height RT_0/g of an isothermal reference atmosphere at constant temperature T_0, then the right-hand side of eq 5 is just gT/T_0, and z differs from z_* only insofar as T differs from T_0.

The physical content of eq 5 is that the vertical pressure gradient is inversely proportional to temperature (warm air at the same pressure is less dense than cold air). Consequently, pressure decreases less rapidly with altitude at higher temperatures. From eqs 3 and 5 we obtain

$$f\frac{\partial u}{\partial z_*} = -\frac{1}{a}\frac{\partial^2 \Phi}{\partial \phi \partial z_*} = -\frac{R}{aH}\frac{\partial T}{\partial \phi} \tag{6}$$

Thus, the vertical gradient of the zonal wind (in terms of the log-pressure coordinate z_*) is proportional to the meridional gradient of the temperature, with an equatorward temperature gradient implying an eastward vertical gradient of the zonal wind. This is known as the "thermal wind relation".

As a consequence of the thermal wind relation, the meridional temperature gradients anticipated from radiative–convective equilibrium imply zonal flows (Figure 5b). Assuming that the zonal flow is approximately zero at the Earth's surface (because of boundary-layer friction), then the equatorward meridional gradient of temperature within the troposphere implies eastward (or, as meteorologists say, "westerly") flow, in both hemispheres, with the maximum winds achieved in the subtropics at the tropopause. This accounts for the observed jet streams at about 10 km altitude. Immediately above, the reversed meridional temperature gradient implies a reversed vertical shear of the zonal wind and a weaker zonal flow. Then above the tropical tropopause, within the stratosphere, the temperature minimum over the winter pole implies an increase again in the strength of the eastward jet with altitude in the winter hemisphere, except that now the temperature gradient is concentrated more in subpolar latitudes and the jet maximum shifts poleward accordingly. This is the so-called "polar night jet" or "polar vortex" since the circumpolar flow constitutes a vortex. In the summer hemisphere, in contrast, the temperature maximum over the pole implies, by the thermal-wind relation, a westward vertical shear to the point that the winds become westward through most of the summer stratosphere. Finally, in the mesosphere, the reversed meridional temperature gradient relative to the stratosphere implies a weakening of both eastward and westward jets with altitude above 65 km to the extent that their directions actually reverse in the lower thermosphere above 90 km.

Zonal flows are the dominant flows in the atmosphere. They are required by the thermal-wind relation, in the presence of the inevitable meridional temperature gradients, and are "free" solutions of the governing equations in the sense that they do not require external forcing in order to exist. In contrast, persistent vertical or meridional motions require some kind of forcing (thermal or mechanical, respectively) and for that reason are comparatively weak. However, precisely because of their strength, zonal flows are often of little chemical relevance because they cause long-lived chemical species to be distributed relatively uniformly in longitude, in which case zonal transport is "do-nothing" transport. Exceptions occur for species in the troposphere with lifetimes of a week or less (e.g., CO), where zonal flows can provide long-range, intercontinental transport. Also, as is discussed further in section 8, strong zonal flows act as barriers to meridional transport because flow streamlines then tend to be aligned with the zonal flow. An example is the Antarctic polar vortex, which confines polar air and acts as a "containment vessel" within which ozone-hole chemistry can proceed undisturbed.[5]

4. Rossby Waves

In any physical system, waves require a restoring force. Water waves result from the gravitational restoring force that acts on displacements of the water's surface. In a continuously stratified fluid such as the atmosphere, the same gravitational restoring force acting on displacements of isentropic surfaces leads to internal gravity waves (see *static stability*). The Earth's rotation also provides a restoring force but now in the meridional direction, which leads to Rossby waves. Rossby waves are the theoretical building blocks of large-scale atmospheric dynamics. The existence of Rossby waves is a direct consequence of the conservation following air parcels of a scalar quantity known as the *potential vorticity* (PV), together with the concept—known as *PV invertibility*—that the velocity field is uniquely determined by the distribution of PV.

The simplest model of Rossby waves is that of two-dimensional incompressible motion on the surface of the sphere (known as the "barotropic model"), which can be seen as a very crude representation of the vertically integrated motion of the atmosphere. In this case the PV, denoted by q, consists of the sum of the *vorticity* associated with the motion, known as the "relative vorticity" $\zeta = \hat{\mathbf{z}}\cdot\nabla\times\mathbf{v}$ (where $\hat{\mathbf{z}}$ is the unit vector in the direction of the local vertical) and the vorticity associated with the Earth's rotation, known as the "planetary vorticity" $f = 2\Omega \sin \phi$ (where Ω is the angular frequency of the Earth's rotation and ϕ is latitude). Conservation of q following air parcels then implies that as ϕ increases, ζ must decrease; and as ϕ decreases, ζ must increase. PV invertibility means that $\mathbf{v}$ is determined by ζ; one speaks of the velocity field "induced" by a vorticity distribution, much as an electric field is induced by an electric charge.

In the absence of motion, we have simply $q = f$, so the PV contours are aligned along latitude circles and PV has a northward gradient at all latitudes (except right at the poles). Now consider nonzero but sufficiently weak motion, so that the PV contours (which deform with the motion, since PV is conserved following air parcels) are displaced but still retain their essentially zonal character, as depicted in Figure 6. Northward movement of a PV contour implies a larger value of f, which implies a negative value of ζ—corresponding to a clockwise circulation. Southward movement likewise induces a positive (counterclockwise) value of ζ. If one now considers the action of the induced velocity field on the background PV gradient, the effect is to move a sinusoidal pattern of undulations westward, as shown in Figure 6. These are Rossby waves, which always propagate to the west, relative to any background wind. The *zonal phase speed* c of any wave satisfies $c = u + \hat{c}$, where u is the zonal wind velocity and $\hat{c}$ is the

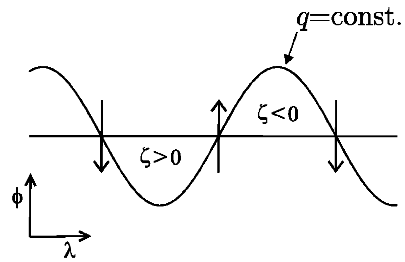

Figure 6. Schematic of Rossby wave propagation. The horizontal line shows a potential vorticity contour (q = const.) for a resting flow, which lies along a constant latitude ϕ. The wavy line shows this q contour displaced by a sinusoidal wave. Where the q contour has moved northward, conservation of q following the motion implies that the relative vorticity ζ is negative; where the q contour has moved southward, ζ is positive. These disturbances to the ζ field induce, respectively, clockwise and counterclockwise velocities, with meridional velocities as shown by the arrows. The action of these arrows on the background q distribution is such as to cause the phase of the wave to move to negative λ, i.e., westward.

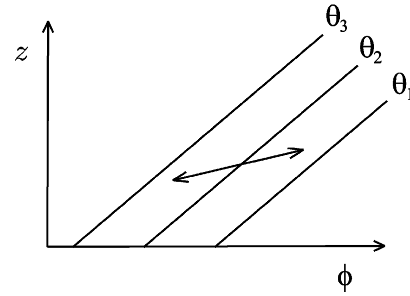

Figure 7. Physics of baroclinic instability (Northern Hemisphere case). In the troposphere, surfaces of constant θ slope upward-poleward or downward-equatorward (Figure 4). Under these conditions, if air parcels move at an angle as indicated by the double-headed arrow, they can ascend and gain buoyancy or descend and lose buoyancy, even though the atmosphere is convectively stable. This leads to instability, which can be seen as a form of slantwise convection. Sloping θ surfaces owe their existence to the Earth's rotation, which provides a balance between the meridional pressure gradient force and the Coriolis force associated with the corresponding zonal flow, satisfying the thermal wind relation.

"intrinsic zonal phase speed", namely, the zonal phase speed in the frame of reference moving with the mean zonal wind. Thus, the phase speed in the ground-based reference frame, c, is Doppler shifted by the wind. For sound waves, for example, $\hat{c}$ is the speed of sound but c can be greater or less than $\hat{c}$.

In the troposphere, Rossby waves are forced by the topography of the Earth's surface and by the thermal forcing associated with land−sea contrasts. Both are of planetary scale and since they are fixed in space tend to force stationary (zero phase speed) waves, with $c = 0$. Since Rossby waves propagate westward relative to the background wind, $\hat{c} < 0$ and stationary Rossby waves require an eastward background wind ($u > 0$) in order to satisfy $u + \hat{c} = c = 0$. As we have already seen, this is the case in the troposphere; furthermore, the zonal wind speeds and spatial scales are conducive to the efficient generation of Rossby waves. Stationary Rossby waves shape climate by introducing longitudinal asymmetries in atmospheric conditions, as reflected in the Aleutian and Icelandic wintertime lows and the subtropical highs. By so doing they can create closed circulation cells which can trap air and cause it to recirculate in the same region.

While stationary Rossby waves shape climate, they also shape climate variability. This is because, to a first approximation, stationary Rossby waves can be understood as the linear response to surface forcing and linear responses can be superposed—hence anomalous forcing (relative to the climatological mean) creates an anomalous response. For example, the El Niño/Southern Oscillation (ENSO), a tropical atmosphere−ocean interaction, leads to climatological anomalies in sea surface temperatures over the tropical Pacific Ocean which can persist for a period of a year or longer.[6] These tropical temperature anomalies force anomalous stationary Rossby wave trains which propagate into the extratropics[7] and cause the well-known ENSO effects over North America.

As well as propagating horizontally, stationary Rossby waves can propagate vertically. As noted earlier, they can only propagate in an eastward wind; however, the wind cannot be too strongly eastward or the stationarity condition $u + \hat{c} = 0$ is impossible to meet. The intrinsic propagation speed of Rossby waves, $-\hat{c}$, increases with horizontal wavelength, so the critical zonal wind speed beyond which stationary Rossby waves cannot exist depends on wavelength and is greater for the longer waves. It happens that for typical wintertime stratospheric conditions, only the largest-scale waves, of *zonal wavenumbers* 1−3, can propagate into the stratosphere.[8] Thus, the wintertime stratosphere is buffeted by planetary-scale Rossby waves which continually disturb it. In contrast, stationary Rossby waves cannot propagate at all into the summer stratosphere where the wind is westward, and the summer stratosphere is therefore relatively quiescent.

