

DYNAMICAL PERTURBATIONS TO THE OZONE LAYER

Global-scale disturbances in the stratosphere produce substantial changes in the abundance and distribution of ozone and prevent the formation over the North Pole of an ozone hole like that observed over Antarctica.

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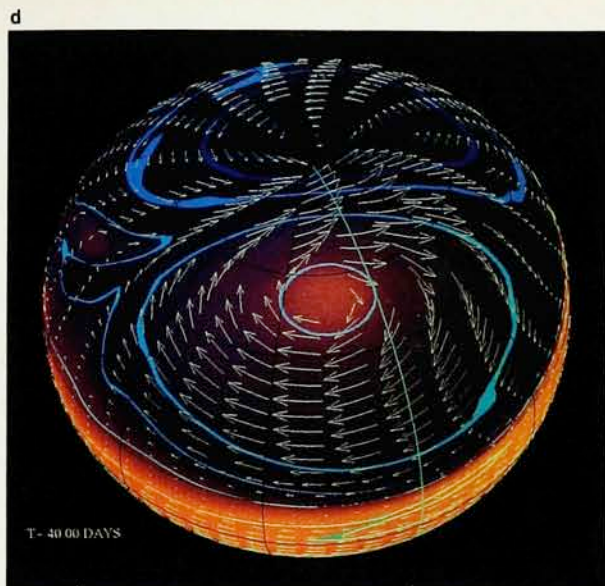
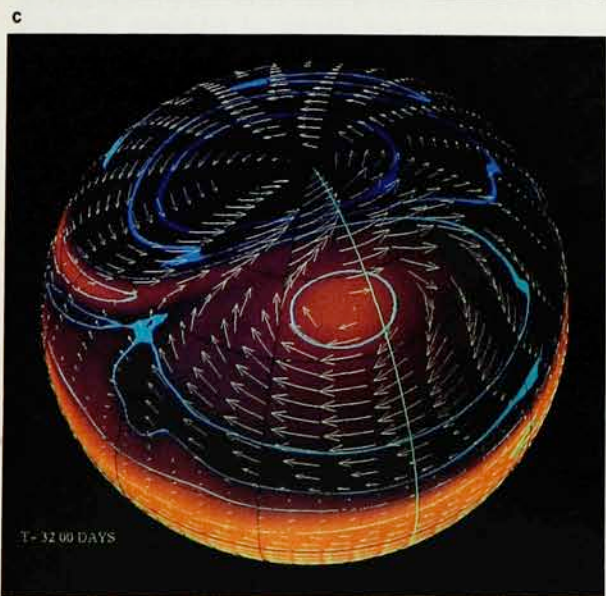
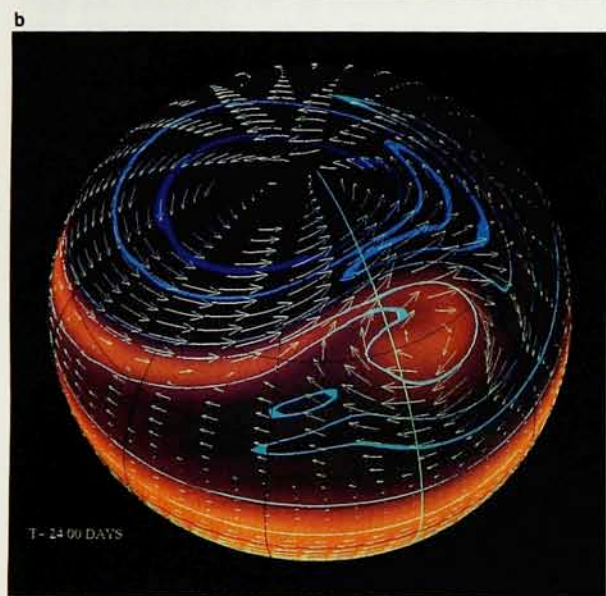
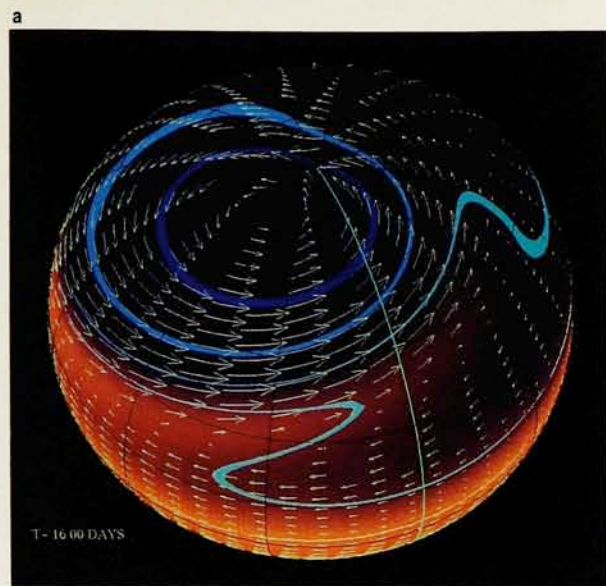
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Several years ago spectrophotometer measurements revealed substantial depletions of atmospheric ozone over Antarctica, reviving concerns over damage to Earth's ozone layer by human activities.¹ These ground-based observations were subsequently verified in global ozone data that had been collected by satellite for years.² The sizable ozone depletions over Antarctica are now believed to arise from complex chlorine chemistry, stimulated ultimately by the emission of man-made chlorofluorocarbons into the atmosphere.³

Ozone occurs naturally in the stratosphere, that layer of the atmosphere extending from about 10 km to 50 km above the surface of the Earth. During Antarctic spring, as much as half of the ozone over the South Pole disappears, with a nearly 100% loss in the layer between 10 and 20 km, where ozone is normally concentrated at these latitudes. Initially, several explanations were proposed for these dramatic reductions. However, recent expeditions to the Arctic and Antarctic have pointed to heterogeneous chemistry as the mechanism responsible.³ (See *PHYSICS TODAY*, July 1988, page 17.) Inside polar stratospheric clouds the presence of both solid and gas phases permits certain chemical reactions to proceed much faster than is possible in the gas phase. For this reason, much of the recent interest in ozone has focused on chemical issues. However, to assess man-made changes to date and to understand how the ozone layer may respond to future perturbations of human origin requires an understanding of how the ozone layer functions and the processes that contribute to its natural variability. In particular, anthropogenic variations must be considered in the context of naturally occurring changes in ozone.

