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Large-Scale Dynamics and Global Warming

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Abstract

Predictions of future climate change raise a variety of issues in large-scale atmospheric and oceanic dynamics. Several of these are reviewed in this essay, including the sensitivity of the circulation of the Atlantic Ocean to increasing freshwater input at high latitudes; the possibility of greenhouse cooling in the southern oceans; the sensitivity of monsoonal circulations to differential warming of the two hemispheres; the response of midlatitude storms to changing temperature gradients and increasing water vapor in the atmosphere; and the possible importance of positive feedback between the mean winds and eddy-induced heating in the polar stratosphere.

1. Introduction

What surprises might the climate system have in store as it responds to the changing atmospheric composition? Some of the uncertainties that prevent us from developing a fully convincing climate model are well publicized; others are not. What follows is a tour along one particular path through the complex of issues associated with the greenhouse warming problem, emphasizing aspects that involve the planetary-scale fluid dynamics of the atmosphere and the oceans.

2. Atlantic overturning

A remarkable aspect of the earth's climate is the gross asymmetry between the circulations of the Pacific and the Atlantic oceans. The bulk of the cold, dense water filling the Atlantic below roughly 1 km depth is referred to as North Atlantic Deep Water, having had its last contact with the surface, and last felt the effects of heat exchange with the atmosphere, in the northern reaches of the Atlantic. The sinking that occurs in this region is balanced by an overturning circulation depicted schematically in Fig. 1. As it enters the Southern Ocean, this Atlantic water mixes with waters that have seen the surface in the Southern Ocean, and then flows into the Pacific and Indian basins. Surface waters in the Pacific are less dense than this water of mixed North Atlantic–Antarctic ori-

gin; as a result the wind- and thermally driven circulation of the Pacific is much shallower, riding upon and mixing only very weakly with this denser layer.

Why is it that denser water is created at the surface in the North Atlantic than in the North Pacific? One might guess that the key to this difference is the basin geometry, with the Atlantic extending somewhat farther north, or possibly the relatively saline outflow from the Mediterranean. However, recent research points to the possibility of a very different explanation, and one with important implications for climatic sensitivity.

The northward surface flow in the Atlantic transports warm water into high northern latitudes (also warming the land downwind, notably Great Britain and Scandinavia). This warmth in itself would act to destroy the circulation by making the surface waters less dense and less likely to sink. However, the equation of state of seawater near 0°C is such that, in practice, it is the salinity of the water and not its temperature that controls the density. The waters of the northern
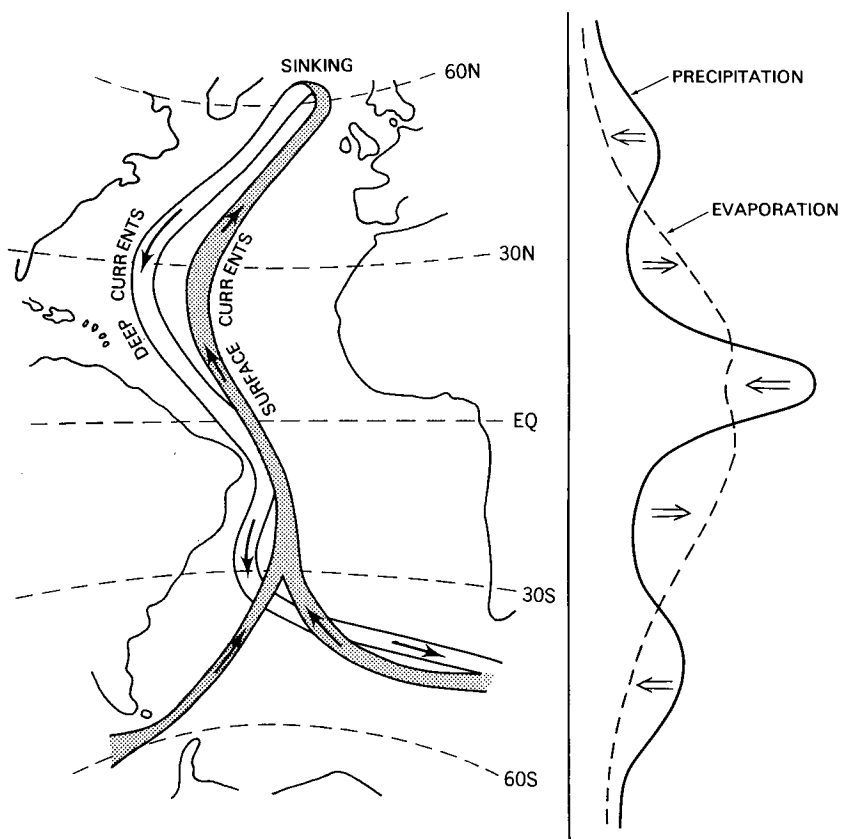

Fig. 1. A schematic of the overturning component of the circulation in the Atlantic, and of the evaporation and precipitation distributions as a function of latitude. Large horizontal currents, primarily wind driven, are not shown.

North Atlantic are much more saline than those of the Pacific. This high Atlantic salinity, in turn, is at least partly due to the overturning circulation itself. Subpolar latitudes are regions of moisture convergence in the atmosphere, where precipitation exceeds evaporation (see Fig. 1). In the Pacific, subpolar waters have a long residence time near the surface, where they are continually freshened and lightened by this excess of precipitation. In the Atlantic, due to the strong overturning circulation, the water spends little time in this region, and remains relatively saline at high latitudes.

This picture of a self-sustaining circulation raises the question of whether a climate without North Atlantic Deep Water formation might also be possible. Inspired by speculations along related lines by Stommel (1961), Rooth (1982), Warren (1983), and Broecker et al. (1985), modeling studies have recently provided some support for this picture. F. Bryan (1986), working with a numerical three-dimensional ocean circulation model developed by K. Bryan and Cox at GFDL, studied an idealized ocean basin symmetric about the equator, in which the boundary conditions controlling the salt, heat, and momentum fluxes across the sur-

face were also symmetric about the equator. He found that such a model can easily sustain a circulation with pole-to-pole overturning and with deep-water formation confined to only one hemisphere, due to the mechanism outlined above. By symmetry, the sinking can occur near either pole, depending on the initial condition.

Marotzke and Willebrand (1991) have elaborated upon this picture. They consider a model with somewhat more realistic geometry, with identical Atlantic and Pacific basins, connected at their southern boundaries by a circumpolar ocean. They find the four stable equilibria illustrated in Fig. 2: one with sinking in both northern oceans, two with sinking in one of these basins only, and one with sinking in neither of the northern basins but with weak sinking north of the circumpolar ocean. The circulation in even these idealized oceans has a complex three-dimensional structure, and a great deal of work remains to be done to better understand the steady and time-dependent behavior of such models.

