The Global Atmospheric Circulation of Saturn

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Over the past decade, the Cassini spacecraft has provided an unprecedented observational record of the atmosphere of Saturn, which in many ways now surpasses Jupiter as the best-observed giant planet. These observations, along with data from the Voyager spacecraft and Earth-based telescopes, demonstrate that Saturn, like Jupiter, has an atmospheric circulation dominated by zonal (east-west) jet streams, including a broad, fast eastward equatorial jet and numerous weaker jets at higher latitudes. Imaging from Voyager, Cassini, and ground-based telescopes also document a wide range of tropospheric features, including vortices, waves, turbulence, and moist convective storms. At large scales, the clouds, ammonia gas, and other chemical tracers exhibit a zonally banded pattern whose relationships to the zonal jets remains poorly understood. Infrared observations constrain the stratospheric thermal structure and allow the derivation of stratospheric temperatures; these exhibit not only the expected seasonal changes but a wealth of variations that are likely dynamical in origin and highlight dynamical coupling between the stratosphere and the underlying troposphere. In parallel to these observational developments, significant advances in theory and modeling have occurred over the past decade, especially regarding the dynamics of zonal jets, and we survey these new developments in the context of both Jupiter and Saturn. Highly idealized two-dimensional models illuminate the dynamics that give rise to zonal jets in rapidly rotating atmospheres stirred by convection or other processes, while more realistic three-dimensional models of the atmosphere and interior are starting to identify the particular conditions under which Jupiter- and Saturn-like flows—including the fast equatorial superrotation, multiple jets at higher latitudes, storms, and vortices—can occur. Future data analysis and models have the potential to greatly increase our understanding over the next decade.

11.1. Introduction

The dynamics of Jupiter and Saturn have long been the subject of fascination. Although Galileo trained the telescope on both objects starting in 1610, the pace of atmospheric discoveries over subsequent centuries was far slower for Saturn than for Jupiter because of the fact that Saturn is dimmer, smaller in Earth’s sky, and more muted in its cloud-albedo contrasts. The first description of Jupiter’s zonal banding occurred in 1630. Cassini and others discovered short and long-lived spots, the equatorial current, and other atmospheric features on Jupiter starting in the latter decades of the 1600s, and ever more refined observations have continued episodically since then (e.g., Rogers 1995). In contrast, early observations of Saturn could not make out features on the planet’s surface, instead emphasizing the discovery of Saturn’s moons and the nature of its rings (van Helden 1984). Cassini reported an equatorial belt on Saturn in 1676, and hints of non-zonal atmospheric features appeared a century later in the 1790s, but it took until the latter decades of the 1800s before spots could reliably be observed on Saturn (van Helden 1984). This disparity in pace of discovery between Jupiter and Saturn continued through the twentieth century, and even over the last few decades, the relative difficulty of observing Saturn relative to Jupiter means that Saturn has received much less attention than its sister planet. Yet the Cassini mission, in orbit around Saturn since 2004, has revolutionized the quality and quantity of observational constraints, which now rival or exceed those available for Jupiter. This for the first time places Saturn at the forefront of understanding giant planets as a class.

The atmospheric circulations on Saturn and Jupiter are dominated by numerous fast east-west (zonal) jet streams, which interact with a wealth of vortices, waves, turbulent filamentary structures, convective storms, and other features in poorly understood ways. The jet profiles exhibit qualitative similarities—on both planets, there exists a broad, fast prograde (eastward) equatorial jet and numerous narrower, weaker jets at higher latitudes. Nevertheless, the zonal jets are considerably broader, faster, and fewer in number on Saturn than Jupiter—peak wind speeds reach $\sim 500 \text{ m s}^{-1}$ on the former but never exceed $200 \text{ m s}^{-1}$ on the latter. Both planets exhibit cloud banding that is modulated by the zonal jets and, at large scales, is far more zonally symmetric than the cloud pattern on Earth. Yet the detailed relationship of the cloud-band structure to the zonal jet structure differs considerably between Jupiter and Saturn. Moreover, despite the existence of many dozens of vortices on both planets, Saturn lacks prominent large vortices like Jupiter’s Great Red Spot and White Oval. These differences remain poorly understood. As case studies in similarities and differences, comparative investigations of the two planets therefore have great potential for insights. Motivations for studying the atmospheric dynamics of Jupiter and Saturn are several. Our understanding of Earth’s
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On Saturn the winds blow primarily east-west and are organized into alternating jet streams, each peaking at its particular latitude. At large scales, the cloud structure is zonally banded—to a much stronger degree than on Earth—but also contains numerous vortices, waves, storms, and various cloud features at small scales (Figure 11.1). Viewed from a non-rotating reference frame, cloud features near the equator exhibit periods of about 10 h 10 min. At latitudes of 35°–40° in each hemisphere, the recurrence period is about 10 h 40 min, and at higher latitude the periods vary within this range up to the pole.