An important source of natural variability in ozone is a class of disturbances known as planetary waves, which perturb the circulation of air in the stratosphere. (See figure 1.) These waves, which are possible in a rotating fluid, occur routinely in the northern winter stratosphere, where they disrupt the circumpolar flow and exert a profound influence on ozone and other chemical species. It now appears that these global-scale disturbances also play a key role in shaping ozone depletions at the poles.

This article presents an overview of these dynamical disturbances, focusing on how they alter the global distribution of ozone through horizontal and vertical transport. To set the stage, we review the basic climatology of ozone: its origin, global distribution and seasonal



Motion of polar air (blue) and tropical air (orange) after a disturbance to the circulation of the stratosphere, as given by a numerical calculation. Colors indicate values of potential vorticity, which is conserved for individual bodies of air and therefore traces out their motion (see figure 3). The color code ranges from orange, which indicates small values of potential vorticity and weak absolute angular momentum, to deep blue, which indicates large values of potential vorticity and strong absolute angular momentum. The sequence shows tropical air being drawn poleward in the form of a tongue of low potential vorticity. The folding of air contours, which develops in the first panel, is dynamically unstable and leads to the formation of a large anticyclone along the periphery of the vortex. This planetary-scale eddy draws midlatitude air from the vortex and entrains it with tropical air, stirring bodies of air down to small dimensions. Air in the resulting anticyclone has characteristics similar to air found at much lower latitudes in the undisturbed circulation. **Figure 1**

variation. This is followed by a description of the essential mechanics of planetary waves. We then use detailed transport calculations to investigate how these disturbances can influence the abundance and distribution of ozone.

Climatology of stratospheric ozone

For some time ozone has been recognized as essential to the existence of life at the Earth's surface.⁴ In addition to its biological importance, ozone plays a central role in the energetics and dynamics of the upper atmosphere. The absorption of ultraviolet solar radiation by ozone is a heat source for the tropical stratosphere, ultimately driving the global circulation of the upper atmosphere.⁵

Ozone, O_3 , is produced in the stratosphere by a series of reactions that begin with the photodissociation of diatomic oxygen. Ozone can be destroyed by recombination with atomic oxygen and by a variety of catalytic cycles involving hydrogen, nitrogen and chlorine. In the absence of atmospheric motion, the distribution of ozone would be determined by a balance between the chemical processes that create ozone and those that destroy it. On purely

chemical grounds, one would anticipate ozone to be most abundant in the tropics, where the photodissociation of O_2 is efficient. However, observations of stratospheric ozone reveal a very different picture.

The ozone abundance at any location over the Earth's surface is measured by the "total ozone column," which is the integral over the depth of the atmosphere of the amount of ozone per unit volume, or ozone number density ρ_{O_3} . The total ozone column is often expressed as the depth that the ozone alone would occupy at standard temperature and pressure; this depth is often measured in thousandths of a centimeter, or Dobson units, DU. At middle and high latitudes, the ozone column is concentrated between 10 and 20 km, where the ozone number density increases sharply with altitude before decreasing exponentially with the air density. In the tropics, ozone is concentrated at slightly higher altitudes, 20–30 km, but still within a layer about 10 km deep.

Figure 2, which is based on the climatological record of ozone, shows the seasonal variation of the total ozone column as a function of latitude.⁶ Column abundances of 300 DU, corresponding to a pure-ozone column depth of just 0.3 cm, are typical. Despite the fact that ozone is produced at low latitudes, the largest column abundances actually appear at high latitudes, as a result of poleward transport out of the tropical source region. In addition, column amounts are enhanced to over 400 DU in winter and spring, when large-scale disturbances of the circulation are prevalent. Abundances even greater than the monthly-mean values shown in figure 2 are observed intermittently.

The chemical lifetime of ozone, which measures how quickly its concentration adjusts to changes in the photochemical environment, is determined by the reactions involved in its production and destruction. In the lower stratosphere, where ozone is concentrated, its lifetime is several weeks, long enough for it to be transported passively, or "advected," by the circulation of the atmosphere.

The stratospheric circulation. On average, stratospheric air moves parallel to latitude circles and counter-

clockwise about the North Pole, or "cyclonically." This strong circumpolar flow, known as the polar night vortex, is brought about by the Earth's rotation when poleward-moving air parcels are deflected toward the east by the Coriolis force. In steering the airstream parallel to latitude circles, the Earth's rotation inhibits the transfer of ozone out of its chemical source region near the equator. The strong zonal flow tends to homogenize constituents along latitude circles, but allows only a very gradual transfer of species between the equator and the poles.

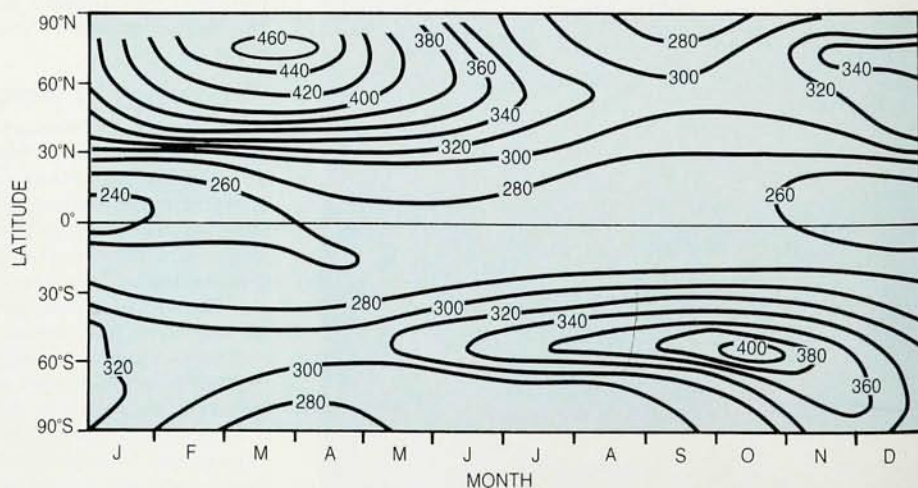
Historically, the stratospheric circulation has been viewed in terms of the overturning of air in the latitude-height plane, averaged over longitude.⁵ This "zonal-mean meridional circulation" is characterized by rising motion in the tropics and sinking motion at middle and high latitudes and operates on a time scale of months. The gradual overturning of air in the meridional plane draws ozone poleward from its chemical source region near the equator and redistributes it globally.