5. Baroclinic Instability

Surface forcing tends to generate stationary Rossby waves of planetary scale. Within the extratropical troposphere, however, there is also internal generation of smaller-scale Rossby waves due to a process called "baroclinic instability". Baroclinic instability is a thermally driven instability constrained strongly by the Earth's rotation and driven by PV dynamics. Its energy source is the gravitational potential energy associated with meridional temperature gradients. When the meridional temperature gradient is nonzero, isentropic surfaces are sloped. As depicted in Figure 7, although the atmosphere is convectively stable, it is nevertheless possible for air to rise in altitude and gain buoyancy by moving into a region with a lower value of θ. This is the basic physics of baroclinic instability.

There are various theoretical models of baroclinic instability,[9] and it is understood to be the fundamental process behind the development of large-scale mobile weather systems. In a way, it can be thought of as a form of sideways convection; while convection fluxes heat upward, baroclinic instability fluxes heat (mainly) poleward. Thus, baroclinic instability is

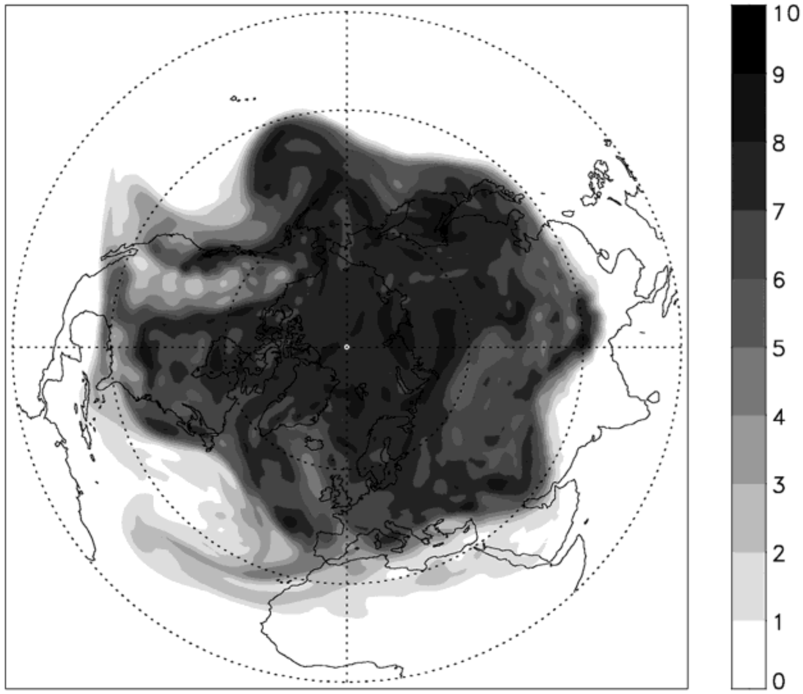

Figure 8. Polar stereographic map (Northern Hemisphere) of the instantaneous distribution of potential vorticity on the 345 K isentropic surface, from the Canadian Meteorological Centre analysis for a day in January, in so-called "PV units" (1 PVU = 10^{-6} m^2 s^{-1} K kg^{-1}). The 345 K isentropic surface intersects the tropopause (cf. Figure 4), with this intersection roughly corresponding to 2 PVU—the edge of the region of high PV. This PV edge also corresponds to the jet stream, which is seen to meander in longitude. The undulations in the PV edge are Rossby waves, although they can become severely distorted and quite nonlinear. Figure courtesy of Yves Rochon, Meteorological Service of Canada.

driven by the meridional temperature gradient within the troposphere and acts to reduce that gradient. Without baroclinic instability, the polar regions would be much colder than they are. It turns out that meridional temperature gradients at the Earth's surface are essential to the dynamics of baroclinic instability; the same meridional temperature gradients, if they occurred in the stratosphere, would not lead to instability.[9] Thus, baroclinic instability is largely confined to the troposphere, although it does penetrate into the lowermost stratosphere.

While the initial development of baroclinic instability can be regarded as an interaction of Rossby waves,[10] fully developed baroclinic disturbances are nonlinear and one may view baroclinic instability more generally as large-scale, quasi-horizontal turbulence.[11] Just as convective instability leads to an equilibrated (stable) state through the process of convective adjustment, one might imagine baroclinic instability to lead to an equilibrated state through a process of baroclinic adjustment. However, it is not possible to entirely remove the meridional temperature gradient between the equator and the pole, and there are fundamental differences between convective and baroclinic turbulence.[11] The central theoretical problem of baroclinic instability is to determine this equilibrated state of the atmosphere, but it has to be said that this problem remains unsolved. A theory of baroclinic adjustment would explain both the static stability and the meridional temperature gradient of the extratropical troposphere and their dependence on external forcing.

As a result of baroclinic instability, the earlier picture of convective instability defining the troposphere needs to be modified in the extratropics. There, convection is relatively shallow and baroclinic instability is the dominant dynamical process, leading to vigorous large-scale quasi-horizontal mixing within the troposphere. In fact, tropopause height must depend on the static stability of the troposphere, and without a theory of the latter we can claim no real understanding of the former. In midlatitudes, the tropopause slopes upward and equatorward (Figure 4). Since, all else being equal, PV is proportional to static stability, the tropopause corresponds to a strong meridional gradient of PV on those isentropic surfaces that pass through it. The distribution of PV on the 345 K isentropic surface (corresponding roughly to 10 km altitude) is shown for a typical day in January in Figure 8; high values of PV correspond to the strongly stratified stratosphere and low values (below about 2 PVU) to the weakly stratified troposphere. Thus, the midlatitude tropopause can be regarded as a sideways interface just as much as a vertical one. Undulations in this interface represent baroclinic disturbances and are seen to be of zonal wavenumbers 5−6, as predicted by baroclinic instability theory. A strong PV gradient, while apparently an anti-diffusive phenomenon, can be established by strongly inhomogeneous stirring of PV, removing gradients in the well-stirred region and enhancing them at the edges. This appears to be the natural outcome of baroclinic instability and provides an explanation for the midlatitude tropopause.[12] However, since the PV stirring associated with baroclinic instability is itself affected by the structure of the tropopause (through the latter's effect on the subtropical jet), the equilibration process is highly nonlinear and a predictive theory of it remains elusive.

As noted above, baroclinic instability is driven by surface temperature gradients and thus is most intense where the surface temperature gradients are most intense—in midlatitudes, below the jet stream. In the Northern Hemisphere there is, furthermore, a longitudinal concentration of baroclinic activity into so-called “storm track” regions over the western Atlantic and Pacific oceans, organized by planetary wave structures and surface forcing. While baroclinic instability is a large-scale phenomenon (typically of zonal wavenumbers 5−6 or a length scale of roughly 1000 km), it leads to the development of so-called “mesoscale” structures (meaning length scales in the tens to hundreds of km), notably fronts.[13] Fronts are forced by the large-scale flow and develop strong vertical motion and significant condensation and precipitation. They are thus important for atmospheric chemistry through vertical transport and rain-out.

6. Large-Scale Tropospheric Circulation

In the tropics, latent heat release in convection drives deep vertical motion. This occurs primarily over a narrow band of latitudes known as the “intertropical convergence zone” (ITCZ), which moves back and forth across the equator, essentially following the Sun, during the course of the seasonal cycle. Once the rising air reaches its level of neutral buoyancy, it cannot rise much further and begins to flow poleward. How far poleward it can reach is constrained by conservation of angular momentum: as the air moves poleward and toward the Earth’s axis of rotation it must develop an eastward velocity, yet the eastward zonal wind speed aloft is limited (through the thermal wind relation) by the maximum possible meridional temperature gradient, which is constrained by the radiative forcing. (As noted in section 5, baroclinic instability fluxes heat poleward and only weakens this gradient.) This is the basic physics behind the observed “Hadley circulation”,[14] which is a pair of massive overturning cells reaching to roughly 30° latitude in both hemispheres and to around 11−12 km altitude (the top of the level of moist convective adjustment). The Hadley circulation is responsible for poleward heat transfer within the tropics; its descending branches are of course undersaturated and undergo adiabatic warming, the moisture having been largely lost as rain in the ascent, which accounts for the hot, dry desert regions found in the subtropics. In contrast, the deep tropics experience heavy rainfall.

Embedded within the Hadley circulation are longitudinal circulation cells, known as the “Walker circulation”, associated with longitudinal variations in sea surface temperature and surface heating. These modulate the intensity of the Hadley circulation. Air parcels do not actually follow a single Walker circulation cell; rather, they move chaotically through the tropics. However, there appears to be a strong dynamical impediment to cross-equatorial motion.[15] This is at least partly explained by the process of *inertial instability*, which acts to break up any cross-equatorial flow. It follows that there is a notable inter-hemispheric contrast in many chemical species, as is evident for example in CO_2, despite its long lifetime. While horizontal mixing times might be several weeks in the troposphere, within each hemisphere, these times extend to a year or more for global mixing.

In the extratropics, baroclinic instability drives a *meridional circulation*, with poleward flow in the upper troposphere and a return flow in the surface layer. This refers to the net Lagrangian transport of fluid parcels; the time-mean (Eulerian) meridional velocity is actually in the opposite direction (the so-called Ferrel cell) but is overwhelmed by the transport by baroclinic disturbances. This circulation results in poleward heat transfer within the extratropics and, together with the Hadley circulation in the tropics, acts to cool the tropics below radiative equilibrium and to warm the extratropics above radiative equilibrium. As a consequence, the net radiative forcing is positive in the tropics, where it is warm, and negative in the extratropics, where it is cold. Although radiative forcing provides zero net energy input (on average), it produces entropy which drives the circulation. Thus, the troposphere, which represents about 80% of the atmospheric mass, is essentially a classical heat engine[16]—temperature contrasts driving a thermally direct circulation.

7. Middle Atmosphere Circulation

While the tropospheric circulation is thermally driven, the middle atmosphere (stratosphere and mesosphere) is quite different in character. As noted earlier, the radiative equilibrium state in the middle atmosphere is dynamically stable. Thus, no motion need arise apart from the zonal flow required by thermal wind balance and the time-dependent meridional circulation resulting from the slight radiative imbalance associated with the annual cycle[17]—this imbalance being a consequence of the finite radiative relaxation time. Yet there is a definite meridional circulation in the middle atmosphere, even under solstice conditions when the radiative conditions are quasi-steady. This can be inferred, for example, from the distribution of long-lived chemical species such as CH_4, shown in Figure 9. CH_4 has a surface source and a middle atmospheric sink via oxidation, so CH_4 undergoes *chemical aging* in the stratosphere, and high values of CH_4 indicate air that has only recently entered the stratosphere. Thus, from the observed distribution of CH_4 one can infer persistent tropical upwelling and extratropical downwelling in the stratosphere. Such a circulation cannot be produced radiatively.