Manabe and Stouffer (1988) have obtained the important result that a fully coupled atmosphere—
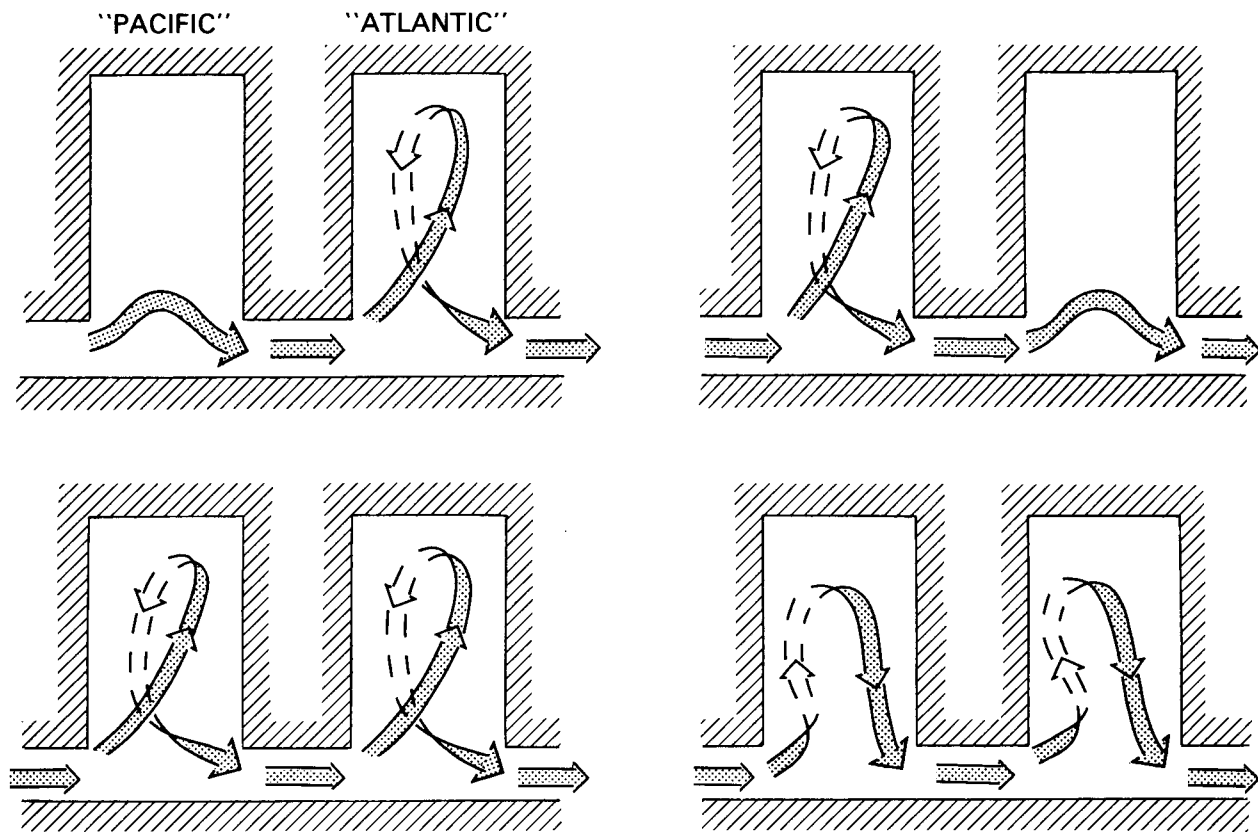
Fig. 2. A schematic of the four distinct climatic states obtained by Marotzke and Willebrand (1991) in a numerical ocean model.

ocean climate model, with realistic ocean basin geometry, can exist in either of two states: one similar to the present climate, with pole-to-pole Atlantic overturning and North Atlantic Deep Water formation, and another with no Atlantic sinking and much greater symmetry between the Atlantic and Pacific circulations. In the limited integrations of this model that have been performed, a spontaneous transition from one state to the other has never been observed. (The differences between the surface air temperatures in these two statistically steady states are reproduced in Fig. 3.) There are a number of open questions concerning the relationship between this result and the ocean-only studies of Bryan, Marotzke, Willebrand, and others, but it is clear that a similar salinity feedback mechanism is playing an essential role. As work with coupled models evolves, it will be important to test the robustness of this result, and to see if coupled atmosphere–ocean models have even more distinct statistical equilibria of physical interest.

As the earth warms in the next century, the salinity of surface waters in high latitudes is expected to decrease. Higher temperatures will result in greater amounts of water vapor and, therefore, greater moisture convergence in high latitudes (assuming that the

atmospheric circulation is not dramatically altered). Therefore, the excess of precipitation over evaporation will also increase at these latitudes (since this excess must balance the moisture convergence), thereby freshening polar waters, weakening vertical convection in the ocean, and increasing the residence time of the subpolar waters in the high precipitation zone. Is it possible that a modest freshening of high-latitude waters, in the presence of this positive feedback, could result in an irreversible transition toward a state in which there is little deep-water formation in the North Atlantic?

Recent calculations at GFDL suggest that by the time the global mean temperature has warmed by 3 K, Atlantic overturning will have weakened by nearly 50%. However, the model has not made an irreversible transition into the "no-sinking" solution at this point (Manabe, personal communication). These results are very preliminary. While such coupled atmosphere–ocean models are capable of simulating many features of the present climate, they also have a variety of deficiencies. It should come as no surprise that these coupled models have difficulty simulating bottom-water formation, given the subtlety of this process. In the study by Manabe and Stouffer cited above, when

their model is left to its own devices it always settles into the no-sinking solution in the North Atlantic. The flux of freshwater from atmosphere to ocean must be adjusted for the model to maintain an oceanic structure resembling that observed. Once the needed adjustment is determined, this fresh water source is held fixed, and the model then is able to maintain itself in either of the two stable states described above. Whether this artificial freshwater source is compensating for deficiencies in the atmospheric or the oceanic model, or both (or in the treatment of sea ice), is an area of current research. Dynamical analysis of a variety of models of different levels of complexity will clearly be needed before we will have confidence in such results.

One naturally would like some observational confirmation of this sensitivity of the Atlantic circulation, short of waiting for the greenhouse warming experiment to be performed. One expects sensitive parts of the system to have large natural variability, and given enough information about the sources of "noise" in the system, one might hope to infer the strength of the restoring forces, as in "fluctuation–dissipation" analyses in many branches of physics. The approach described by Hasselmann (1976) has been useful in studies of the variability of surface ocean temperatures and sea-ice extent on time scales of months to a few years. However, the time scales on which natural fluctuations of the entire North Atlantic circulation are expected to occur range from decades to centuries,

and oceanic data are barely sufficient to allow one to define the basic features of the circulation throughout the depth of the ocean, let alone its variability on these time scales. There are some tantalizing hints, but it is difficult to see how such sparse data can be translated into independent estimates of sensitivity.