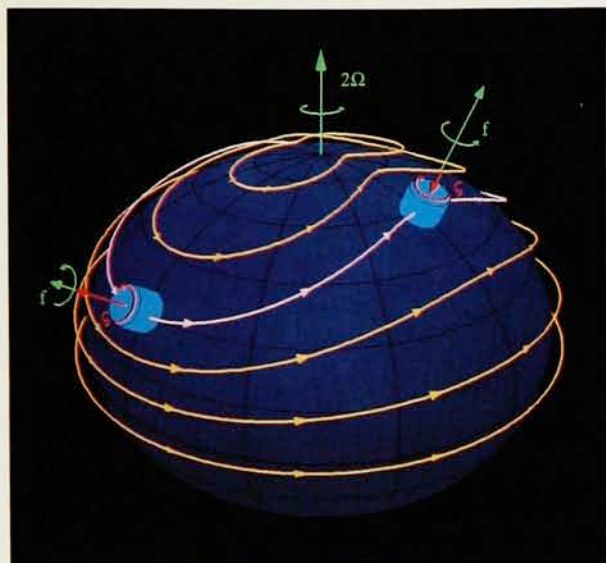
Although it provides a simple and conceptually appealing picture of stratospheric air motion, much of this zonal-mean overturning is actually the net result of motions that are not zonally symmetric but rather are more complex. Theoretical advances and large-scale numerical modeling have underscored the intimate relationship between the zonally averaged view of stratospheric air motions and large-scale disturbances to the circulation (see reference 5 and references therein). These disturbances, which form the subject of this article, operate on much shorter time scales than the zonally averaged circulation. They give rise to air motions that are variable with longitude but that when averaged around latitude circles produce the classical picture of the mean meridional circulation in the stratosphere. The stronger and more frequent these disturbances are, the more rapid is the zonally averaged overturning of air in the latitude-height plane and the more quickly ozone is exported out of its tropical source region.

As the distribution of solar radiation changes during the year, chemical and dynamical processes alter the abundance and distribution of ozone. Figure 2 indicates

Seasonal march of ozone abundance. The average column abundance of ozone—the vertically integrated ozone density—is plotted here as a function of latitude and time. (The abundance values are given here in Dobson units.) Although ozone is produced at low latitudes, the greatest column abundances are actually found at middle and high latitudes. Strong seasonal and intraseasonal variations in ozone abundance also occur at middle and high latitudes during winter and spring.

(Adapted from ref. 6.) **Figure 2**





Conservation of angular momentum of bodies of air moving through the planetary wave pattern. Represented here are elements of the potential vorticity $(f + \zeta)/h$, which measures the absolute angular momentum of a body of air. To leading order, the planetary vorticity f , which is twice the local vertical component of angular velocity, is exchanged with the relative vorticity ζ of the atmospheric flow. Air parcels displaced meridionally react to changes in planetary vorticity by spinning up cyclonically or anticyclonically, deflecting the airstream back and forth across an equilibrium latitude. **Figure 3**

only a weak seasonal variation in column abundance at low latitudes. However, at middle and high latitudes the seasonal variation is considerable. Maximum column amounts appear in spring in both hemispheres, the northern maximum occurring in early spring, whereas the southern maximum is delayed until late austral spring. Except during southern spring, when ozone is depleted in the Antarctic stratosphere, chemical processes do not differ appreciably between the hemispheres. Thus differences in ozone abundance must follow from differences in the dynamics of the Northern and Southern Hemispheres.

It is well known that the Antarctic vortex is stronger and less disturbed than its counterpart over the Arctic. The more organized circumpolar flow in the Southern Hemisphere is associated with a deep temperature depression over the South Pole. Very cold temperatures there are responsible for the formation of polar stratospheric clouds over the Antarctic, which are believed to have played a central role in the ozone depletions observed in recent years. The much warmer temperatures characteristic of the Arctic stratosphere do not support the formation of polar stratospheric clouds as readily and therefore inhibit the destruction of ozone through heterogeneous chemistry. It turns out that planetary waves are the key to these hemispheric differences. By disrupting the polar night vortex, these disturbances are responsible for the zonally averaged transfer of ozone out of the tropics. As we shall see, large-scale wave motions also affect the ozone distribution in a more direct and dramatic fashion.

Stratospheric disturbances

During winter and spring, the circumpolar flow is disrupted by global-scale, or planetary, waves that propagate upward out of the troposphere. Planetary waves are large-scale Rossby waves, a type of wave motion possible in a rotating fluid medium. When they amplify in the stratosphere, planetary waves displace the vortex out of polar symmetry, causing air to flow across latitude circles and allowing chemical constituents to be transferred meridionally on a time scale on the order of only a day.

When the planetary wave field amplifies, as occurs sporadically during winter, the strong zonal-mean flow (the flow averaged around latitude circles) is decelerated and temperatures over the polar cap increase. In its

extreme form, this disturbance is known as a "stratospheric sudden warming" because its effects were first observed as a reversal of the normal pole-to-equator temperature gradient.⁷ During a sudden warming, stratospheric temperatures in the polar night increase dramatically—by as much as 50 K in just a few days—actually becoming higher than those in the sunlit tropics. At the same time, the zonal-mean flow reverses direction, and ozone increases sharply at high latitudes.⁸ Although stratospheric warmings are often described in terms of zonally averaged quantities, they actually involve a complex distortion of the circulation in which air and chemical constituents are exchanged rapidly across the entire winter hemisphere.