These observations indicate that rotation plays a crucial role in the atmospheric dynamics of Saturn. The importance of rotation can be characterized using the Rossby number, which is the ratio of the advection force to the Coriolis force in the horizontal momentum equation. The advection force per mass has characteristic magnitude $U^2/L$, and the horizontal Coriolis force per mass has characteristic magnitude $fU$, where $U$ is the characteristic wind speed, $L$ is the characteristic horizontal length scale of the circulation (e.g., the jet width), $f = 2\Omega \sin \phi$ is the Coriolis parameter, $\Omega$ is the planetary rotation rate (2\pi over the rotation period), and $\phi$ is latitude. Thus, the Rossby number is given by $Ro = U/fL$. The cloud measurements described above suggest a characteristic rotation period of order 10.5 hours, with zonal speed differences between the cloud bands of typically $\sim 100 \text{ m s}^{-1}$ (Figure 11.1 and 11.2). Inserting appropriate values of $f \approx 2 \times 10^{-4} \text{ s}^{-1}$, $U \approx 100 \text{ m s}^{-1}$, and $L \approx 10^7 \text{ m}$ (relevant to Saturn’s zonal jets), we obtain $Ro \approx 0.05$. Thus, on Saturn—as well as on Jupiter, Uranus, Neptune, and Earth’s atmosphere and oceans—the large-scale circulation exhibits small Rossby number. This estimate implies that, for the large-scale flows on the giant planets, advective forces are weak compared to the Coriolis force. Since the pressure-gradient force is the only other significant force in the horizontal momentum equation, the smallness of $Ro$ implies that the dominant balance in the horizontal momentum equation is between the Coriolis force and the pressure-gradient force—called geostrophic balance (see, e.g., Holton and Hakim 2013, Chapter 2).

Although the observed differential rotation is evidence of wind, we do not know the precise wind speeds relative to the planetary interior—simply because Saturn’s interior rotation rate is not well determined. The other giant planets have internally generated magnetic fields that are tilted with respect to the rotation axis, and those fields are presumably locked to the electrically conducting fluid interiors, at least on time scales of years to decades, so the daily wobble of the field gives us a good estimate of the internal rotation period. Saturn has an internal field as well, but it is not tilted. Thus far, the instruments on Cassini and other spacecraft have been unable to detect a wobble and therefore unable to provide an exact period from which to measure the winds.

Voyager flew past Saturn in 1980 and measured magnetic...
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Fig. 11.1.— Cassini images of Saturn’s southern hemisphere. Left: A cylindrical projection mosaic comprised of Cassini ISS images using the MT2 filter, centered in a methane absorption band at 727 nm wavelength and showing the equator to \( \sim 85^\circ \) latitude and \( 180^\circ - 360^\circ \) longitude. The image is overlain with the zonal-wind profile obtained from cloud tracking (plotted such that \( 0 \, \text{m s}^{-1} \) in System III is at the center of the panel). Right: Polar stereographic projection of Cassini ISS images in the CB2 filter, which is centered at 750 nm, in the continuum between methane absorption bands. From Vasavada et al. (2006).

Fig. 11.2.— Mean zonal wind profile for the 2004–2009 time period. The black curve is from clear filters that sense clouds in the 350–700 mbar pressure range. The red curve is from methane band filters that sense clouds in the 60–250 mbar pressure range. From García-Melendo et al. (2011).
rotation period. Anderson and Schubert (2007) and Read et al. (2009b) used theoretical arguments combined with the observed wind field to suggest rotation periods that are 5–6 minutes shorter than the System III period, and Helld et al. (2015) used the measured low-order gravitational coefficients and the shape of the planet to suggest a rotation period of 10 h 32 min 45 s ± 46 s. If these estimates are correct, then the winds measured relative to the interior would be shifted westward (by a latitude-dependent value) relative to those plotted in Figure 11.2.