Paleoclimatic reconstructions are more promising in this regard. Indeed, studies of the transition from the last ice age into the present interglacial (from about 18 000 to 9000 years ago) have spurred climate modelers to focus on the North Atlantic as a location of dramatic circulation shifts. The retreat of the glaciers not only changed the atmospheric circulation over the Atlantic, but also resulted in a large input of fresh water. Various paleoclimatic indices show that the climate in the North Atlantic and surrounding continents underwent a large fluctuation, known as the Younger Dryas event, during this period. A rapid transition to a warmer climate was interrupted by a retreat, near 10,000 B.P., back toward cold ice-age temperatures, followed by a recovery, all within 1000 years. (The December 1990 issue of *Paleoceanography* is devoted to this phenomenon.) Interactions between the salinity and circulation of the North Atlantic are likely to be at the heart of these fluctuations. The precise timing of the temperature changes as compared to the evidence for changes in freshwater discharge and oceanic circulation is rapidly being refined (e.g., Fairbanks 1990). It may be that simulation of the transition from glacial to interglacial climates with coupled atmosphere–ocean models
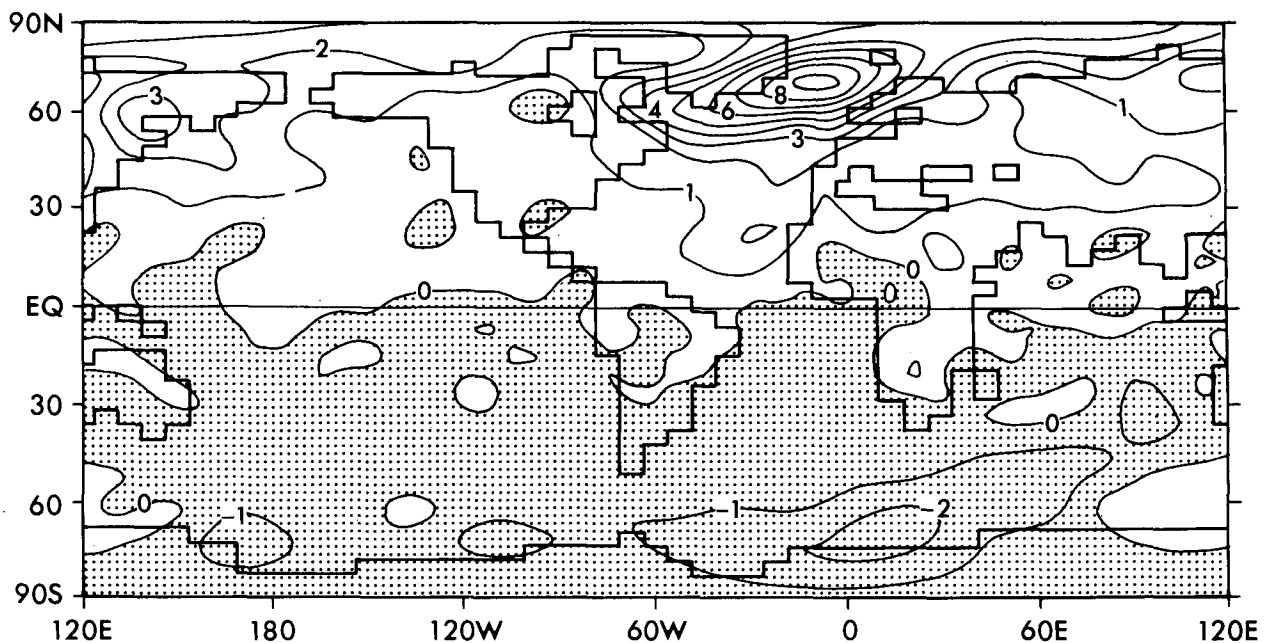

Fig. 3. The temperature difference in the surface air between the two climatic equilibria found by Manabe and Stouffer (1988) in a coupled atmosphere–ocean model (the state with North Atlantic sinking minus the state without). The contour interval is 1 K, with negative values shaded.
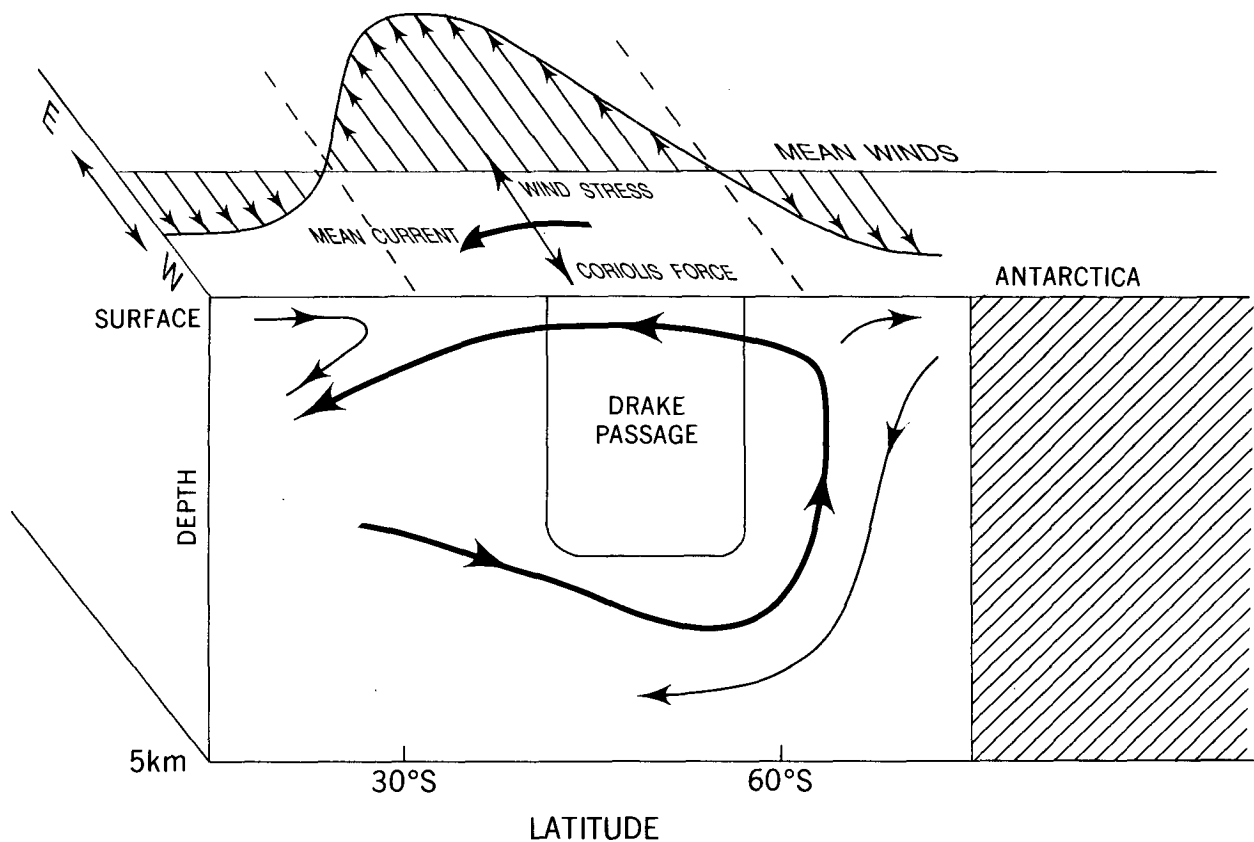
Fig. 4. A schematic of the prevailing surface winds in the Southern Hemisphere, and the overturning circulation, averaged around latitude circles, at the latitude of the Drake Passage.

will be one of the more significant ways of verifying the models, and gaining confidence in their predictions for the sensitivity of the North Atlantic circulation.

3. The Drake Passage

There is no land along the entire length of the latitude circles that pass through the Drake Passage between the South American and Antarctic continents. While this would appear no more than a geographical curiosity, in fact it has important consequences for the World Ocean circulation, as first emphasized by Gill and Bryan (1971).

Large-scale flows in the ocean are in geostrophic balance to first approximation. The most important region where this constraint is broken is the surface mixed layer, within which the stress exerted by the winds is distributed by turbulence, resulting in a three-way balance between the Coriolis force, pressure gradient, and wind stress. Beneath the mixed layer, geostrophic balance implies that the zonally averaged meridional flow in an ocean basin, such as the Atlantic, is proportional to the pressure difference between the eastern and western walls of the basin. By the same

token, the meridional geostrophic flow averaged around a latitude circle passing through the Drake Passage, and above the seafloor topography on this latitude circle must be zero.

The Drake Passage is located within the region of predominantly eastward winds at the surface, and the resulting zonal mean wind stress in the mixed layer must be balanced by the Coriolis force acting on a northward ageostrophic current (see Fig. 4). It follows that sinking in Antarctic waters cannot be balanced, either in the surface layer or in the geostrophic interior, by a simple overturning that advects in warm water from lower latitudes. A sharp density gradient between Antarctic waters and waters north of the passage results, and the pressure gradients that are thereby created balance one of the strongest large-scale currents in the World Ocean, the eastward Antarctic Circumpolar Current. The current is substantially strengthened by direct wind forcing as well.