Thus, what drives the observed meridional circulation? To answer this question we must consider the thermodynamic and angular momentum balances (first law of thermodynamics and Newton’s second law). In the presence of vertical entropy stratification, which is characteristic of the middle atmosphere, persistent vertical motion requires diabatic heating or cooling in order to move air parcels across isentropic surfaces. However, this is no real constraint since the relaxational nature of radiation easily

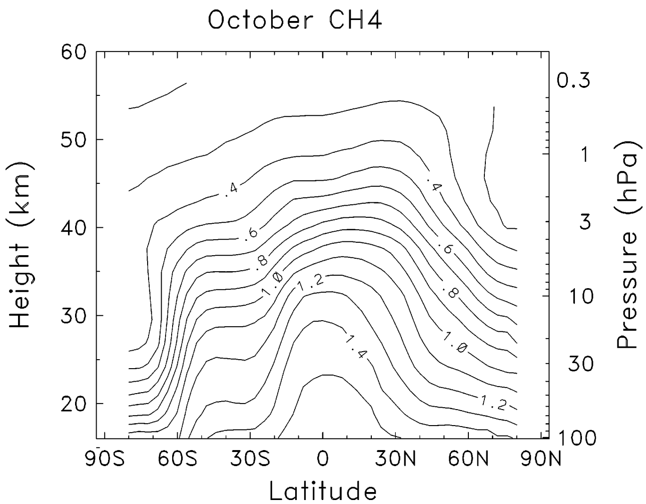

Figure 9. Latitude−height cross-section of CH_4 mixing ratio for October, in parts per million by volume, from a climatology produced from the HALOE and CLAES instruments on the UARS satellite. The data set is described in ref 18. As CH_4 is emitted at the Earth's surface and destroyed in the stratosphere by oxidation, high values of CH_4 correspond to air that has only recently entered the stratosphere while low values correspond to air that has been *chemically aged*. Thus, from the CH_4 distribution one can immediately infer the sense of the stratospheric *Brewer−Dobson circulation*—ascent in the tropics and descent in the extratropics. Figure courtesy of Bill Randel, National Center for Atmospheric Research.

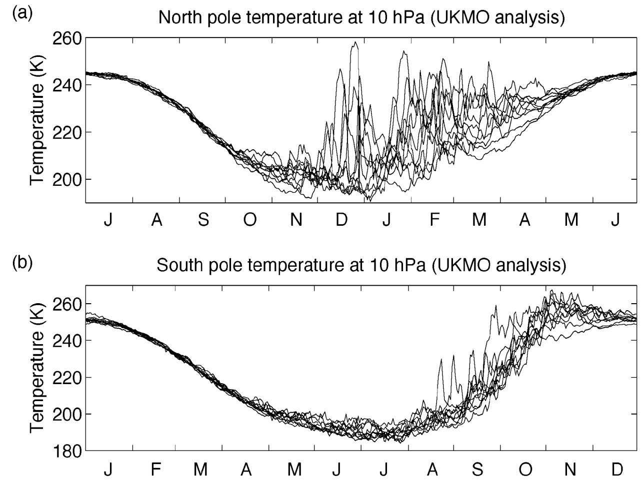

Figure 10. Time series of polar temperatures at 10 hPa for (a) North Pole and (b) South Pole for the period 1993−2002, as produced by the Met Office stratospheric analysis. Each year is represented with a different curve, and the two panels are offset by 6 months so that the seasons are aligned. The intermittent warm events are *stratospheric sudden warmings*. All of the high values over the South Pole between mid-August and late October occurred in the anomalous year of 2002, when the Antarctic ozone hole split in two—the signature of a zonal wavenumber 2 sudden warming.

allows this. If, for example, downwelling is suddenly imposed at a given location, the immediate response is for the atmosphere to warm adiabatically. The adiabatic warming raises the temperature above radiative equilibrium, inducing radiative cooling, and this radiative cooling allows the downwelling to continue. However, the same considerations do not apply to meridional motion. Because of the Earth's rotation, persistent meridional motion requires a torque in order to move air parcels across surfaces of constant angular momentum, either away from or toward the Earth's axis of rotation. As there are no external angular momentum sources in the atmosphere, such torques can only arise from angular momentum transfer by atmospheric motions. If, for example, poleward motion is suddenly imposed at a given location, the immediate response is for the zonal wind to accelerate (toward the east). However, there is no particular reason this would lead to a change in the angular momentum transfer that would provide exactly the right amount of negative torque, and the motion is unlikely to be sustainable.

The above considerations highlight the fundamental difference in the nature of the thermodynamic and angular momentum balances in the middle atmosphere and lead to the concept of *downward control*.[19] Although diabatic heating must coincide with persistent rising motion and diabatic cooling with persistent sinking motion, this is only a diagnostic relationship and does not imply causality. In fact, it is the diabatic heating or cooling that is caused by the vertical motion, while the vertical motion is caused by wave-induced angular momentum transfer (commonly referred to as *wave drag*) via the meridional circulation. Morever, the circulation induced by wave drag extends downward in the steady-state limit, pulling the atmosphere away from radiative equilibrium. Thus, wave drag drives the middle atmosphere circulation, while temperature is determined by a balance between radiative equilibrium and circulation-induced departures from that state. One has to think of radiative forcing as determining the radiative equilibrium state and *not* as determining the net radiative heating or cooling rates.[20]

This picture is oversimplified in several respects. First, radiative cooling depends nonlinearly on temperature, so the radiative damping of a wave can give a net local heating or cooling. (This effect is not represented in the Newtonian cooling approximation commonly considered by dynamicists, which is linear.) Second, there are feedbacks. In particular, wave drag depends on the zonal wind structure which is determined by temperature [through eq 6] and thus reflects the radiative properties of the atmosphere. So, for example, a radiative perturbation (e.g., ozone loss) could lead to lower temperatures, and the resulting zonal wind changes could lead to a change in the wave drag which would induce a change in circulation. However, this last feedback is a second-order effect and in distinction to the chain of causality discussed in the previous paragraph is not predictable. Thus, it is useful to ignore such feedbacks in a first analysis. (On the other hand, the time-dependent circulation induced by a radiative change in the annual cycle[17] should be regarded as a first-order effect.)

Note that while the meridional circulation pulls the middle atmosphere away from radiative equilibrium locally, it does not do so globally. This is because at any given altitude, the warming induced in regions of mean descent is balanced, by mass conservation, by the cooling induced in regions of mean ascent. (The

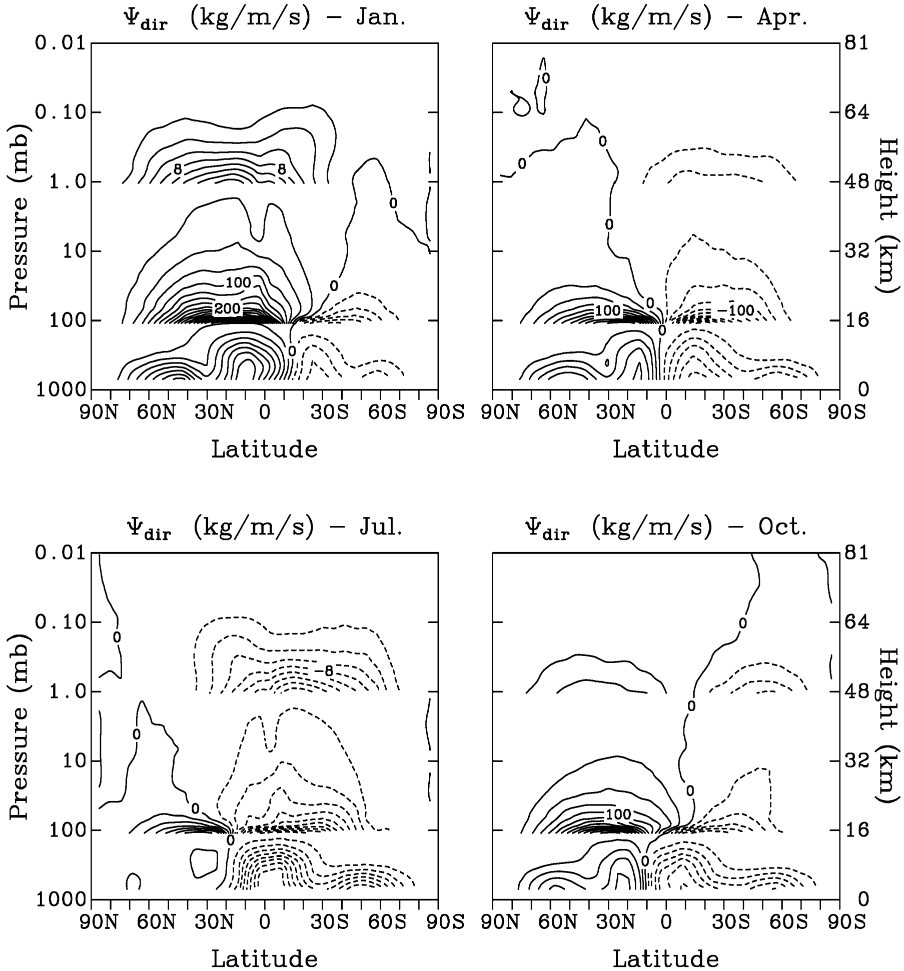

Figure 11. Mass streamfunction for the four cardinal months taken from an integration of the Canadian Middle Atmosphere Model, a comprehensive middle atmosphere climate simulation model. What is shown is actually the streamfunction of the so-called *residual or TEM circulation*, which is closely related to the diabatic transport of chemical species.[22,23] A different set of contour intervals is used in different regions so that the middle atmosphere circulation can be seen. Solid lines indicate counterclockwise circulation and dashed contours clockwise. Contour intervals are 500 kg m^{-1} s^{-1} between 1000 and 100 hPa, 25 kg m^{-1} s^{-1} between 100 and 1 hPa, and 2 kg m^{-1} s^{-1} above 1 hPa. Evident are the poleward circulations in both the troposphere and stratosphere, strongest in the winter hemisphere, and the pole-to-pole solstitial circulation in the mesosphere. (Reprinted with permission from ref 24. Copyright 1997 Canadian Meteorological and Oceanographic Society.)

cancellation is exact only if the static stability and the radiative relaxation rate are independent of latitude, but this is true to a reasonable approximation.) The situation is different in the troposphere, where vertical heat fluxes due to convective and baroclinic instability play a first-order role in the energy balance and cause global departures from radiative equilibrium (Figure 2).