On an oblate planet like Saturn, where the rotational flattening \((1 - R_p/R_e)\) is 9.8%, where \(R_p\) and \(R_e\) are the polar and equatorial radii, there are two ways to define the latitude. Planetocentric latitude \(\phi_{pc}\) is the angle of a line from the center of the planet relative to the equatorial plane. Planetographic latitude \(\phi_{pg}\) is the angle of the local vertical relative to the equatorial plane. If we approximate the planet’s shape, e.g., on a constant-pressure surface, as an ellipse rotated about its short axis, then the relation between \(\phi_{pg}\) and \(\phi_{pc}\) is \(\tan \phi_{pg} = (R_e/R_p)^2 \tan \phi_{pc}\). Thus \(|\phi_{pg}|\) is always greater than \(|\phi_{pc}|\) except at the poles and equator, where they are equal. Both kinds of latitude are used in the literature and this review.

### 11.2.1. Zonal Velocity

Figure 11.2, from García-Melendo et al. (2011), shows the zonal mean zonal wind as a function of planetocentric latitude, measured in the System III reference frame. The curves are time averages from 2004 to 2009, but at high and mid latitudes the variations from year to year are generally less than the uncertainty of the measurement. Departures from the zonal mean—the eddies—do not show up on this figure. They have been averaged out, but at least on a large scale the eddy winds are small.

The two curves in Figure 11.2 refer to different altitudes. In both cases the winds are obtained by tracking small clouds in images separated by one Saturn rotation period. The black curve uses images in the CB2 and CB3 filters, centered at 752 and 959 nm, where the main sources of opacity are the clouds themselves. Generally this bandpass senses clouds at the 350–700 mbar pressure range, but clouds outside this range may show up at certain times and places. The red curve uses images in the MT2 and MT3 filters, centered at 727 and 890 nm, where methane, one of the well-mixed gases in the atmosphere, limits the depth one can see. Thus the red curve shows the motion of clouds and haze at the 60–250 mbar pressure range, depending on latitude. This range spans the tropopause—the temperature minimum where the pressure scale height (the e-folding scale) is of order 30 km, so the clouds contributing to the red curve are \(\sim 40\) km higher than those contributing to the black curve.

One notable feature of the zonal wind profile is the high speed in the equatorial band—at least 360 m s\(^{-1}\) faster than the minimum speeds at higher latitudes. These differential speeds are two times larger than those on Jupiter and are 7–8 times larger than the difference between the easterlies and westerlies on Earth. Nevertheless, they are comparable to the speed differential on Uranus, and less than that on Neptune. These differences are not well understood.

The four high-latitude eastward jets, separated by four high-latitude westward jets (eastward jet minima) in each hemisphere are another notable feature. Both in terms of their speeds and their numbers, these jets are more like Jupiter’s jets, although the latter are somewhat more numerous and somewhat less speedy. Away from the equator, the winds are remarkably steady in time and remarkably constant in altitude. There is a general tendency for the profiles at the zonal jet minima to be more rounded than the profiles at the zonal jet maxima, which are sharper. The zonal wind must go to zero at the pole, and Figure 11.2 hints at the extremely rapid drop in winds speed within one degree of the south pole. This is the south polar vortex, like the eye of a hurricane, with 150 m s\(^{-1}\) winds circling around the pole at \(-89^\circ\) latitude (Dyudina et al. 2008; O’Neill et al. 2015, 2016). The circulation direction is cyclonic, meaning that it is in the same direction that the planet is rotating as seen looking down from above. In the southern hemisphere a cyclonic vortex is clockwise. A cyclonic vortex exists at the north pole as well (Antuñano et al. 2015), although Figure 11.2 doesn’t show it because the north pole was in darkness when the data were taken. Hurricanes on Earth are also cyclonic, although they form in the subtropics and drift around, unlike the polar vortices on Saturn (Dyudina et al. 2008). See the chapter by Sayanagi et al. on polar phenomena.