Calculations by Cox (1989) with an ocean model have shown how the closing of the Drake Passage, and the consequent increase in communication between high southern latitudes and the rest of the ocean, results in waters around Antarctica becoming the dominant deep-water formation regions for the

World Ocean. The North Atlantic evidently owes its present predominant role as a source region for deep water to the existence of this passage, and the resulting tendency toward isolation of the Southern Ocean.

The equatorward flow in the surface mixed layer generated by the wind stress cannot easily be compensated in the oceanic interior at the latitude of the passage. In the absence of comparable Reynolds stresses to those in the mixed layer, these interior layers are geostrophic, and if the ocean bottom were flat, the return flow would be confined to a turbulent bottom boundary layer. In the presence of bottom topography, a geostrophic return flow can be supported by pressure differences across this bottom relief, but the result is still a deep overturning circulation (Fig. 4). Together with the strong small-scale convective mixing poleward of the passage that exists due to the weak gravitational stability of the water column, this overturning provides unusually strong coupling between the deep ocean and the surface waters.

The deep overturning and mixing within the Southern Ocean should have a profound influence on the pattern of greenhouse warming. Figure 5a shows the change in ocean surface temperature predicted at $t \approx$ 75 years in a recent calculation with a coupled atmosphere–ocean model in which the CO_2 concentration increases at a rate of 1% per year, starting with the value of 300 ppm at $t = 0$ (Manabe et al. 1991). (This increase is more rapid than that observed for CO_2 in the atmosphere, but was chosen to account crudely for the other greenhouse gases as well. It should be emphasized that this model's sensitivity is partly controlled by the predicted changes in cloudiness, which are very uncertain.) After 75 years, the Southern Ocean has not yet warmed significantly, while the global mean surface air temperature has increased by 3 K. Immediately to the south of the Drake Passage, in the region of intense vertical mixing and mean upwelling, surface waters are being mixed with and replaced by deep waters that know nothing of the CO_2 increase.

Can the model prediction of a long delay in the warming of the southern oceans be trusted? One possible criticism is that the model could be misrepresenting the force balance in the Circumpolar Current. This current is observed to be unstable, spinning off eddies on scales of a few tens to hundreds of kilometers. Numerical models that have been used to date to study global warming have insufficient resolution to capture these instabilities. It is possible that these eddies could redistribute sufficient momentum in the vertical so as to generate an ageostrophic mean meridional flow in the interior, just as the turbulent stresses within the mixed layer do near the surface,

resulting in a shallower circulation and weaker coupling to very deep layers. Calculations are under way in several research centers to model the dynamics of this current at high resolution so as to address such issues. A British consortium, the FRAM Group (1991), has initiated a particularly ambitious effort.

Ocean observations must also be used to directly test the circulation patterns predicted by the model, but this is not straightforward because of the scarcity of direct current measurements. Measurements of transient tracers, such as the ^{14}C produced by bomb tests and the CFCs, promise to provide rigorous tests of the mixing and transport patterns predicted by numerical ocean models (e.g., Toggweiler et al. 1990).

The extent to which the oceanic circulation, constrained as it is by the existence of the Drake Passage, can prevent warming of the Southern Ocean promises to become a central issue in global warming research. It is clearly central to the issue of how the Antarctic ice sheet will evolve, but it is also important for the atmospheric circulation in the Southern Hemisphere.

4. Tropospheric dynamics

If high southern latitude surface temperatures are unable to warm rapidly, due to strong coupling with the deep ocean, while lower latitudes do warm (see Fig. 5), the result will be an increase in the north–south temperature gradient in the Southern Hemisphere. Through the process of baroclinic instability, the potential energy associated with this gradient is converted into the energy of the atmospheric eddies that dominate horizontal heat and momentum transports outside of the tropics. One intuitively expects these eddies to grow and their transports to increase if the north–south temperature gradient increases, an intuition that is confirmed by a variety of models of finite-amplitude baroclinic instability.

When these eddies disperse out of their region of excitation in midlatitudes, they transport eastward angular momentum into this source region. The stress exerted by the surface on the atmosphere, in the region of surface westerlies, balances this momentum flux convergence into midlatitudes. As temperature gradients increase, one expects larger eddy energies in midlatitudes of the Southern Hemisphere and greater radiation of disturbances to the north and south, producing larger angular momentum fluxes as well. The strength of the surface winds must then also increase to maintain angular momentum balance.

Among the consequences of this invigoration of the Southern Hemisphere circulation, the effect on the oceanic circulation is of particular interest. If the momentum transports do increase in the Southern Hemi-
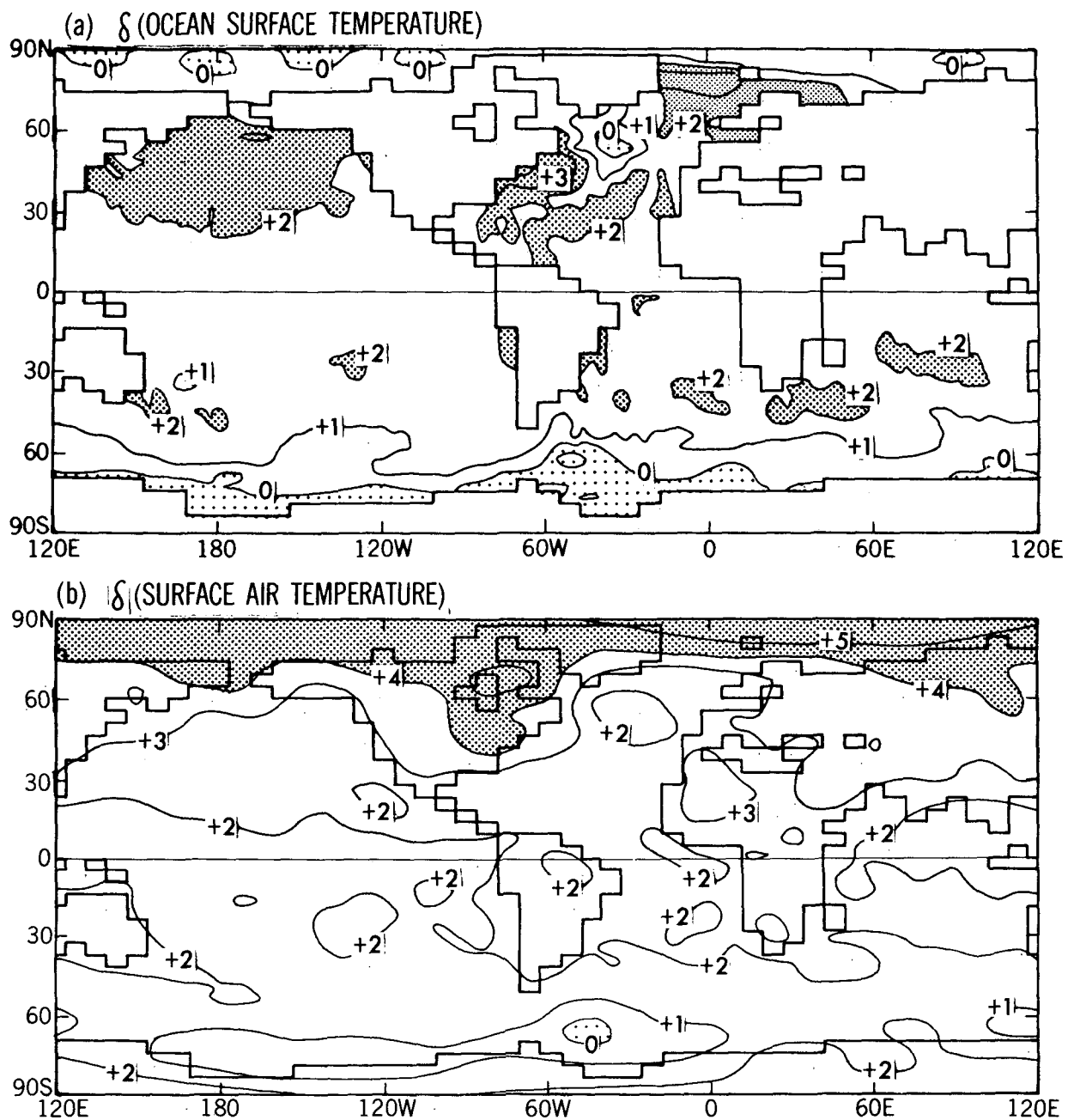
(a) δ (OCEAN SURFACE TEMPERATURE)