Propagation of planetary waves. Planetary waves are excited at the Earth's surface by the forced ascent of air over elevated terrain, such as the Tibetan plateau and major mountain ranges, and also by convective heating in the tropical troposphere, where large quantities of latent heat are released in cumulus cloud systems.

The restoring force for these disturbances is supplied by the variation of the Coriolis force with latitude. Air parcels flowing meridionally experience a change in rotating environment reflected in the Coriolis parameter

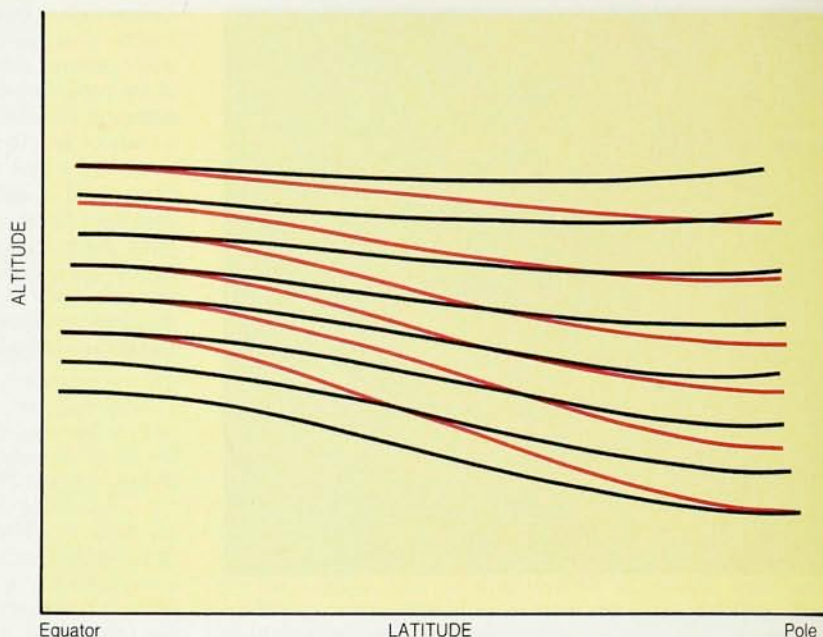
$$f = 2\Omega \sin \phi \quad (1)$$

Here Ω is the angular velocity of the Earth and ϕ is latitude. The Coriolis parameter f is the local vertical component of planetary vorticity (twice the local vertical component of angular velocity), varying from 0 at the equator to $\pm 2\Omega$ over the poles. Air parcels deflected across latitude circles adjust to changes in local rotation so as to preserve absolute angular momentum, that is, the angular momentum seen by an observer in an inertial reference frame. The latter is measured by the "potential vorticity,"

$$Q = \frac{f + \zeta}{h} \quad (2)$$

The quantity ζ is the relative vorticity of the flow, $\nabla \times \mathbf{v}$, as seen by an observer fixed with respect to the Earth, and h is the effective thickness of the fluid column. To leading order, the absolute vorticity $f + \zeta$ is constant, the relative vorticity ζ simply being exchanged with the planetary vorticity f . Thus air streaming toward the equator will experience decreasing planetary vorticity and spin up

Equivalent barotropic system defined above the 350-K isentropic surface, which is near the tropopause. Most of the ozone is concentrated in a 10-km layer just above the tropopause. Air moves along isentropic surfaces (solid lines), which slope downward toward the pole and intersect surfaces of constant ozone mixing ratio (colored lines). The lowest colored curve represents an ozone mixing ratio of 1 part per million. Because air density decreases exponentially with altitude, the ozone density falls off sharply above the lower boundary. **Figure 4**



cyclonically (increasing ζ) to compensate, while air flowing poleward will see increasing f and spin up anticyclonically (decreasing ζ).

Planetary waves owe their existence to the adjustment of air parcels to alternating changes in planetary vorticity. As they are deflected across latitude circles, air parcels experience a decrease or increase in planetary vorticity according to equation 1. To preserve potential vorticity (equation 2) they spin up cyclonically or anticyclonically, oscillating about an equilibrium latitude and giving the circulation a wave-like appearance (figure 3).

Once excited, planetary waves propagate both horizontally and vertically. It turns out that these waves are able to propagate only where the zonal flow is toward the east relative to the reference frame of the wave. Consequently, wave activity is excluded from the summer stratosphere, where winds are toward the west. However, wavelengths of planetary dimension are able to propagate freely into the eastward flow of the winter stratosphere, amplifying vertically as the air density decreases exponentially.⁹

Transport by planetary waves. Planetary waves disrupt the circumpolar flow of the winter stratosphere so that air streams meridionally. From a global perspective, the vortex is displaced from the pole and air flows across latitude circles. In this disturbed circulation atmospheric constituents are transferred meridionally two orders of magnitude faster than is possible in the unperturbed flow.

As an air parcel moves through the disturbed airstream, its dynamical characteristics change to preserve its potential vorticity. A property that is conserved for individual fluid particles is referred to as a "material tracer." Particular values of such a quantity trace out the motion of individual air parcels, and with them all other conserved species. In the lower stratosphere, radiative damping times are on the order of 50 days¹⁰ which is long enough (relative to the time scale for horizontal advection) for potential vorticity to behave as a tracer. Likewise, the long photochemical lifetime of ozone in the lower stratosphere means that its mixing ratio μ_{O_3} , which is the number of ozone molecules divided by the number of air molecules, also behaves as a tracer. Therefore the distributions of both the potential vorticity and the ozone

mixing ratio in the lower stratosphere are, to a good approximation, rearranged by air motions.