From the insight of downward control, much follows.[21] In the extratropical stratosphere, wave drag comes dominantly from planetary-scale Rossby waves which propagate up from the troposphere and dissipate in the stratosphere. Rossby waves always exert a negative wave drag where they dissipate. Thus, Rossby-wave drag drives a poleward circulation within the stratosphere, and mass conservation then dictates descending motion in the extratropics and ascending motion in the tropics, as already inferred from Figure 9. This is the origin of the celebrated *Brewer–Dobson circulation*. To reiterate: the upward motion in the tropics is *not* because “hot air rises”, as is sometimes mistakenly stated; in fact, at the tropical tropopause, it is *cold* air that is rising! Rather, the upward motion is mechanically driven and results from stratospheric wave drag.

As noted in section 4, stratospheric Rossby waves are restricted to the winter hemisphere where the winds are eastward, so the Brewer–Dobson circulation is mainly a wintertime phenomenon. It is moreover stronger in the northern winter than in the southern winter, because the distribution of land masses is such as to generate stronger planetary waves in the Northern Hemisphere. Since stronger downwelling implies warmer temperatures (because of adiabatic warming), this explains why the Arctic wintertime stratosphere is warmer and has weaker circumpolar eastward winds than the Antarctic wintertime stratosphere. The stronger dynamical forcing in the Arctic often leads to *stratospheric sudden warmings*, where the downwelling becomes focused over the pole to the extent that polar temperatures increase by several tens of Kelvin over the course of a few days. Figure 10 shows time series of polar temperatures at 10 hPa (approximately 35 km altitude) for the period 1993−2002. It is evident that in the late summer and early fall, the two hemispheres are both quiescent and rather similar in temperature. However, in the winter, the Arctic is much warmer and more active than the Antarctic. The one exceptional year in the Antarctic is 2002, which accounts for all the high temperatures between mid-August and late October. In that year the ozone hole split in two in late September, in the first ever recorded Antarctic stratospheric sudden warming.

In the mesosphere, wave drag comes dominantly from gravity waves which also propagate up from the troposphere and dissipate. Unlike Rossby waves, gravity waves can exert either a positive or a negative wave drag, but the drag tends to oppose the zonal flow in the mesosphere. Thus, the eastward zonal flow of the winter hemisphere induces negative gravity-wave drag and meridional flow toward the pole, while the westward zonal flow of the summer hemisphere induces positive gravity-wave drag and meridional flow toward the equator. Together, this implies a pole-to-pole solstitial circulation, with mass conservation then implying rising motion (and lower temperatures) over the summer pole and sinking motion (and higher temperatures) over the winter pole. In fact, the wave-driven circulation is so strong that the mesopause temperatures are actually higher over the winter pole than they are over the summer pole (Figure 5a).

The meridional circulation cannot be observed directly, given present instrumentation, but the circulation obtained from a comprehensive climate simulation model is shown in Figure 11 and illustrates all of these features. It may be noted, with reference to section 6, that the tropospheric mass circulation is also evident in this figure. As with the stratospheric circulation, it exhibits a strong seasonality but this is driven by the seasonality of the meridional temperature gradient (strongest in the winter hemisphere) rather than by the seasonality of Rossby-wave propagation.

8. Transport and Mixing

The earlier discussion has highlighted the different nature of transport and mixing in different parts of the atmosphere, which is mainly related to considerations of atmospheric stability. This heterogeneity is illustrated by Figure 4, which shows the latitude−height structure of the atmosphere in terms of potential temperature. Recall that adiabatic motion must occur along isentropic surfaces.

The shortest transport time scales are found in the planetary boundary layer, which covers the lowest kilometer or so of the atmosphere and within which air is mixed on a time scale of a few hours. The next shortest time scales are in regions of convection, where surface forcing warms and moistens air parcels to the extent that they become positively buoyant and are lofted from the boundary layer into the free atmosphere. Convective transport is an extremely complicated process involving uplift, detrainment and entrainment with the surrounding air, scavenging by precipitation, isolated downdrafts, and mixing. In the tropics, convection regularly reaches about 11−12 km altitude and up to this level the environmental lapse rate is observed to follow a moist adiabat. To a first approximation then, the tropical atmosphere below 11−12 km undergoes moist convective adjustment and can be regarded as being well mixed (in the vertical) on a time scale of a day or two. Intercontinental transport in longitude relies on the zonal wind and can take a week or so; for a species to be well mixed longitudinally, its lifetime would have to exceed several weeks. Time scales for meridional transport out of the tropics are likewise several weeks, while interhemispheric transport is much slower. In the extratropics, transport and mixing time scales are also on the order of weeks, with mixing primarily occurring quasi-horizontally on the sloping isentropic surfaces, albeit with a nonnegligible diabatic component.

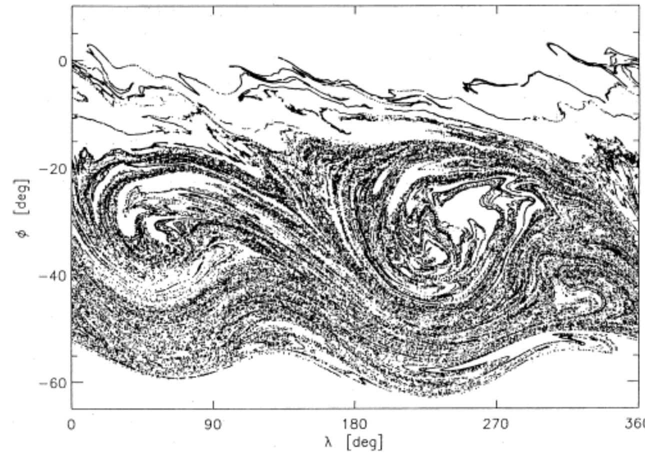

Figure 12. Thirty-day particle advection calculation in the Southern Hemisphere using isentropic winds from the Canadian Middle Atmosphere Model during a model July. The isentropic level is 1000 K, which corresponds roughly to 35 km altitude, the middle stratosphere, and the particles are initially aligned along latitude circles. There is coherent horizontal stirring by a breaking zonal wavenumber 2 planetary Rossby wave in the stratospheric “surf zone”. (Reprinted with permission from ref 26. Copyright 1999 American Meteorological Society.)

Within the stratosphere, transport time scales are considerably longer. The overturning time scale of the Brewer−Dobson circulation is around 5 years. Planetary-scale Rossby waves dissipate in the wintertime stratosphere in a process known as “wave breaking”, which is analogous to the breaking of water waves on a beach except that since the velocities are quasi-horizontal, the waves break sideways.[25] (Water waves grow in amplitude as they approach a beach and the depth of the water decreases; atmospheric waves grow in amplitude as they propagate upward and the density of air decreases.) This leads to intense quasi-horizontal mixing on isentropic surfaces, as is evident from Figure 12, which shows the result of a particle advection calculation on the 1000 K isentropic surface (approximately 35 km altitude) in the Southern Hemisphere winter using winds from a comprehensive climate simulation model. Large-scale stirring by a planetary wave with zonal wavenumber 2 is evident in this figure, extending over several tens of degrees latitude, creating a well-mixed “surf zone”. This process is similar to the

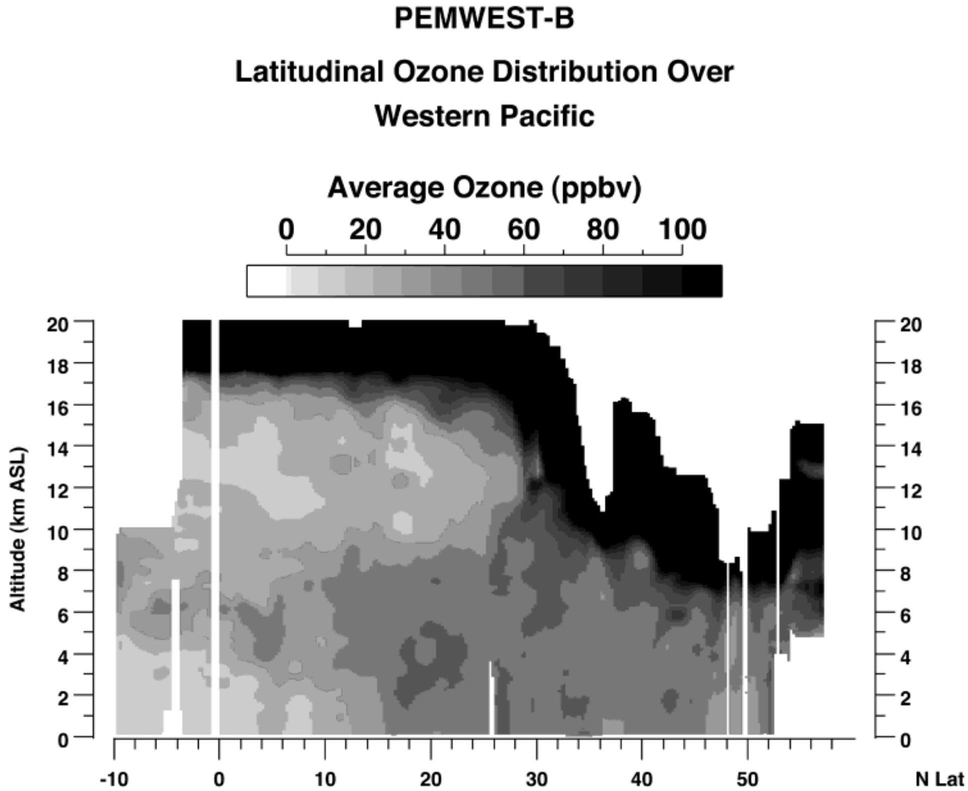

Figure 13. Composite latitudinal ozone distribution obtained with an airborne lidar over the western Pacific during the NASA Pacific Exploratory Mission conducted in February–March 1994. The distribution of ozone corresponds closely with the structure of the tropopause. (Reprinted with permission from ref 31. Copyright 1997 American Geophysical Union.) See also ref 32.

PV stirring associated with baroclinic instability, discussed in section 5. The mixing time scale associated with such large-scale stirring is several weeks.