(b) |δ|(SURFACE AIR TEMPERATURE)

Fig. 5. (a) The change in the annual mean ocean surface temperature predicted by a coupled atmosphere–ocean model after 75 years of increasing CO_2 concentration (1% per year); (b) the corresponding change in surface air temperature. Contour interval is 1 K. The idealized continental outlines are those of the model. (The model data is actually averaged over years 70 to 80.) [From Manabe et al. (1991).]

sphere in response to the increased temperature gradient, the eastward wind stress at the latitude of the Drake Passage would increase, driving a stronger overturning cell in the ocean, with stronger upwelling to the south and increased northward spreading of this upwelled cold water in the surface layers. The result is a greenhouse-induced cooling of polar waters. This is a temporary effect, admittedly lasting for centuries, that would disappear as the source of the upwelling waters is eventually mixed with warmed waters from other regions. Cooling in this region is, in fact, apparent in Fig. 5a. The problem is complicated by the existence of an alternative, somewhat more direct mechanism that promotes cooling in regions of deep oceanic convection: the freshening of subpolar waters due to increased moisture convergence in the atmosphere lightens the surface water, allowing it to cool more before becoming denser than the underlying

layer. Both of these mechanisms appear to be operating in the model that generated Fig. 5.

Models of the equilibrium response of climate to an increase in the atmospheric CO_2 concentration generally show the largest surface warming in polar latitudes. (The equilibrium response is obtained by waiting until the entire atmosphere–ocean system has adjusted to the new CO_2 level, ignoring the interaction between the ocean circulation and the carbon cycle that helps control the CO_2.) This polar amplification is a result of two effects: the retreat of highly reflective snow and ice cover; and the large gravitational stability of the atmosphere near the surface in the high-latitude winter, which has the tendency to concentrate the warming at low levels. (This is a wintertime effect; during the summer there can be little warming at polar latitudes for the simple reason that temperatures are pegged at the melting temperature of ice.) In contrast to the Southern Hemisphere, the effective heat capacity is small enough in the north, due to the large landmasses, that polar amplification is seen in the transient response to a gradual buildup of greenhouse gases (see Fig. 5b). Should one expect this decrease in the pole-to-equator temperature gradient to weaken the midlatitude baroclinic eddies during northern winter?

One complicating factor is the prediction of most models that the pole-to-equator temperature gradient will increase in the upper troposphere, as shown schematically in Fig. 6. This increase is a consequence of the large warming predicted to occur in the tropical upper troposphere. The upper-tropospheric maximum in the warming, in turn, reflects the fact that if we start with two air parcels near the surface, one warmer and therefore most likely containing more water vapor, the difference in temperature amplifies as the parcels travel upward, since more latent heat is released in the warmer parcel. The quantitative prediction of this upper-tropospheric warming differs from model to model, being sensitive to detailed assumptions about convection and clouds.

The dominant wintertime baroclinic eddies are coherent through the depth of the troposphere in midlatitudes. As a result, it is unclear whether the eddies would respond primarily to the decrease in lower-tropospheric temperature gradient or the increase in the upper-tropospheric gradient. (In the Southern Hemisphere, both gradients increase if the scenario associated with Fig. 5 is correct, so the issue does not arise.) Linear instability theory points to the lower-tropospheric gradients as being dominant, and some recent highly idealized homogeneous geostrophic turbulence calculations (Held and O'Brien 1992) support this conclusion. The issue is still open; for example, it seems plausible that eddy heat fluxes are
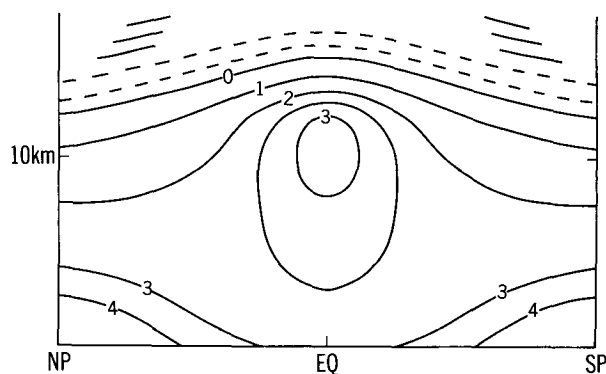

FIG. 6. A schematic of the equilibrium annual mean temperature response to a doubling of CO_2, as typically predicted by GCMs, emphasizing the maxima at upper-tropospheric levels in the tropics and at low levels in the polar regions. Polar amplification is present only in winter.

sensitive to lower-tropospheric temperature gradients, but that the eddy momentum fluxes, which are strongest near the tropopause, are sensitive, in addition, to upper-tropospheric conditions that determine the extent to which disturbances are able to radiate away from their source latitudes. The decrease in temperature gradient in the lower troposphere, and the increase in the upper troposphere, could also result in eddies that are more top heavy, with a larger fraction of their total energy at upper levels.

The response of the midlatitude eddies is greatly complicated by the fact that these storms are strongly influenced by moisture in the atmosphere. The increase in moisture that accompanies the warming will itself have two distinct and competing effects on these eddies.

Warm air rises as it moves poleward in a midlatitude eddy, while cold air sinks as it moves equatorward, and the resulting lowering of the center of mass is ultimately the source of the kinetic energy of the storm. Latent heat release strengthens the storm, as it occurs primarily in the warm sector of rising air, thereby generating additional vertical motion and kinetic energy. On this basis one intuitively expects more intense storms in a moister atmosphere.

But there is another, potentially more powerful effect. These eddies transport energy poleward, and it is the balance between this transport and the pole-to-equator heating gradient that determines the statistically steady temperature gradient on the earth as well as the average strength of the eddies. The eddies carry moisture poleward also, since the warm air tends to be moist and the cold air dry. This moisture transport contributes to the energy transferred poleward: if water evaporated in the subtropics is transferred to higher latitudes before condensing and releasing its latent heat, the temperature gradient is reduced com-

pared to that which would exist if the vapor condensed at the latitude of evaporation. As the atmosphere becomes moister, this latent heat transport increases, making the eddies more efficient at transporting energy polewards. Smaller eddies are then required to maintain the same temperature gradient, so we can anticipate that eddy amplitudes will decrease as the moisture content increases.

Both of these effects could be important. I suspect that the latter effect is more important at the planetary scales that are responsible for most of the horizontal energy transport, while the direct enhancement of eddy energy by latent heat release is more likely to be important on the smaller-scale storms that dominate the "weather." It is possible that the predictions of the relatively coarse-resolution atmospheric models that have been used to date to study climate change may be significantly modified by a new generation of models that captures the dynamics of these smaller storms more faithfully. Much work is evidently needed to sort out the various controls on the strength, and the spectrum, of the midlatitude eddy field.