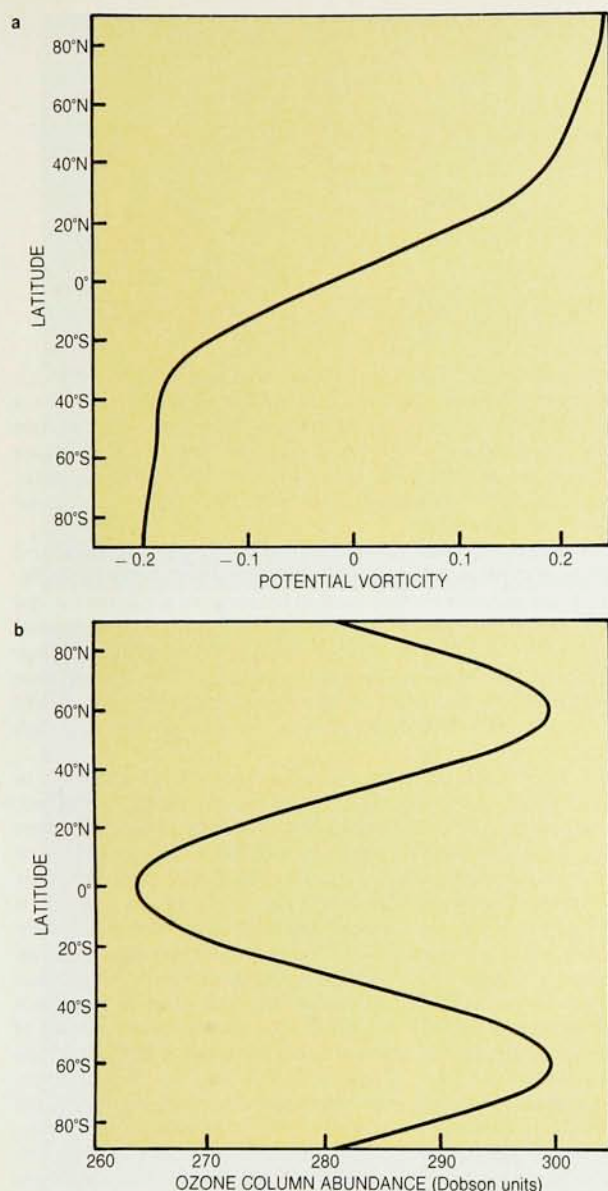
As long as dissipation is unimportant, the motion of air through the wavy stream pattern is completely reversible. Air parcels simply cycle back and forth across latitude circles with no net rearrangement. However, dissipation alters this picture by preventing air parcels from returning to their original latitudes, leading to a net redistribution of potential vorticity and other long-lived quantities. When the wave field amplifies, excursions of the airstream across latitude circles steepen to the point that the flow becomes dynamically unstable. The waves then "break,"¹¹ and secondary vortices develop through instability. As they amplify, these unstable eddies stir bodies of air down to small dimensions where nonconservative processes are efficient. Dissipation then results in an irreversible rearrangement of air and a net transport of constituents across latitude.

In addition to causing horizontal transport, planetary waves also induce vertical motion. Air moving vertically experiences changes in pressure and temperature through compression and expansion. If the time scale for heat transfer is long compared with that for advection, as is true in the lower stratosphere, the motion of an air parcel is nearly adiabatic. The pressure p and temperature T of a parcel then adjust to conserve its "potential temperature" θ :

$$\theta = T \left(\frac{p_0}{p} \right)^\kappa \quad (3)$$

The reference pressure p_0 is 1000 millibars. The exponent κ is the ratio of the gas constant to the specific heat at constant pressure; its value is 0.286 for air.

Like potential vorticity, potential temperature behaves approximately as a tracer in the lower stratosphere. It follows that air parcels must move along surfaces of constant potential temperature. These quasihorizontal surfaces, which are coincident with isentropic surfaces, undulate under the influence of planetary waves. Air descending along an isentropic surface responds to increases in pressure by undergoing compression, but with its temperature changing according to equation 3 to preserve its potential temperature. Actually, diabatic, or



Initial conditions in a numerical calculation of the circulation, before a disturbance is applied. Profiles of potential vorticity (a) and ozone column abundance (b), both zonally symmetric, are plotted as functions of latitudes. **Figure 5**

These dynamical influences have been explored through high-resolution transport calculations in which quasi-horizontal air motions are represented very accurately. Previous studies using equations similar to those that describe waves in shallow water have addressed horizontal transport in this manner.¹² However, vertical transport by planetary waves, which can be especially important for column abundances, is not described easily in this traditional framework.

The global transport calculations presented below are performed with the equivalent barotropic equations, which govern column-averaged motions and reflect behavior in a layer about 10 km deep.¹³ The ozone column abundance emerges naturally in this system when the three-dimensional equations of motion are integrated vertically over pressure. Strictly speaking, this system governs motion for which the vertical scale is large compared with the vertical scale of the mass distribution, as is typical in the stratosphere. One feature of the equivalent barotropic system is that it accounts for the vertical displacements associated with deep atmospheric motion.

An equivalent barotropic circulation, shown schematically in figure 4, is defined above a material surface that bounds the ozone layer from below. In this system the horizontal velocity $\mathbf{v}$ and ozone mixing ratio μ_{O_3} are given by

$$\mathbf{v} = A(\theta) \cdot \langle \mathbf{v} \rangle (\lambda, \phi, t) \quad (4)$$

$$\mu_{O_3} = M(\theta) \cdot \langle \mu_{O_3} \rangle (\lambda, \phi, t) \quad (5)$$

The potential temperature θ is used as the vertical coordinate; $A(\theta)$ and $M(\theta)$ are prescribed profiles; λ , ϕ , and t are longitude, latitude and time, respectively; and angle brackets denote column-averaged quantities:

$$\langle \bullet \rangle = \frac{1}{p_0} \int \bullet dp \quad (6)$$

Here p_0 is the pressure along the lower boundary.

The representation given by equations 4–6 accounts for variations in the ozone mixing ratio along isentropic surfaces and is in reasonable accord with observed behavior in the lower stratosphere, where ozone is concentrated. The column abundance m of ozone follows from these equations as

$$m(\lambda, \phi, t) = p_0(\lambda, \phi, t) \cdot \langle \mu_{O_3} \rangle (\lambda, \phi, t) \quad (7)$$

Because both the potential vorticity $\langle Q \rangle$ and the mixing ratio $\langle \mu_{O_3} \rangle$ are approximately conserved within individual air parcels, the redistribution of one of these quantities

heat transfer, effects cause air to drift slightly across isentropic surfaces. This small residual motion turns out to be important for understanding the gradual sinking and rising of air in the mean meridional circulation (see reference 5 and references therein).