The balance between the mean vertical motion and horizontal mixing leads to a characteristic shape to concentrations of long-lived chemical species, as shown for CH_4 in Figure 9. The concentration of many such species is closely related to their stratospheric "age" (see *chemical aging*), so Figure 9 shows that the oldest air is found in the polar regions of the upper stratosphere and (during October) in the Antarctic lower stratosphere. Evidently age is directly related to dynamics: stronger planetary wave drag implies a faster meridional circulation and more rapid horizontal mixing, both of which would act to reduce stratospheric age. This provides a direct way by which dynamical changes could affect the lifetimes of certain greenhouse gases and ozone-depleting substances. It also means that the impact of climate forcings in the stratosphere can be delayed by several years. For example, the stratospheric sulfate aerosol resulting from a volcanic eruption warms the lower stratosphere through increased absorption of thermal radiation, and all else being equal this should allow more water vapor into the stratosphere through a modulation of the *freeze-drying mechanism* (see also section 9). Yet there would be a delay of several years before this perturbation to the stratospheric HO_x budget had an impact on upper stratospheric ozone, because of the long time scales involved in transport to the upper stratosphere.[27]

Horizontal mixing is spatially inhomogeneous, as manifested in Figure 12; in particular, transport across the edges of the surf zone is much slower than transport within it.[23] This leads to weak meridional gradients of long-lived species within the surf zone and to enhanced gradients at the edges. For this reason, the edges are sometimes referred to as "mixing barriers". Mixing barriers create distinct air masses and are reflected in the structure of correlations between long-lived chemical species.[28,29] However, one must not necessarily interpret them as barriers to transport.[27] In principle, there could be a uniform flux (or transport) through a given region with enhanced gradients just reflecting subregions of reduced meridional diffusivity (or mixing). At the subtropical edge of the surf zone, for example, there is a systematic poleward flux from the Brewer–Dobson circulation across the mixing "barrier". It follows that while the tropics are relatively isolated from the extratropics, forming a distinct "tropical pipe",[30] the converse is not true. However, at the polar edge of the surf zone—which corresponds to the edge of the wintertime polar vortex—the barrier works both ways. This is because for Rossby waves in the extratropics there is a tight connection between horizontal mixing and the net meridional mass flux. (In the subtropics, this connection is weakened by transient effects.)

The polar vortex isolation is particularly strong in the Antarctic and is clearly evident in the sharp meridional gradient in CH_4 found around 65°S in Figure 9. The extremely low values of CH_4 found within the vortex in October are indicative of air that has descended unmixed during the Antarctic winter. In fact, it is this vortex isolation together with the low temperature (which are just two sides of the same coin, since they both result from relatively weak downwelling compared to the Arctic and are linked by the thermal wind relation), which allows the anomalous chemistry associated with the Antarctic ozone hole to occur.[5]

The tropopause itself also represents a mixing barrier, as can be seen in measurements of many chemical species which exhibit sharp contrasts in concentration across the tropopause. (Yet this does not prevent significant cross-tropopause transport.) An example is ozone, shown in Figure 13. Earlier it was noted that the extratropical tropopause is a region of enhanced horizontal PV gradients on isentropic surfaces (Figure 8), the result of inhomogeneous stirring by tropospheric baroclinic disturbances. The same velocity field that stirs the PV also stirs chemical species, and one therefore expects an enhanced horizontal gradient on isentropic surfaces of any long-lived chemical species at the tropopause.

9. Lowermost Stratosphere and Tropical Tropopause Layer

It is evident from Figure 4 that there is a distinct region of the stratosphere, known as the "lowermost stratosphere", which is connected to the troposphere along isentropic surfaces.[4] Above this region, which is to say above about 17 km or the 380 K isentropic surface, air is well isolated from the troposphere; moreover, because of the persistent downwelling in the extratropics, air can only enter the deep stratosphere through the tropical tropopause. On the other hand, air can enter the lowermost stratosphere either via downwelling from the deep stratosphere or via horizontal mixing from the troposphere. However, tropospheric air entering the lowermost stratosphere cannot ascend into the deep stratosphere and will eventually get flushed back into the troposphere. Downwelling dominates during the winter because of wintertime stratospheric planetary-wave drag, and during that period of the year air in the lowermost stratosphere has a definite stratospheric character. During summer, however, air of tropospheric character can often be found in the lowermost stratosphere.

Just as the lowermost stratosphere is a distinct region of the stratosphere, so too is the upper troposphere a distinct region of the troposphere. The distinction in both cases has to do with transport. However, in this case the defining characteristic is not so much the possibility of rapid connection with the stratosphere, but rather the low likelihood of rapid connection with the boundary layer. As noted earlier, tropical convection establishes a state of moist convective adjustment up to about 11−12 km. This is the height of the maximum Hadley-cell outflow and is quite distinct from the tropical tropopause which is located at around 16−17 km. Realistic models of radiative−convective adjustment exhibit a similar structure, with CO_2 long-wave heating leading to a clear separation between the region of convective adjustment and the temperature minimum[33] (see Figure 2). Between these two levels, there is a rapidly decreasing (with height) likelihood of convection and a rapidly increasing stratospherically driven upwelling. In clear-sky regions, the transition from downwelling to upwelling occurs in the range 14−16 km.

There has been considerable recent interest in this transition region, known as the "tropical tropopause layer" (TTL). Together with the lowermost stratosphere, the region is referred to as the "upper troposphere/lower stratosphere" (UTLS). The TTL is of great chemical interest because air within it can be chemically aged. Both radiatively and chemically, it is in many ways like the stratosphere. However, convection does penetrate into the TTL and deliver boundary-layer air, and there is lateral mixing with subtropical air of stratospheric character. The balance between these mixing processes and photochemical aging sets the chemical boundary conditions for the stratosphere. Simple one-dimensional models based on convective outflow, clear-sky ascent, and chemical aging seem able to explain a considerable part of the observed structure in this region.[34,35] Ozone profiles within the tropics show a continual transition from low tropospheric ozone to high stratospheric ozone, with no particular signature at the tropical tropopause itself.

From a transport point of view then, one might regard the boundary condition for the stratosphere as being set at the level at which the clear-sky vertical motion is upward and modified by convective intrusions above that level. However, water vapor is special because it condenses and can sediment. While the thermodynamic effects of condensation are restricted to below about 10 km or so, from the chemical and radiative point of view the water vapor distribution in the TTL and the stratosphere is a key quantity. The classical view,[36] based on the *freeze-drying mechanism*, is that water vapor is reduced to the saturation mixing ratio (over ice) at the minimum temperature—which is to say at the tropical tropopause—and then provided the ice is somehow removed, the dehydrated air enters the warmer stratosphere where it remains unsaturated (except in cold wintertime polar conditions in the lower stratosphere). In this respect the tropical tropopause has a special significance as the cold point. This view is still broadly held today and is believed to account for the *tropical tape recorder*.[37] However, important aspects of exactly what determines the water vapor entering the stratosphere remain controversial. In particular, recent observations of increasing stratospheric water vapor are inconsistent with the classical view of dehydration, given that the tropical tropopause temperature has if anything decreased in recent decades.[38]

10. Chemistry−Climate Coupling

Chemistry affects climate through its effects on radiatively active species. These include important well-mixed greenhouse gases such as N_2O, CH_4, and halogens as well as the relatively short-lived greenhouse gases O_3 and H_2O. (Chemistry also affects

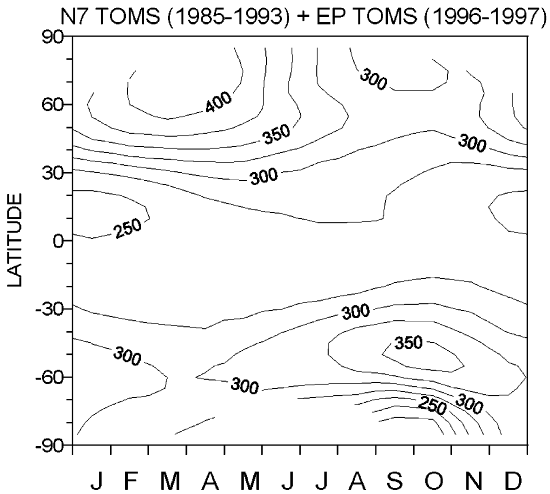

Figure 14. Seasonal cycle of total ozone as a function of latitude, based on TOMS satellite data covering the years 1985–1997, in Dobson units. Figure courtesy of Vitali Fioletov, Meteorological Service of Canada.

climate radiatively through aerosols and clouds, which are beyond the scope of this review.) For the long-lived greenhouse gases, changes in atmospheric temperature and circulation could alter their lifetime and, thereby, their global warming potential. However, the strongest chemistry–climate coupling occurs through O_3 and H_2O, because their spatial distributions are so inhomogeneous and thus sensitive to changes in atmospheric circulation and climate. The spatial inhomogeneity arises because of their relatively short lifetimes compared to the longer-lived species (see section 1.1) and is particularly notable in the UTLS region which, because of its low temperatures, plays an important role in the greenhouse effect.

It is perhaps arguable whether UTLS H_2O should be considered a chemical species because it is under such strong thermodynamic control in the troposphere. However, ozone is quite another matter. Unlike the well-mixed greenhouse gases, ozone is produced within the atmosphere rather than being emitted into it. The distribution of ozone accounts for the stratosphere (see Figure 2) and thus determines (with other factors) the minimum temperatures found at the tropopause. Ozone is also the classical example of stratospheric transport; ref 39 inferred a poleward circulation from the fact that the maximum total ozone concentration was found in the extratropics, not in the tropical ozone source region. The seasonal cycle of the observed total ozone distribution is shown in Figure 14. In the extratropics, ozone increases during the winter season and reaches a maximum during spring. The ozone buildup is the result of poleward and downward transport by the Brewer–Dobson circulation and shuts down once the winds become westward and stratospheric wave drag disappears. Thereafter ozone decays photochemically until the next year's buildup begins.[40] The springtime ozone maximum is larger in the Northern than in the Southern Hemisphere, because planetary-wave driving (and thus the Brewer–Dobson circulation) is strongest in the Northern Hemisphere. The maximum also reaches to the pole, whereas in the Southern Hemisphere it is prevented from reaching the pole by the mixing barrier associated with the strong Antarctic polar vortex. Of course, the low springtime Antarctic ozone values observed today are also due to the ozone hole, but minimum ozone was found over South Pole in springtime even before the ozone hole.[41]

The interhemispheric contrast in ozone illustrates the potential for chemistry–climate coupling because it shows the strong sensitivity of the ozone distribution to atmospheric circulation and climate. Because ozone has such a strong radiative effect, there is a feedback of ozone changes onto climate. The coupling between dynamics and ozone chemistry is nonlinear, with many feedbacks, and it is not possible to separate the two in an additive fashion.