There is some concern about a poleward shift of the midlatitude rainbelts in winter, a response that would be of great importance to areas with Mediterranean climates, in which much of the rain falls from the equatorward edge of the wintertime storm track. In models where this result holds when averaging around a latitude circle, it can still be overwhelmed by local effects in particular regions. It is most clearly seen in model calculations with flat continents and simplified continental boundaries (e.g., Manabe et al. 1981). The effect of changing horizontal temperature gradients and moisture on the latitudinal structure of the eddy activity is not well understood. But it may also be necessary to focus on changes in vertical temperature gradients. Baroclinic instability is enhanced by reducing the gravitational stability of the air column. The maximum in the warming near the surface reduces the gravitational stability in high latitudes during winter, which should tend to shift the region of greatest baroclinic instability poleward in the Northern Hemisphere.

In the northern winter, the most energetic storms tend on average to develop near the east coasts of Asia and North America and decay in the western Pacific and Atlantic (see Fig. 7). Modest changes in the location of these storm tracks can have dramatic effects on regional climate, overcoming the effects of the zonally symmetric climatic responses discussed above. Progress in understanding the factors that control these storm tracks is being made by analyzing year-to-year variability in their location and strength, by performing linear stability analyses of realistic planetary-scale flows, by studying the nonlinear dynamics of a variety of idealized atmospheric models, and by

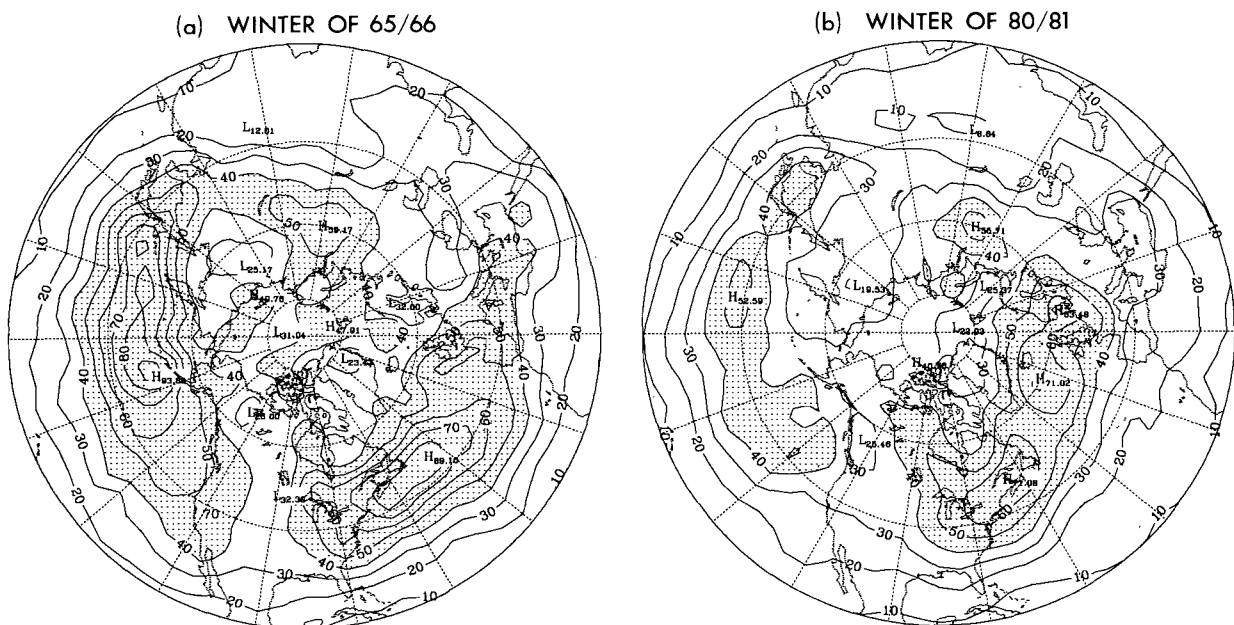(a) WINTER OF 65/66

(b) WINTER OF 80/81

FIG. 7. Stereographic plot of the Northern Hemisphere poleward of 20°N, with contours of the standard deviation of the height of the 500-mb pressure level in the atmosphere. The data has been filtered so that only periods from 2 to 7 days are considered. (These are the periods characteristic of large-scale baroclinic eddies; the comparable or greater variance at lower frequencies has a distinct horizontal distribution.) Contour interval is 10 m. Two different winters have been selected to illustrate the interannual variability of the storm tracks (these are extreme cases). Figure kindly provided by N.-C. Lau and M. J. Nath, based on data from the National Meteorological Center.

performing sensitivity studies with GCMs, but this remains one of the major challenges in dynamical meteorology.

Preferred regions of storm growth and decay exist because of the zonal asymmetries of the larger-scale flow and of the temperature field in the atmosphere. These asymmetries result, in turn, from asymmetries in the lower boundary condition for the atmosphere, primarily the major orographic features (such as the Tibetan plateau and the Rockies) and the land–ocean heating contrast. Numerical modeling is providing new information on how the Tibetan plateau, for example, or the large east–west surface temperature gradient across the tropical Pacific, affects the global circulation. The responses are often surprisingly nonlocal; for example, changes in the tropical Pacific Ocean temperature distribution associated with the El Niño phenomenon are known, through both modeling and observational studies, to have a significant effect on the flow field and storm tracks over the North Pacific and North America.

Since the observed zonal asymmetries are the (nonlinear) superposition of the response to various features of the lower boundary, it is difficult to predict how the dominant planetary-wave pattern and the associated storm tracks will change as the climate warms. As an example of the kind of speculation that is possible, it can be argued that the zonal asymmetries in the flow forced by midlatitude orography, such as the Rockies, should increase in amplitude if the north–south temperature gradient decreases near the surface: the adiabatic heating that occurs as the air is forced up and down over the orography is primarily balanced by the north–south advection of temperature; with a smaller temperature gradient, air must be brought in from farther away to balance the temperature changes associated with the vertical motion. [Cook and Held (1988) have argued that this relation is important for the ice-age climate, where the increased north–south temperature gradient reduces the response of the flow to the Laurentide ice sheet.] But this is only one of several factors that could affect the longwave pattern over North America.

Storms are "steered" by the larger-scale flow, but they also help to determine the zonal asymmetries in this flow through their fluxes of heat, momentum, and water vapor. Studies of natural variability of storm tracks suggest that this feedback is often positive: an anomaly in the longwave pattern creates an anomalous storm track whose rectified effects tend to enhance the anomalous larger-scale flow. Indeed, it seems that planetary-wave patterns that persist for significant periods are those for which the storm track provides positive feedback. In related work, numerical studies of the atmospheric response to perturbations

in the sea surface temperature distribution suggest that modest changes in these boundary conditions can have significant effects on the storm tracks, and that the feedback of these storms onto the larger-scale flow is essential to this response (e.g., Palmer and Sun 1988; Lau and Nath 1990). Consistently, recent analysis suggests that the storm track's fluxes play an important role in determining the changes in the wintertime stationary-wave pattern as the climate warms in a GCM (Stephenson and Held 1993).

There is substantial interannual variability in the storm tracks, as exemplified by Fig. 7. It appears that a significant part of this variability can be thought of as the response to variations in the ocean temperatures that provide a lower boundary condition for the atmosphere. To the extent that this part of the interannual

The aspect of the summertime climate that has attracted the most attention in the context of global warming is the possibility that continental regions will become drier in the summer.

variability can be isolated, it should prove invaluable in testing model predictions of how the path of the wintertime storm tracks will respond to the global warming.