Planetary waves and the ozone distribution

Planetary waves can modify ozone in two important ways. By transporting air horizontally, planetary waves draw ozone-rich air poleward from its chemical source region near the equator. Nonconservative influences can then result in a net redistribution of this air, dispersing higher mixing ratios globally.⁸ Planetary waves can also modify the distribution of ozone by moving air vertically. This is especially true in the lower stratosphere, where isentropic surfaces slope downward toward the pole. Air streaming northward along these surfaces descends to higher pressure. The attending compression of that air can increase the ozone density and column abundance substantially, as we will see below.

determines that of the other. Thus, the distribution of $\langle Q \rangle$ predicted in the equivalent barotropic system uniquely determines the distribution of $\langle \mu_{O_3} \rangle$ and, through equation 7, the column abundance.

The numerical calculation is defined by prescribing column-averaged properties above the 350-K isentropic surface, which is roughly coincident with the tropopause. This lower boundary varies in altitude from approximately 100 millibars over the equator to 200 millibars over the poles and underlies most of the ozone layer. An initial circumpolar flow having a column average of 20 m/sec is prescribed, giving the profile of potential vorticity shown in figure 5a. The column-averaged ozone mixing ratio $\langle \mu_{O_3} \rangle$ is scaled to give the distribution of column abundance shown in figure 5b. The values of $\langle \mu_{O_3} \rangle$, which range from about 260 DU over the equator to 300 DU at high latitudes, are representative of conditions during autumn (figure 2), before the planetary wave field amplifies.

The initial zonal circulation is perturbed in the calculation by deflecting its material lower boundary. A wavenumber-1 displacement, in which there is one oscillation around the entire latitude circle, is amplified over 10 days to a limiting deflection of about a kilometer, which is representative of disturbed conditions¹⁴ near 100 millibars. Air motion resulting from this mechanical perturbation is revealed by the behavior of the potential vorticity $\langle Q \rangle$, shown in figure 1. On day 16, the vortex has been displaced out of zonal symmetry, resulting in cross-polar flow at high latitudes. Velocities have increased along the equatorward flank of the vortex, where the meridional gradient of potential vorticity has steepened. Shear accompanying this jet leads to a folding of material lines across a wide belt of latitude. Tropical air is being advected poleward along the periphery of the vortex, in the form of a tongue of low $\langle Q \rangle$ values. A complementary tongue of midlatitude air, with high $\langle Q \rangle$ values, is being drawn equatorward from the vortex.

The folding of potential vorticity contours that has resulted by day 16 is dynamically unstable, setting the stage for the wave field to break. Unstable disturbances amplify near the fold in potential vorticity, leading to secondary eddies that distort the material field further. By day 24, the tongue of tropical air drawn poleward is beginning to spin up anticyclonically to conserve potential vorticity. Material lines are being wound about the induced anticyclone, and midlatitude air drawn from the vortex is being entrained with air of tropical origin. A second anticyclone is forming on the globe's opposite side.

On day 32 the first anticyclone is intensified, as marked by a sharp minimum in potential vorticity and strong flow across the polar cap. Tropical and midlatitude air has been stirred together within this secondary vortex, dissolving into a large pool with potential vorticity values characteristic of much lower latitude. At the same time, the anticyclone on the opposite side of the globe has also

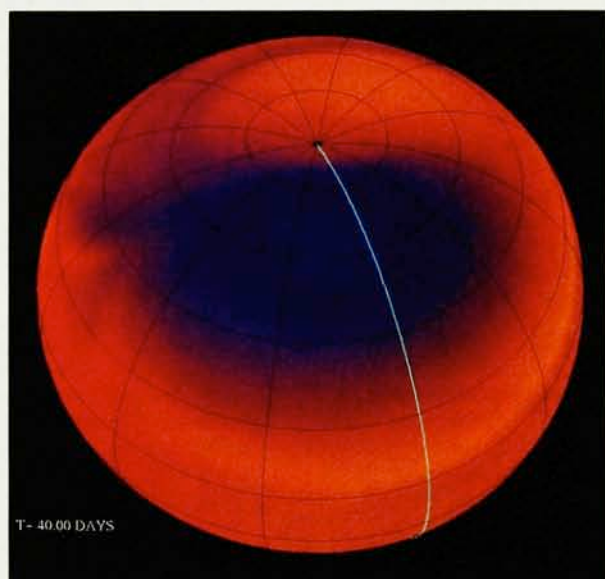
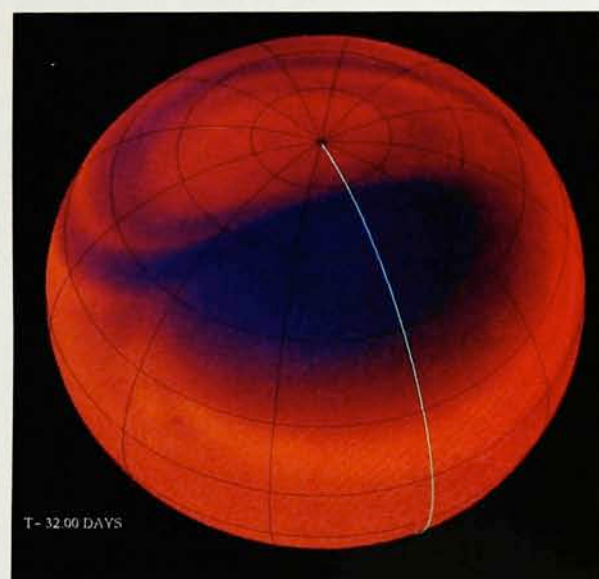
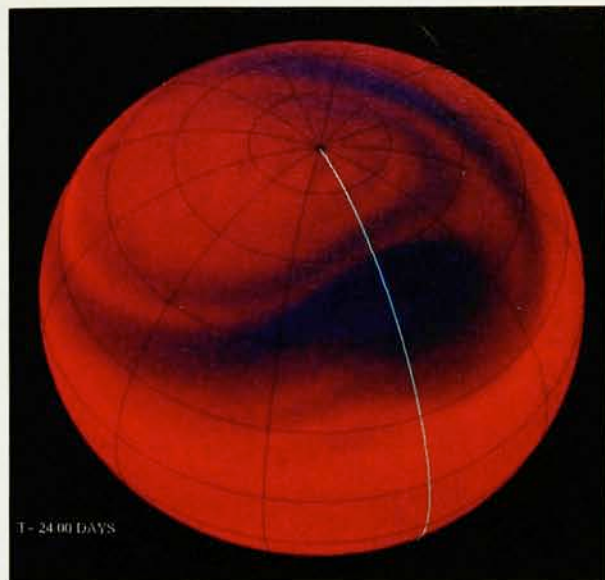
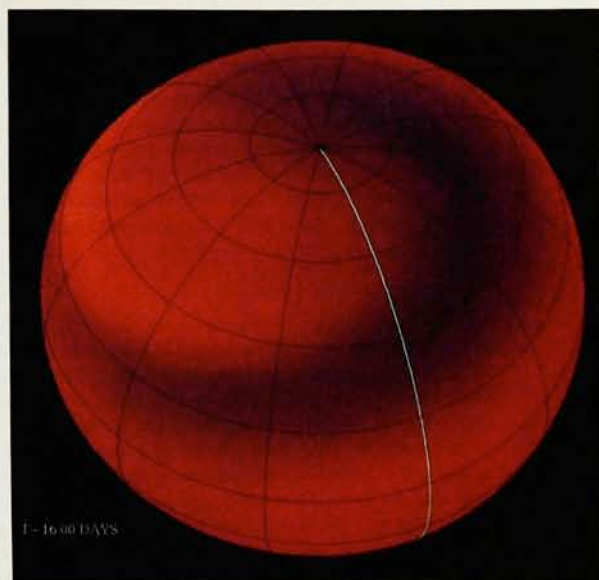
strengthened. Together, these unstable eddies draw air from the vortex and wind it together with air from lower latitudes down to very fine scales. This process continues until bodies of air have been strained down to dimensions where background diffusion is efficient, at which point anomalies in potential vorticity and other long-lived quantities dissolve.