A clear and comparatively simple case of chemistry–climate coupling occurs in the upper stratosphere, where transport of ozone is relatively unimportant (compared to the lower stratosphere) and the primary coupling between ozone chemistry and dynamics occurs through temperature. Anthropogenic increases of stratospheric chlorine have reduced upper stratospheric ozone abundance over the past few decades, which has cooled the upper stratosphere due to less absorption of solar radiation. At the same time, the increase of CO_2 has also cooled the upper stratosphere due to increased emission of thermal radiation. The most recent estimates suggest that the contribution of the two effects to the observed stratospheric cooling since the late 1970s is roughly equal.[42] However, the two effects are not really separable in this way, since the cooling due to the CO_2 increase has decreased ozone loss rates and thereby offset some of the ozone loss due to chlorine. Because of this CO_2-induced cooling, the ozone abundance in the upper stratosphere is expected to increase more rapidly in the future than would be predicted based on the reduction in chlorine alone.[42]

Chemistry–climate coupling is more complicated in the lower stratosphere, because of the crucial role of transport in establishing the ozone distribution. The interhemispheric contrast in the seasonal cycle of ozone (Figure 14), a consequence of the interhemispheric contrast in stratospheric dynamics, has already been noted. This interhemispheric contrast in stratospheric dynamics has important implications for understanding the behavior of polar ozone over the past few decades (Figure 15). In the Antarctic polar stratosphere, wintertime temperatures are nearly always low enough to form the polar stratospheric clouds (PSCs) that allow ozone-hole chemistry to occur.[44] Thus, the record of Antarctic springtime ozone roughly follows the halogen loading of the stratosphere—with the exception of 2002, when the Antarctic experienced an unprecedented stratospheric sudden warming, and 1988, when the vortex was severely disturbed but not quite to the point of a sudden warming. In the Arctic, in contrast, because of the stronger planetary wave drag and associated downwelling, temperatures are higher and in many winters never get low enough for extensive PSC formation. As seen in Figure 15, the year to year behavior of ozone in the Arctic is highly variable.

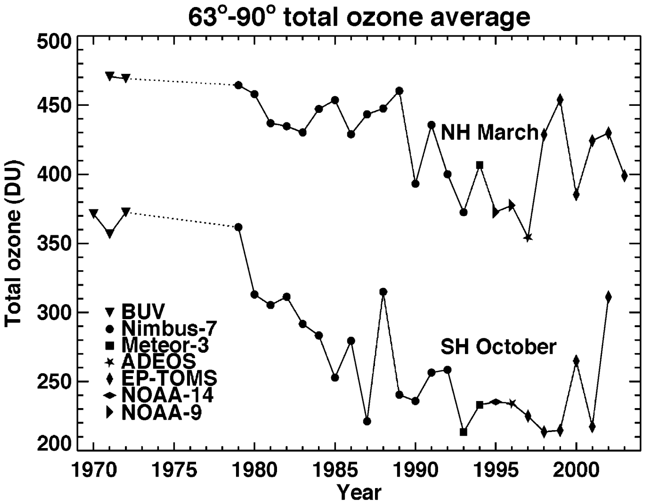

Figure 15. Total ozone poleward of 63° latitude in the springtime of each hemisphere (March for the Northern Hemisphere and October for the Southern Hemisphere), in Dobson units, based on data from various satellite instruments as indicated. This is an update of the figure given in ref 43. Figure courtesy of Paul Newman, NASA Goddard Space Flight Center.

When severe cold winters do occur, as, for example, in 1996 or 1997, then severe ozone depletion results, and when they do not, it does not. This means that the observed record of Arctic ozone does not follow the halogen loading of the stratosphere particularly well and reflects more the interannual meteorological variability. In fact, the low ozone years are the result of both chemistry and dynamics; when temperatures are low enough for PSC chemistry to proceed, this is the result of a weaker meridional circulation, which means less ozone transport. This contrast in ozone transport between warm and cold winters is analogous to that between the Arctic and the Antarctic itself. One can therefore think of halogen chemistry as an amplifier of dynamically induced polar ozone variability. The most recent estimates suggest that in the low ozone years of the 1990s, roughly one-half the ozone decrease in the Arctic vortex (compared to high ozone years) came from halogen chemistry and one-half from reduced transport.[42]

The fact that the development of the Antarctic ozone hole was such a direct response to halogen loading suggests that we may expect Antarctic ozone to recover as halogen levels decline in the future; changes in climate over the next 50 years seem unlikely, on present evidence, to alter the widespread formation of PSCs in the Antarctic winter. However, the situation in the Arctic is quite different. Arctic conditions are highly variable from year to year, and this variability straddles the temperature threshold for PSC formation and severe ozone depletion. Thus, just as the past record of Arctic ozone can only be interpreted in the context of dynamical conditions, the same applies to the future—yet we do not know how the future will evolve.

The difficulty here is 2-fold. The first problem is that of signal-to-noise, namely, the fact that the signal of halogen impact on Arctic ozone has to be discerned over the noise of internal climate variability. (In the Antarctic, the signal is stronger and the noise weaker.) The time scale over which Arctic "climate" can be defined—several decades—overlaps with the time scale over which halogen loading is changing, preventing a clear separation of the signal from the noise based on data alone. It is becoming increasingly clear that chemistry—climate model simulations of Arctic ozone need to consider an ensemble of possible behaviors in order to identify the impact of halogen loading, but the real atmosphere will only exhibit one realization.

The second problem is that, in contrast to the Antarctic, even a relatively small change in stratospheric climate could have a large impact on Arctic ozone. However, current predictions of Arctic climate change in the stratosphere are inconclusive. The wild card is the possible change in stratospheric planetary wave drag, which could easily trump the (relatively small) direct radiative response to greenhouse gas changes in the lower stratosphere, yet which is highly sensitive to various factors in climate and extremely difficult to predict. It is instructive to recall that in 1997 many were predicting severe Arctic ozone depletion in the future, comparable to that in the Antarctic, on the basis of the dramatic observed decreases seen in the 1990s up to 1997 (Figure 15) together with the prediction of a single chemistry—climate model.[45] The proposed mechanism was a greenhouse gas induced decrease in stratospheric planetary wave drag. Today, with the latest generation of chemistry climate models unable to reproduce those results[46]—indeed, several models suggest that stratospheric planetary wave drag will increase and the Arctic vortex will warm—and following a number of relatively warm years with only modest ozone decreases (Figure 15), there is no consensus on how climate change will affect Arctic ozone.[42] Yet whatever does happen, it can be safely said that changes in ozone chemistry and in climate will be closely linked.

11. Summary

Atmospheric dynamics affects atmospheric composition through both temperature and transport. Atmospheric composition feeds back on dynamics via radiatively active species. This is especially the case for those species with relatively short lifetimes (O_3 and H_2O) whose distribution is most sensitive to dynamics, but it applies also to well-mixed greenhouse gases with significant loss processes within the atmosphere (CH_4, N_2O, CFCs). Thus, atmospheric dynamics and chemistry are nonlinearly coupled. Natural (internal) variability is generally driven by dynamics but has chemical implications and is also modified by chemistry.

There is an important distinction, both chemically and dynamically, between the troposphere and the middle atmosphere (stratosphere and mesosphere). In the troposphere, the radiatively determined state is dynamically unstable to convective instability in the tropics and to baroclinic instability in the extratropics. The resulting thermally driven motions lead to mixing time scales that are relatively rapid in the troposphere—typically, air parcels in the free atmosphere will have been in contact with the boundary layer within the last week or so. In addition, the troposphere is characterized by clouds and rainout of soluble species. The middle atmosphere, in contrast, is dynamically stable and its weak meridional circulation is driven mechanically by waves propagating up from the troposphere. Air within the middle atmosphere can have a residence time of several years and can be photochemically aged. Yet the distinction between troposphere and stratosphere is not rigid, and one of the most fascinating regions of the atmosphere—both chemically and dynamically—is the upper troposphere/lower stratosphere (UTLS), which shares properties of both regions. In particular, the tropospheric part of the UTLS, known as the tropical tropopause layer, represents a significant portion of the mass of the free troposphere and plays a major role in the tropospheric ozone budget and, hence, in determining the oxidizing capacity of the troposphere.

Ozone represents the classic example of a chemical species strongly influenced by dynamics, and the understanding of past ozone changes due to halogens requires an appreciation of atmospheric dynamics. Given the central role of ozone in climate—it is a greenhouse gas which is produced entirely within the atmosphere, but it also affects the lifetimes of other greenhouse gases through its control of oxidation processes—future changes in climate will involve fascinating chemical–dynamical interactions which are only beginning to be elucidated.

12. Glossary

Adiabatic: Involving no heat transfer from or to an air parcel (consisting of a given set of molecules). For adiabatic motion, the first law of thermodynamics states that the entropy of an air parcel is conserved as the parcel moves around; this is expressed by eq 2 with $Q = 0$, which is nothing more than the familiar $dU + p dV = 0$ where $U = c_v T$ is the specific internal energy of dry air, c_v is the specific heat at constant volume, and $V = 1/\rho$ is the specific volume. (To see this, $dV = -(d\rho)/\rho^2$, the mass continuity equation can be written $d\rho = -\rho\nabla\cdot\mathbf{v}$, and d is identified with $\partial/\partial t + \mathbf{v}\cdot\nabla$ which is the time rate of change following an air parcel.) Adiabatic motion conserves the *potential temperature* of the air parcel; thus, when the atmosphere is stably stratified, adiabatic motion is confined to quasi-horizontal isentropic surfaces of constant potential temperature (although these surfaces themselves move).

Balanced Dynamics: The theoretical description of the slow, vortical motion in the atmosphere. In practice, "slow" means time scales of a day or more—which is to say, weather systems and climate. Balanced dynamics is described by *PV invertibility* and excludes faster motions that are not constrained by PV, such as internal gravity waves and convection. There are various theoretical models of balanced dynamics, but the most widely known is quasi-geostrophic (QG) theory. [The thermal wind relation (eq 6) is a special case of QG theory.] QG theory is not quantitatively accurate, but it does capture a great deal of the essence of atmospheric dynamical phenomena;[9] in fact, most detailed theoretical understanding of large-scale extratropical atmospheric dynamics is based on QG theory. However, it is important to know that the general principles of QG dynamics can be expressed in terms of PV and can thus be generalized to more accurate models of balanced dynamics.