The aspect of the summertime climate that has attracted the most attention in the context of global warming is the possibility that continental regions will become drier in the summer (Manabe et al. 1981). The reader is also referred to the IPCC report (1990). Earlier snowmelt and the poleward movement of the midlatitude precipitation maximum in spring both lead to drying (the latter effect being dominant in middle latitudes and the former at somewhat higher latitudes), which then amplifies through the summer due to enhanced heating of the drier surface and reductions in cloudiness. Wintertime precipitation tends to increase in middle and higher latitudes, but the memory of this wetness does not survive into the summer since the soil in winter is typically saturated in any case. Better understanding of this "summer dryness" will depend as much on improved theories for the relevant ground hydrology as on improvements in modeling the atmospheric flow.

Changes in monsoon circulations can also be expected. Model results summarized in the IPCC report tentatively suggest that the East Asian monsoon will be strengthened on average. This is evidently a response to the relatively large warming of the Asian landmass, which, in turn, is sensitive to the aforementioned summer dryness.

A plausible case has been made by Folland et al. (1986) that a small warming of the Southern, relative to the Northern, Hemisphere has been at least partly responsible for the Sahel droughts of the 1970s and 1980s. It appears that a warming of one hemisphere with respect to the other can displace the monsoonal rains slightly toward the warmed hemisphere. The argument is based partly on observed correlations and partly on GCM sensitivity studies. This result is of great interest in any case, but may also have implications for global warming. If the Northern Hemisphere warming does outpace that in the Southern in the next century, as suggested by Fig. 5 (but counter to the observed warming from the 1960s to the 1980s), this might spell some relief for regions, such as the Sahel, that depend on the northermost extension of the summer monsoons for their rainfall. Corresponding regions in the Southern Hemisphere would suffer. On the other hand, the Sahel drought would intensify if the Southern Hemisphere, for whatever reason—perhaps because of the counteracting effect of aerosol pollution in the north—continues to warm more rapidly.

The relative importance of local and remote forcing for monsoonal circulations remains very poorly understood, however. The central complication is that the precipitation is strongly controlled by the large-scale monsoonal flow that is itself primarily driven by this latent heating. Relatively small radiative- or sensible-heating gradients can drive a weak flow, forcing moisture convergence and excess precipitation in some regions; the resulting latent heating then forces a circulation that strongly reinforces the precipitation pattern, overwhelming the initial impulse. As a result, it is very difficult to isolate the ultimate from the proximate cause of the circulation by simply diagnosing the observed flow and heating distributions. Numerical atmospheric models therefore play an essential role in these studies, even though there remain many uncertainties on how to incorporate moist convection in such models.

5. Stratospheric cooling

Finally, consider the cooling of the middle atmosphere (the stratosphere and mesosphere) that is anticipated to result from the increase in the concentration of carbon dioxide.

To first approximation, temperatures in the middle atmosphere are determined by the balance between absorption of solar photons by ozone and emission by carbon dioxide in the infrared. Well-established radiative models predict that a doubling of the carbon dioxide concentration will result in a cooling larger than 5 K throughout most of the stratosphere (reaching $\approx$

10 K near 50 km), if 1) ozone concentrations do not change, and 2) it is in fact adequate for this purpose to assume radiative equilibrium, or, more precisely, that the nonradiative heating of the atmosphere is unchanged (Fels et al. 1980).

Needless to say, the first of these assumptions must be rejected, given the consensus that the catalytic action of chlorine compounds resulting from the photodissociation of man-made CFCs will result in a significant reduction in global ozone levels. In the present climate, the chemistry responsible for the Antarctic ozone hole, involving reactions on the surfaces of cloud particles that form only at very cold temperatures, is activated only fleetingly in the Arctic. There is concern that the general cooling of the stratosphere associated with the CO_2 increase will help create conditions in the Arctic that will also be conducive to rapid ozone loss. At present the temperatures at high northern latitudes as they emerge from polar night are on on the order of 20 K warmer than their analogs in the Southern Hemisphere, a fact that we return to below.

The second assumption (fixed nonradiative heating) may also be inappropriate. The middle atmosphere in wintertime is far from radiative equilibrium. During the polar night, the only radiative heating in the middle atmosphere results from absorption of the infrared flux upwelling from the troposphere. As Fels (1987) has emphasized, this upwelling flux, acting alone, would maintain stratospheric temperatures as much as 70 K colder than observed near the North Pole, taking tropospheric temperatures as given. A small part of this difference can be explained by taking into account the slow large-scale atmospheric motions directly induced by the seasonally varying radiative forcing (the atmosphere responds to transient cooling by generating downward motion that increases temperatures through adiabatic compression), but most of the discrepancy remains. This departure from radiative equilibrium is now understood as being induced by the deceleration of the zonal winds by waves of various scales propagating into the region from below.

The middle atmosphere would be relatively quiescent if it were not constantly perturbed by disturbances generated in the much more turbulent troposphere. Hydrostatic and geostrophic balance requires a zonal flow from west to east that increases with height in the winter hemisphere, resulting in the polar-night jet in the mesosphere (see Fig. 8). Instabilities of this jet appear to be relatively unimportant. The large disturbances that exist in the middle atmosphere, particularly in the winter hemisphere, have propagated up from their sources in the troposphere. On scales of a few hundred kilometers or less, these disturbances
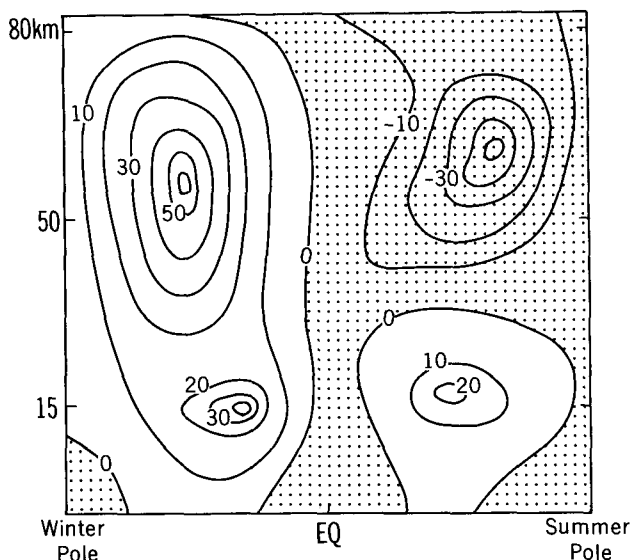
Fig. 8. A qualitative picture of the zonal wind distribution in the middle atmosphere averaged around latitude circles, showing the tropospheric subtropical jets and the mesospheric jets. Areas of easterlies are shaded.

are primarily gravity waves; on larger scales, they are Rossby (or "planetary") waves. Gravity waves are excited by the deflection of the flow near the surface by small-scale orographic features, by turbulent moist convection, and by the cascade of the "macroturbulence" of the troposphere to small scales, particularly in regions of frontal formation. Planetary waves are excited by large-scale orography, such as the Tibetan plateau and the Rocky mountain complex, by slowly evolving large-scale heat sources associated with the continent–ocean geometry, and by the cascade of the tropospheric macroturbulence to larger scales.