By day 40 the flow has achieved a wavenumber-1 pattern, with a jet flowing almost directly across the pole. This strong meridional motion transports air between low and high latitudes in only a day. Contours of potential vorticity at low latitudes, caught in this circulation, are quickly advected across the polar cap. The anticyclone continues to intensify at the expense of the polar night vortex, which has suffered considerable erosion through large-scale stirring.

Ozone mixing ratios undergo an evolution similar to that shown in figure 1 for potential vorticity. However, the ozone column abundance evolves differently (see figure 6). Unlike the mixing ratio, the column abundance is not conserved for individual bodies of air. As tropical air is drawn poleward along the periphery of the vortex, it descends along isentropic surfaces to lower altitude, where it undergoes compression. The resulting increase in ozone density leads to a significant enhancement of column abundance. Already by day 16, column abundances have approached 400 DU in a broad area overlying the tongue of tropical air drawn poleward, an increase of 30% over the initial values (see figure 5b). On day 24 two column-abundance maxima exist, both associated with tropical air being drawn poleward along the northern flanks of the two anticyclones that have developed through instability. The stronger of these ozone anomalies has values in excess of 470 DU. These large column abundances surround much of the polar cap, which remains under the influence of the vortex (compare with figure 1).

By day 32, an extensive region of anomalously large ozone column abundance has formed at high latitude. Column abundances in excess of 520 DU appear, an increase of nearly 100% over the original values. These magnified ozone abundances correspond to the large pool of tropical air (shown in figure 1) that has been drawn poleward, descended to higher pressure and spun up anticyclonically. By day 40 this secondary circulation has strengthened through the introduction of additional tropical air. Corresponding ozone abundances are increased further as the anticyclone spreads poleward and descends to higher pressure.

One can compare the above results with the variability representative of the middle and upper stratosphere. Satellite observations of the ozone column lying above 30 mb reveal similar features.⁸ However, because isentropic surfaces are nearly level at these altitudes, fluctuations in column abundance are considerably smaller (under 30%) and are due primarily to horizontal transport. The much larger enhancements of total column described above



Ozone column abundance associated with the air motion indicated in figure 1. Orange represents 250 Dobson units; deep blue, 600 DU. Tropical air drawn poleward along isentropic surfaces descends to higher pressure, causing increases in ozone density and column abundance. Ozone column abundance increases markedly inside the anticyclone that develops alongside the vortex (compare with figure 1), exceeding 500 DU by the end of the sequence. The enhanced values result from the poleward displacement of tropical air, which is stirred with midlatitude air, descends along isentropic surfaces and is compressed. Peak ozone column abundances along the highlighted meridian in the four frames are 374 DU, 472 DU, 511 DU and 517 DU. **Figure 6**

result primarily from vertical transport and compression, when air descends along isentropic surfaces in the lower stratosphere. Increases in column abundance of the magnitude seen in these calculations occur routinely in satellite data from the Total Ozone Mapping Spectrometer, which reflect the complete ozone overburden at any location, with the dominant contribution arriving from the lower stratosphere.

Implications for ozone destruction

The Antarctic ozone hole is of profound importance because it represents the first concrete evidence that

human activities have had a substantial impact on this protective layer of the atmosphere. However, large changes in ozone abundance occur naturally, especially in the winter stratosphere, where planetary waves are common. Poleward transport of tropical air in combination with vertical transport and compression can increase column abundances by as much as 100%. However, it should be emphasized that there is no evidence in these calculations of comparable *decreases* in column abundance. The absence of such decreases follows from the fact that the intrusion of tropical air seen in the second panel of figure 1 is compensated by equatorward displacements

of air that are distributed across much of the hemisphere and are therefore smaller, resulting in much weaker vertical transport.

Global-scale disturbances such as those considered here are common in the Northern Hemisphere, where elevated terrain is able to excite a strong planetary wave field. Ozone increases induced by these disturbances can be as large or larger than chemically induced losses over Antarctica, where a 50% reduction in the column is typical, and they can cover an area of hemispheric dimension. Like chemically induced changes, these perturbations are irreversible. Air that is wound up in global-scale eddies such as those depicted in figure 1 is eventually acted upon by diffusion and other nonconservative processes. Once these processes have dissolved inhomogeneities, a net transfer of ozone to high latitudes has taken place.