Brewer–Dobson Circulation: The chemical transport circulation of the stratosphere. Reference 36 inferred a tropical entry to the stratosphere based on the water vapor distribution (see *freeze-drying mechanism*), while ref 39 inferred poleward transport within the stratosphere based on the ozone distribution. Both characteristics match the *diabatic and residual circulations*, so there is a tendency to use all three terms synonymously. However, chemical transport involves both the mean mass circulation and two-way mixing (i.e., without net transport of mass), and only the former is related to the diabatic or residual circulation. Thus, for example, air in the extratropical lower stratosphere is not as *chemically aged* as it would be if it were advected purely by the diabatic or residual circulation.[47]

Chaotic: Deterministic yet ultimately unpredictable. A deterministic system has a unique solution for a given initial condition and is to be distinguished from a stochastic system which has some externally imposed random element. Yet deterministic systems can exhibit "sensitive dependence on initial conditions" (SDIC), meaning that small differences in the initial conditions of two solutions will generally amplify until after some time the two solutions are as different as any two states of the system chosen at random. This is most famously illustrated by E. N. Lorenz's "butterfly effect"[48] and is understood to be the reason for the ultimate futility of weather prediction beyond a certain time limit generally regarded as being about 2 weeks. Chaotic solutions cannot be steady, or periodic, or even quasi-periodic (meaning that they represent a superposition of a finite number of periodic solutions with periods that are not all rational multiples of each other, with the consequence that the same state never repeats itself) for such solutions do not exhibit SDIC and are predictable arbitrarily far into the future. Chaotic systems are predictable for some finite time but are ultimately unpredictable.

Chemical Aging: The persistent conversion of one chemical species to another by photochemistry (chemistry associated with the absorption of ultraviolet and visible radiation) or oxidation. One might reasonably include radioactive decay as well. Air that enters the stratosphere through the tropical tropopause tends to remain in the stratosphere for a considerable time. Because all air in the deep stratosphere (above the lowermost stratosphere) has to have passed through the tropical tropopause, one may define the "age" of stratospheric air, at a given location, as the time since it passed through the tropical tropopause. In fact there is a *PDF* of ages, because of mixing;[47] a given air parcel is composed of molecules with different ages. The mean of this PDF is the mean age (or simply age, for short). In the upper stratosphere, the age of air is several years. Photochemical and oxidation processes can therefore act on stratospheric air over long time scales, and the age of air is inversely related to the concentrations of long-lived species whose primary loss process occurs in the middle atmosphere. In fact, the spatial distributions of such species and of age tend to coincide.

Diabatic: Involving heat transfer from or to an air parcel (consisting of a given set of molecules). Diabatic processes in the atmosphere include heat production by molecular viscosity, molecular diffusion of heat, chemical heating, fluxes of sensible or latent heat from the Earth's surface, and radiative heating or cooling. Latent heat release or uptake is fundamentally an *adiabatic* process (in thermodynamic equilibrium) and conserves entropy but is sometimes represented as a diabatic process in simple models of dry dynamics.

Diabatic Circulation: The mean meridional circulation inferred from the steady-state thermodynamic equation in isentropic coordinates, where eddy fluxes and meridional advection of heat vanish by construction.[49] There is then a balance between adiabatic heating or cooling from vertical flow and diabatic cooling or heating, which in the stratosphere is mainly from radiative processes. The associated meridional flow is inferred from mass conservation.

Downward Control: The constraint between *wave drag* and the *residual (or TEM) circulation* under steady-state or time-mean conditions.[19] Under these conditions, the TEM formulation of the zonal momentum (or angular momentum) equation represents a balance between the applied torque from wave drag or frictional drag and the Coriolis torque from meridional flow. The associated vertical flow is inferred from mass conservation. The key assumptions of downward control are that radiative damping can accommodate any steady vertical flow and that frictional drag is restricted to the planetary boundary layer where it can accommodate whatever steady meridional flow is demanded by mass conservation. It then follows that the vertical flow at any given altitude and latitude is determined by the total amount of wave drag above that location. A corollary of downward control is that a radiative heating anomaly cannot, on its own, lead to a steady meridional circulation.

Freeze-Drying Mechanism: The process by which air is conventionally imagined to be dehydrated as it enters the stratosphere in the tropics. Because the tropical tropopause is a local minimum of temperature, it also represents a minimum in the saturation mixing ratio of water vapor over water or ice (it turns out that ice is the relevant factor here). Thus, the water vapor in a parcel ascending through the tropical tropopause region will continuously condense into ice clouds. Provided the ice is somehow removed, the dehydrated air will then enter the warmer stratosphere and remain undersaturated. Water vapor is produced within the stratosphere by methane oxidation, but the quantity $H_2O + 2CH_4$ (which is conserved under methane oxidation) is remarkably uniform throughout the stratosphere. The only exception occurs in the Antarctic lower stratosphere in winter and in unusually cold winters in the Arctic, where the extremely low temperatures cause further condensation and dehydration through sedimentation of polar stratospheric clouds.

Hydrostatic Relation: The relation between pressure and density for a fluid at rest, in gravitational equilibrium. For a shallow atmosphere with constant gravitational acceleration g, the balance of forces between the vertical pressure gradient and gravity leads to the relation $\partial p/\partial z = -\rho g$, where p is pressure, z is altitude, and ρ is density. This equation is the differential form of the statement that the pressure at any altitude is equal to the weight of air (per unit area) above that altitude. The hydrostatic relation is exact for a fluid at rest but is also the dominant balance in the vertical momentum equation (and thus an excellent approximation) for large-scale atmospheric motion. In fact, it is a good approximation provided the vertical length scale of the motion is much smaller than the horizontal length scale, which is true for horizontal length scales larger than a few tens of kilometers. The form of the governing equations used in weather prediction models and in climate models is the so-called "primitive equations", in which the hydrostatic relation is imposed in place of the vertical momentum equation.

Inertial Instability: A fluid dynamical instability associated with rotation about the Earth's axis in the absolute frame. (It has a laboratory manifestation in Taylor–Couette flow.) A steady zonal flow satisfies geostrophic balance (eq 3). If an air parcel is displaced toward the equator while conserving its angular momentum (in the absolute frame), then the zonal wind u has to decrease since the absolute zonal velocity associated with the Earth's rotation increases; the same physics explains the motion of a rolling ball on a rotating turntable. However, the new value of u may yet have a stronger Coriolis force than the ambient meridional pressure gradient, in which case there is a positive feedback and the motion is unstable. This condition is known as "inertial instability" and is satisfied whenever the maximum angular momentum occurs away from the equator. (When $u = 0$—solid-body rotation—the angular momentum maximum is at the equator.) Inertial stability requires that air in the Northern Hemisphere has positive potential vorticity (PV) while air in the Southern Hemisphere has negative PV. Yet air parcels conserve their PV under adiabatic evolution. Thus, whenever air parcels cross the equator on sufficiently fast time scales for the motion to be adiabatic, inertial instability develops and breaks up the motion. This dynamical constraint severely restricts cross-equatorial flow.

Lapse Rate: The rate of decrease of atmospheric temperature with altitude. (Increasing temperature with altitude, as in the stratosphere, corresponds to a negative lapse rate.) The "dry adiabatic lapse rate" is the lapse rate that corresponds to that of a parcel of dry air undergoing adiabatic ascent or descent. Thus, an environmental lapse rate equal to the dry adiabatic lapse rate corresponds to a state of uniform dry entropy (or equivalently uniform potential temperature) within which a parcel of dry air has neutral (zero) buoyancy. An environmental lapse rate greater than the dry adiabatic lapse rate is convectively unstable; if an air parcel is displaced up a small distance by atmospheric motion, it will be warmer than its environment and thus positively buoyant and thus will continue to rise. The extent to which the environmental lapse rate is less than the dry adiabatic lapse rate determines the *static stability* of the atmosphere. For moist air, one can similarly define a "moist adiabatic lapse rate" corresponding to that of a saturated parcel of air undergoing adiabatic ascent or descent (evaporation is assumed on descent to maintain saturation), although there is no simple expression for its value.

Lapse-Rate Tropopause: The altitude above which the lapse rate decreases below 2 K/km through a layer of at least 2 km depth. This is the official definition of the tropopause according to the World Meteorological Organization and is that generally used by meteorologists; it is often called the "thermal tropopause". The thermal tropopause is generally—though not always—located close to a temperature minimum in the vertical profile, and therefore, the temperature minimum is also often considered to be the tropopause, known as the "cold point tropopause".

Meridional Circulation: Circulation along a meridional plane, that is to say the latitudinal–vertical plane lying along a meridian. Thus, "meridional circulation" is really shorthand for meridional–vertical circulation. By default it refers to the zonal mean of the circulation. There are various definitions of the meridional circulation used by dynamicists, but the usual definition is the *residual (or TEM) circulation*, which is closely related to the *diabatic circulation*. It can be thought of as the mean mass circulation.

Moist Entropy: The entropy of a mixed system consisting of dry air plus water. The water can be in either the vapor, liquid, or ice phase, and the entropy of the mixed system needs to include the contribution from latent heat associated with the phase transitions of water. The moist entropy of a parcel of air is conserved under adiabatic motion if the condensate is kept with the air parcel. The equivalent *potential temperature* is a function of moist entropy.

PDF (Probability Distribution Function): The function describing the frequency distribution of the possible values of a given quantity, normalized such that its integral is unity. For a geophysical quantity (e.g., concentration of a chemical species), one may construct the PDF at a given altitude and latitude by binning all the available measurements at different longitudes and times and constructing a histogram of the concentrations: the value of the PDF at each concentration value is the likelihood of observing that concentration value in a random measurement (at the given altitude and latitude).

Potential Temperature: The temperature that a parcel of dry air would have if displaced *adiabatically* to a given reference pressure p_{00}, normally chosen to be 1000 hPa. Thus, potential temperature is conserved under adiabatic motion and is a function of entropy. For an ideal gas, potential temperature $\theta = T(p/p_{00})^{-R/c_p}$, where R is the gas constant of air and c_p is the specific heat at constant pressure. For an adiabatic process

$$\begin{aligned} d\theta &= \frac{\theta}{T}\,dT - \frac{R}{c_p}\frac{\theta}{p}\,dp = \frac{c_v}{c_p}\frac{\theta}{T}\,dT + \frac{R}{c_p}\,\theta\rho d\left(\frac{1}{\rho}\right) \\ &= \frac{\theta}{c_p T}(dU + p\,dV) = 0 \end{aligned}$$

confirming that θ is indeed a function of entropy. Here $U = c_v T$ is the specific internal energy of dry air, c_v is the specific heat at constant volume, and $V = 1/\rho$ is the specific volume. The second equality follows from the equation of state $p = \rho RT$ and the relation $c_p = c_v + R$, and the final equality is the first law of thermodynamics. If $\partial\theta/\partial z = 0$, where z is altitude, then the atmosphere is convectively neutral to dry ascent or descent; if $\partial\theta/\partial z < 0$, then it is convectively unstable. Thus, a state with $\partial\theta/\partial z < 0$ cannot be maintained and will spontaneously convect. Convective overturning and mixing will lead to a state of uniform entropy with $\partial\theta/\partial z = 0$, a process known as convective adjustment. The depth of convective adjustment is constrained by conservation of total energy. For a moist atmosphere, one can account for the entropy associated with the phase change of water, in which case θ is replaced by the so-called equivalent potential temperature θ_e.[50]

Potential Vorticity: The generalization of the *vorticity* of a two-dimensional incompressible fluid to a rotating, stratified fluid. Stratification introduces quasi-horizontal isentropic surfaces to which adiabatic motion is confined. This motion can be characterized by the vertical component of vorticity—a scalar. Kelvin's circulation theorem (see *vorticity*) applies in the absolute reference frame, so one must include the vertical component of the vorticity associated with the Earth's rotation. The latter is twice the angular frequency of the Earth's rotation and points along the rotation axis, so its vertical component is (by simple geometry) this quantity multiplied by the sine of latitude, which is just the Coriolis parameter f. Thus, the vorticity associated with the Earth's rotation can be seen as a potential form of so-called "relative vorticity" (namely, that associated with the fluid motion in the rotating frame), with southward motion (in both hemispheres) leading to an increase in relative vorticity. The motion of the isentropic surfaces themselves allows a weak three-dimensional component to the fluid motion in physical space, so fluid elements may expand or contract in the vertical direction, leading to a weak form of vortex stretching. Thus, *static stability* can also be seen as a potential form of relative vorticity, with weakened stratification (for a given value of PV) leading to an increase in relative vorticity.