As they propagate into regions of ever-smaller density, these waves ultimately break in complex ways. When planetary waves break, they decelerate the zonal flow (the interested reader is referred to Andrews et al. 1987). Gravity waves have a similar effect, although the mechanism is quite different. When the zonal flow is decelerated, the north–south temperature gradient must be reduced and the atmosphere warmed in high latitudes (by adiabatic compression due to induced sinking motion) to maintain the pressure gradients required by hydrostatic and geostrophic balance. An extreme example of this phenomenon is the "sudden stratospheric warming," which occurs in roughly 50% of northern winters, in which a particularly large disturbance from the troposphere breaks in the stratosphere and temporarily disrupts the entire polar vortex.

The relative importance of the gravity-wave and planetary-wave drag in different regions of the middle

atmosphere is a subject of active research. However, there is general agreement that gravity waves dominate near the mesopause ($\approx$ 80 km) and above, and that they control the height at which the polar-night jet reaches its maximum strength. Within the wintertime stratosphere in the Northern Hemisphere, planetary waves are thought to predominate. The Southern Hemisphere stratosphere in winter has a stronger polar-night jet and colder polar temperatures than its northern counterpart. The primary reason for the difference is the weaker planetary-wave activity in the Southern Hemisphere, the wave activity being weaker in turn because of the near absence of the continent-ocean heating asymmetries and large-scale topographic features that are the largest source of planetary waves in the Northern Hemisphere.

There are suspicions that this planetary-wave drag is a sensitive part of the climate. To understand the basis for these suspicions, we need one more piece of information concerning planetary-wave propagation.

Whether or not a large-scale disturbance is able to propagate from the troposphere into the stratosphere is determined by its horizontal wavelength and the strength of the mean zonal flow U, as illustrated schematically in Fig. 9. (Our convention is that $U > 0$ if the flow is westerly.) Note first that $U > 0$ is a requirement for propagation. (Strictly speaking, this result is appropriate in a frame of reference moving with the wave source, but the dominant tropospheric
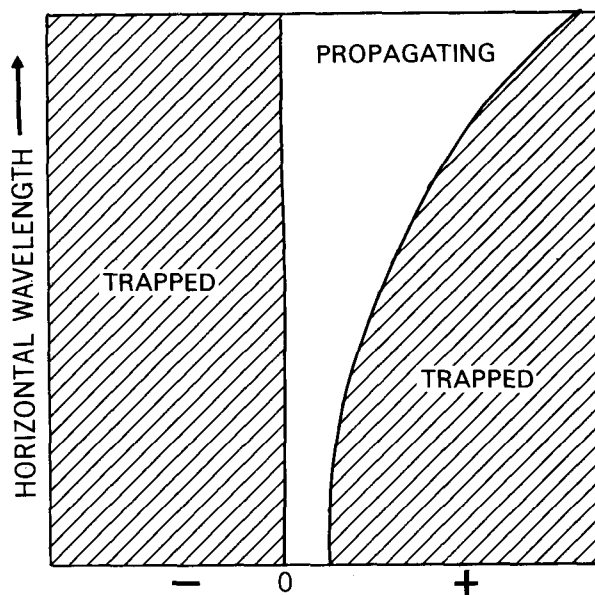

Fig. 9. A plot showing when a planetary-scale wave can propagate vertically, and when it is trapped at the height of excitation, as a function of background zonal wind and horizontal wavelength. The horizontal axis is the mean zonal flow (U).

sources are all moving slowly compared to the strong stratospheric winds, so it is adequate for our purposes to apply this result in a frame of reference fixed with respect to the surface.) In the summer hemisphere, the stratospheric wind U is strongly negative, and on this basis large-scale tropospheric disturbances are not expected to radiate upward, in excellent accord with observations.

When $U > 0$, planetary waves can still be trapped if the winds are sufficiently strong, and the strength of the wind required for trapping is an increasing function of the horizontal scale of the wave. The parameters are such that only the very largest planetary scales can propagate into the wintertime stratosphere. This is again in agreement with observations: the energy containing eddies in the wintertime stratosphere have larger horizontal wavelengths than the eddies in the troposphere.

This scale-selective trapping creates the possibility for positive feedback between the vertically propagating waves and the zonal winds through which they propagate. To the extent that they penetrate into the stratosphere and break there, these waves decelerate the winds and warm polar latitudes. If the strength of the tropospheric wave source is reduced, then the flow will increase in strength, and this, in turn, will tend to trap a larger part of the wave field, further reducing the wave-induced drag. Simple one-dimensional models of vertically propagating waves interacting with the mean zonal flow have been used to analyze this feedback in Plumb (1988) and Yoden (1990). These models undergo a rapid transition, as the strength of the wave source is reduced from a Northern Hemisphere–like solution, with weak winds and strong wave penetration, into a solution that more nearly resembles that seen in the Southern Hemisphere, with stronger winds, weaker penetration, and polar temperatures closer to radiative equilibrium.

Should we be concerned that the climatic changes associated with the changing atmospheric composition could induce such a transition, thereby (among other things) creating conditions in the Arctic similar to those that have resulted in the Antarctic ozone hole? One reassuring fact is that despite substantial interannual variability in the planetary waves propagating from the troposphere into the stratosphere, the wintertime circulation in the Northern Hemisphere stratosphere never resembles that in the Southern Hemisphere. There are also several dynamical reasons, which we do not have the space to discuss here, to question the relevance of the one-dimensional models of this feedback. Furthermore, it is not obvious that one should expect a reduction in the tropospheric source of the very large-scale waves that are able to propagate upward. In fact, one of the arguments

mentioned in the previous section, relating the meridional temperature gradient to the amplitude of the waves generated by midlatitude orography, suggests that this source could actually increase in strength. [Rind et al. (1990) describe a numerical simulation in which the waves in the northern winter stratosphere do increase in amplitude as the climate warms.]

But there is one effect that does push the system in the direction of stronger stratospheric winds: as the CO_2 increases, the rate at which temperatures relax toward their radiative equilibrium state also increases [by roughly 20% for a doubling of CO_2 (Fels 1987)]. For fixed nonradiative heating, an atmosphere with a stronger radiative spring constant will be driven closer to radiative equilibrium, and have a larger north–south temperature gradient and stronger winds. Even with a fixed tropospheric planetary-wave source, the positive feedback described above could then enhance this response. The interested reader is referred to Fels (1987) for further discussion of this and related topics.

Numerical modeling of the flow in the stratosphere and mesosphere is a challenging task, given the importance of disturbances propagating up from the troposphere, and given the complex sources of these disturbances. Progress along these lines is described by Mahlman and Umscheid (1987). Results from a new generation of models will be closely examined in the coming years for signs of significant positive feedback between changes in the mean zonal flow and the wave-induced deceleration, in response to changes in stratospheric climate caused by greenhouse warming and ozone depletion.

6. Final remarks

We are far from anything resembling a "theory" of climate, and cannot expect a theoretical (necessarily computational) approach, in isolation, to yield a totally convincing prediction of climatic sensitivity in the near future. This becomes especially clear when attention is focused on the interactions between the large-scale flow and the various small-scale, moist-convective, cloud-determining processes (a topic I have managed to avoid discussing in this essay). Comprehensive numerical climate models will continue to play a central role in this field, and the analysis of a hierarchy of simpler models will be needed to help develop confidence in the robustness of predicted responses, as well as to develop rational strategies for analyzing and comparing the predictions of various comprehensive models. But for such a complex system, it should be obvious that the most significant progress is likely to be made through a creative blend of theory and observation. The challenge will be to find those key observa-

tions, whether in paleoclimatic records or in the interannual variability of the atmosphere and oceans, or in the first hints of the structure of the greenhouse warming itself, that can be used to constrain or verify the models in relevant ways.

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