Eddy motions such as these play a key role in driving the zonally averaged circulation and in producing the large monthly mean ozone abundances indicated in figure 2, which occur at middle and high latitudes during winter and spring. When the vortex is disturbed, as in figure 1, diabatic effects introduce nonconservative behavior. Warm tropical air, displaced poleward, cools radiatively and therefore sinks slightly across isentropic surfaces. Likewise, cold polar and midlatitude air, displaced equatorward, warms radiatively and therefore rises slightly across isentropic surfaces. Each time the vortex is disturbed, air sinks a little at middle and high latitudes and rises a little at low latitudes. Averaging these motions zonally and over many such episodes produces the classical view of the mean meridional circulation and average transfer of ozone from low latitudes to high latitudes. However, the latter that process actually occurs through more complex behavior, such as that shown in figure 1.

Planetary waves such as those examined in this article occur routinely in the Northern Hemisphere during winter. In the Southern Hemisphere, where the Earth's surface features are smaller, the planetary wave field is generally weaker. Cyclone-scale disturbances, decaying vertically from the troposphere in both hemispheres, are ubiquitous in the lowest levels of the stratosphere and also lead to fluctuations in ozone.¹⁵ Natural ozone variability due to these disturbances should be considered when the implications of chemically induced changes are evaluated.

The air motions responsible for these natural changes in ozone also play a key role in shaping chemical depletions. Disturbances such as those shown in figure 1 allow air to flow between the polar cap, where it is in darkness and cools radiatively to space, and the rest of the hemisphere, where it can warm through absorption of solar radiation. Advection of midlatitude air into the polar cap offsets radiative cooling there. By interrupting the symmetric cooling to space, these dynamical effects keep air inside the vortex warmer than it would be otherwise.¹⁶

This process is important in establishing the sharp difference between the circulations over the Arctic and Antarctic and thereby mediates chemical depletions over the North and South Poles. The relative weakness of planetary-wave activity in the Southern Hemisphere allows the Antarctic vortex to become more organized about the South Pole and colder than its counterpart in the Northern Hemisphere. The very low temperatures over the South Pole favor the formation of polar stratospheric clouds, which serve as sites for heterogeneous chemical reactions that liberate chlorine and set the stage for the springtime depletion of ozone. The stronger

planetary wave activity in the Northern Hemisphere keeps the Arctic vortex considerably warmer. The higher temperatures over the North Pole do not support polar stratospheric clouds as readily and therefore inhibit destruction of ozone through heterogeneous chemistry. Stronger planetary wave activity also brings about the springtime ozone maximum sooner in the Northern Hemisphere than the corresponding maximum in the Southern Hemisphere (figure 2). Larger and more frequent exchanges of air transport ozone more efficiently into the Northern Hemisphere than into the Southern Hemisphere.

In addition to these hemispheric differences, planetary waves are involved in the final breakdown of the vortex during spring in both hemispheres. When the equator-to-pole temperature contrast weakens, the vortex is disturbed out of polar symmetry one last time and is replaced by an anticyclonic flow about the pole. In that process, air is overturned across much of the hemisphere, with ozone-rich air from low latitudes replacing depleted air at high latitudes. Hence planetary waves play a key role in the eventual dispersion of ozone-lean air trapped within the Antarctic vortex and in the recovery toward normal ozone abundances. A better understanding of these mechanisms and their influences on the stratospheric circulation will be essential to reaching a fuller appreciation of how the ozone layer functions and how it will respond to changes brought about through human activities.

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The numerical calculations discussed in this article were performed on the Cray-XMP computer at the National Center for Atmospheric Research. The computer graphics were produced by Patrick Callaghan at the Center for Atmospheric Theory and Analysis at the University of Colorado at Boulder and assembled by the NCAR graphics department.

References

1. J. C. Farman, B. G. Gardiner, J. D. Shanklin, *Nature* **315**, 207 (1985).
2. M. R. Schoeberl, A. J. Krueger, P. A. Newman, *Geophys. Res. Lett.* **13**, 1217 (1986).
3. S. Solomon, *Rev. Geophys.* **26**, 131 (1988).
4. National Research Council, *Causes and Effects of Changes in Stratospheric Ozone: Update 1983*, Natl. Acad. P., Washington. (1984).
5. D. G. Andrews, J. R. Holton, C. B. Leovy, *Middle Atmosphere Dynamics*, Academic, Orlando, Fla. (1987).
6. J. London, in *Proc. NATO Adv. Study Inst. on Atmospheric Ozone: Its Variations and Human Influences*, M. Nicolet, A. C. Aikin, eds., US Dept. of Transportation, Washington. (1980).
7. R. Scherhag, *Ber. Dtsch. Wetterdienst.* **6**, 51 (1952).
8. C. B. Leovy, C.-R. Sun, M. H. Hitchman, E. E. Remsburg, J. M. Russell, L. L. Gordley, J. C. Gille, L. V. Lyjak, *J. Atmos. Sci.* **42**, 230 (1985).
9. J. G. Charney, P. G. Drazin, *J. Geophys. Res.* **66**, 83 (1961).
10. J. T. Kiehl, S. Solomon, *J. Atmos. Sci.* **43**, 1525 (1986).
11. M. E. McIntyre, T. Palmer, *J. Atmos. Terr. Phys.* **46**, 825 (1984).
12. M. Jukes, M. E. McIntyre, *Nature* **328**, 590 (1987).
13. M. L. Salby, R. Garcia, D. O'Sullivan, J. Tribbia, *J. Atmos. Sci.* (January 1990).
14. World Meteorological Organization, *Global Ozone Research and Monitoring Project*, report. 16, vol. 1, available from NASA, Earth Science and Applications Division, Code EE, Washington, DC (1986).
15. M. Schoeberl, A. Krueger, *Bull. Am. Meteorol. Soc.* **64**, 1358 (1983).
16. M. L. Salby, R. R. Garcia, D. O'Sullivan, P. Callaghan, *J. Atmos. Sci.*, in press.