Potential Vorticity (PV) Invertibility: The concept that the dynamical fields can be obtained from the distribution of PV on isentropic surfaces (together with the mass distribution). In general, a three-dimensional velocity field cannot be characterized by a single scalar field. However, in the atmosphere, stratification and rotation together approximately confine the motion to isentropic surfaces and provided the motion is sufficiently slow compressibility effects are weak. Under these conditions, known as *balanced dynamics*, the velocity field is determined by the PV[51,52] and one can view the dynamics in terms of the evolution of the PV field on isentropic surfaces.[10] This is a generalization of the concept that for two-dimensional incompressible flow, the velocity vector is determined by the scalar vorticity. In fact, not only the velocity field but also the temperature and pressure fields are determined by the PV. It is for this reason that atmospheric dynamicists find PV to be such a valuable diagnostic and theoretical construct. Indeed, virtually every theory of large-scale atmospheric dynamics can be cast in terms of PV. Note that surface temperature gradients are equivalent to a "surface" PV and must be included in the concept of PV invertibility.

Pressure Coordinates: The coordinate system obtained when the vertical coordinate is changed from geometric height z to pressure p. This transformation is appropriate when the *hydrostatic relation* holds, in which case the transformation from z to p is well-defined (i.e., monotonic). Pressure coordinates are essentially mass coordinates and therefore have a number of advantages over geometric coordinates. In the momentum equation, the horizontal pressure gradient in height coordinates transforms to a horizontal gradient of geopotential height $\Phi \equiv gz$ in pressure coordinates. This can be seen as follows. Consider a function $f(x, z)$ where x is a horizontal coordinate. We can also regard f as a function of x and p. To determine how the partial derivative of f with respect to x transforms between z and p coordinates, we may write an arbitrary variation in f as $\delta f = f_{x,p}\delta x + f_{p,x}\delta p$, where the notation $f_{x,p}$ denotes the partial derivative of f with respect to x, at constant p. However, we may equally regard f as a function of x and z and write $\delta f = f_{x,z}\delta x + f_{z,x}\delta z$. Now regarding z as a function of x and p, it follows that $\delta z = z_{x,p}\delta x + z_{p,x}\delta p$, whence $\delta f = f_{x,z}\delta x + f_{z,x}(z_{x,p}\delta x + z_{p,x}\delta p)$. Comparing this last expression with the first expression for δf, the coefficients of δx and δp must be the same since δx and δp are arbitrary. Therefore, $f_{x,p} = f_{x,z} + f_{z,x}z_{x,p}$. Now consider the special case $f = p$, for which the left-hand side of the last equation is zero by definition. Using the hydrostatic relation $p_{z,x} = -\rho g$, it follows that $p_{x,z} = \rho\Phi_{x,p}$.

Residual (or TEM) Circulation: The mean meridional circulation in the transformed Eulerian mean (TEM) formulation of the governing equations, denoted $(\bar{v}^*, \bar{w}^*)$. The TEM formulation involves a transformation of the circulation variables that, to good approximation, collects all the eddy forcing terms in the zonal momentum equation, leaving only diabatic heating and mean advection in the thermodynamic equation.[49] In that sense it can be considered a (log-)pressure coordinate approximation to the isentropic formulation of the equations. The residual circulation is not quite equal to the (transient) diabatic circulation, but the differences between them are not quantitatively significant. The great merit of the TEM form of the equations is that the eddy forcing term in the zonal momentum equation is the divergence of the flux of a conserved measure of wave activity, the "Eliassen−Palm (EP) flux", written $\nabla\cdot\mathbf{F}$. It turns out that the EP flux divergence represents the convergence of angular momentum flux, when properly formulated in terms of angular pseudomomentum[49] (see also *wave drag*). Thus, the TEM equations concisely express the physical relation between angular momentum transfer by waves and the induced meridional circulation. In contrast, nonbreaking waves cannot transfer heat across isentropic surfaces, and so it is appropriate that the wave forcing terms disappear from the TEM thermodynamic equation.

Static Stability: The extent to which the environmental lapse rate of the atmosphere is less than the adiabatic lapse rate (dry or moist, depending on the context). An environmental lapse rate less than the adiabatic lapse rate is convectively stable; if an air parcel is displaced up a small distance by atmospheric motion, it will be colder than its environment and thus negatively buoyant and will sink back down. Similarly, an air parcel that is displaced down a small distance will be positively buoyant and will rise back up. This situation leads to oscillations, with a characteristic frequency given by the square root of $(g/\theta)\partial\theta/\partial z$, where θ is the *potential temperature*. (In the moist case, θ must be replaced with the equivalent potential temperature, θ_e.) Since θ is a function of entropy, static stability corresponds to the entropy increasing with altitude. The full three-dimensional wave motions for which static stability provides the restoring force are known as internal gravity waves.

Stratospheric Sudden Warming: Rapid warming of the wintertime polar stratosphere (over a few days) induced by the polar downwelling (and adiabatic warming) resulting from a focusing of planetary-wave drag over the pole. The phenomenon can also be understood in terms of a breakdown of the polar vortex through the absorption of negative angular momentum from the anomalous planetary-wave drag.[21] Sudden warmings appear to be essentially a chaotic process and are the primary cause of Arctic wintertime variability. In our present climate, planetary-wave drag is usually too weak to produce sudden warmings in the Antarctic, although 2002 was a notable exception.

Tropical Tape Recorder: The propagation of tropical tropopause water vapor values into the stratosphere. Persistent upwelling through the tropical tropopause, combined with the *freeze-drying mechanism*, implies that the saturation mixing ratio of water vapor at the tropical tropopause gets carried upward into the stratosphere. It follows that changes in the tropical tropopause temperature should change the amount of water vapor entering the stratosphere, with the signal propagating upward, much as a tape recorder marks a tape; the vertical profile of water vapor in the tropics then provides a record of the time history of the saturation mixing ratio at the tropopause. This effect is demonstrated most clearly in the annual cycle, where the annual cycle of tropical tropopause temperatures gets imprinted on the ascending water vapor signal.[37] In fact, water vapor is produced within the stratosphere by methane oxidation, so the tape recorder signal is best seen in $H_2O + 2CH_4$.

Vorticity: The vector given by the curl of the velocity field, which measures the "swirliness" of the flow, with the vorticity pointing along the axis of rotation according to the right-hand rule. Vortices—concentrated regions of vorticity—occur in everyday life, for example, at the edges of a canoe paddle being drawn through water or in water draining from a bathtub. In incompressible, inviscid, homogeneous fluid, vorticity vectors are carried with the fluid, a result known as Kelvin's circulation theorem. However, the magnitude of the vorticity vector can intensify through a process known as vortex stretching, as in the bathtub vortex. Since incompressible flow is nondivergent, the velocity field is uniquely determined by the vorticity field and one may view the dynamics solely in terms of vorticity. This is a special case of *potential vorticity (PV) invertibility* and is characteristic of *balanced dynamics*. For a two-dimensional fluid, the vorticity vector is necessarily normal to the plane of motion and there is no vortex stretching; in this case, "vorticity" refers to the magnitude of the vector (a scalar) and is conserved following fluid parcels.

Wave Drag: Angular momentum transfer by atmospheric waves, whether gravity waves, Rossby waves, or equatorial waves. It is essential to realize that angular momentum transfer by waves is not equal to the flux of wave angular momentum. Rather, one has to begin from the conservation law for angular momentum and derive what is known as the "angular pseudomomentum", representing the contribution by the waves.[53] In the atmosphere, wave drag is represented by the Eliassen−Palm (EP) flux divergence (see *residual circulation*), which is a combination of wave momentum and heat fluxes, even though it is angular momentum and not heat that is being transferred by the waves! (The heat-flux contribution to the EP flux occurs through the deformation of isentropic surfaces, much as topography deforms a flow, in a process known as "form drag".) The sign of wave drag is related to the phase speed of the waves relative to the mean flow, so planetary Rossby waves always give a negative drag, equatorial Kelvin waves always give a positive drag, and gravity waves can give a drag of either sign. Wave drag tends to act in a decelerative sense (hence the name), but not always, as for example with the quasi-biennial oscillation.[49]

Zonal Phase Speed: The speed at which the phase of a wave moves along latitude circles. The zonal propagation of a wave along latitude circles can be represented by the functional form $\psi(\lambda, t) = A \sin[k(\lambda - ct) + \varphi]$, where ψ is some field representative of the wave (for example, vorticity or temperature), λ is longitude, A is the amplitude of the wave, k is the *zonal wavenumber*, c is the zonal phase speed, and φ is some constant phase. The argument of the sine function is the phase of the wave; 0 and π are the nodes, $\pi/2$ is the maximum, and $3\pi/2$ is the minimum. As t increases, λ has to increase at c times the rate of increase of t in order to maintain the same phase. Thus, the wave phase propagates to the east (increasing λ) at the speed c. $c < 0$ corresponds to westward propagation.

Zonal Wavenumber: The number of wavelengths that fit around a latitude circle. Thus, zonal wavenumber 1 corresponds to a single wavelength (of 40 000 km at the equator), zonal wavenumber 2 corresponds to two wavelengths (each of 20 000 at the equator), and so on.

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