The final notable feature is the vertical variation near the equator. From 10° to 25° in each hemisphere, the wind is slightly stronger at the higher altitude. From 2° to 10°, the high-altitude winds are weaker, and within the narrow jet inside \(\pm 2^\circ\), the high-altitude winds are stronger. The equatorial winds are also variable in time. García-Melendo et al. (2011) point out that the equatorial winds were 450 m s\(^{-1}\) at the time of the Voyager encounters in 1980 and 1981 (Sanchez-Lavega et al. 2000), and ranged up to 420 m s\(^{-1}\) as measured by Hubble in 1990 (Barnet et al. 1992). García-Melendo et al. (2011) conclude that there was a real slowdown of the wind speed at the equator. Nevertheless, Choi et al. (2009) showed by tracking features in 5-\(\mu\)m images from Cassini VIMS—which sense as deep as \(\sim 2\) bars—that the near-equatorial winds reach speeds of \(\sim 450\) m s\(^{-1}\), similar to the Voyager wind measurements. This suggests that any real slowdown in the winds may have been confined to the lower-pressure levels of the upper troposphere (significantly above the level where VIMS 5-\(\mu\)m images sense). Moreover, a stratospheric oscillation in temperature has been observed from Earth-based telescopes since 1980 (Orton et al. 2008) and implies an oscillation in the winds via the thermal-wind equation (see Section 11.3), which may have some connection to the oscillation in equatorial jet speeds seen in Figure 11.2.
11.2.2. Clouds and Temperatures

The clouds that we see in the clear filters CB2 and CB3 are thought to be crystals of ammonia. From spectroscopic observations, we know that ammonia vapor is close to saturation at these levels, so it is likely that ammonia is condensing. Solid particles do not give as clear a spectroscopic signature as the vapor, so it is difficult to say what other substances are present in the cloud particles. What condenses below the ammonia cloud base depends on composition and temperature, and both are uncertain. One must rely on models based on plausible assumptions, which we now describe.

Figure 11.3, from Atreya (2006), is a cartoon showing a possible cloud structure for Saturn for three different assumptions about the mixing ratios of condensable gases relative to hydrogen and helium, which are the major constituents. One takes the ratios O/H, N/H, S/H, and Ar/H on the Sun, determined spectroscopically, and multiplies (enriches) them all by a single factor, either 1, 5, or 10. Then one allows the reactive elements to combine with hydrogen to form H₂O, NH₃, and H₂S. The Ar and He stay in atomic form, and the remaining hydrogen becomes H₂. Even with 10-fold enrichment, the minor constituents make up less than 2% of the molecules in the gas. One assumes the atmosphere is convecting up to the tops of the ammonia clouds, which means temperature $T$ and pressure $p$ follow an adiabat—a moist adiabat in this case to take into account the latent heat released when the vapors condense. The particular adiabat is chosen to match the observed $T$ and $p$ at the top of the clouds, which is about as deep as we can see with remote sensing instruments. In the figure the clouds are labeled and color-coded by their composition, and $T$, $p$ values are given along the sides. For a rising parcel, the less volatile substance, water, condenses out first—at higher temperatures, and the more volatile substance, ammonia, condenses out last—at lower temperatures. The location of cloud base is fairly certain provided the composition is known. The cloud density is much less certain, since it assumes that none of the condensate falls out of the cloud and none is carried upward from where it condenses. Despite these uncertainties, the three-cloud structure is generally accepted. Enrichment factors up to 10 times the solar abundances are supported by the C/H ratio implied by the methane abundance (Fletcher et al., 2009, 2012). Methane, which doesn’t condense at Saturn atmospheric temperatures, is easier to measure above the clouds by remote sensing and is therefore a good measure for the planet as a whole.

Temperature profiles obtained from radio occultations and infrared spectra are shown in Figure 11.4. They show that the temperature is close to an adiabat below the 300-400 mbar level, presumably indicating that convection is occurring, but becomes significantly stratified at higher levels (Li et al. 2013). The two techniques agree reasonably well, and can only sense to levels near ~1 bar, below which we have few observational constraints. The temperatures are fairly steady deeper than the ~300-mbar level, but exhibit greater spatial and temporal variations in the stratosphere.

11.2.3. Inferences From Dynamical Balance

Given the observations of zonal winds at cloud level and temperatures in the overlying layers, basic dynamical arguments can be used to infer the zonal winds above the cloud deck. On planets that rotate rapidly, with small Rossby number, there exists a dynamical link between winds and temperatures. Specifically, combining geostrophic balance, hydrostatic equilibrium, and the ideal-gas law leads to the thermal-wind equation for a shallow atmosphere (e.g., Holton and Hakim 2013, p. 82). In the meridional direction, this reads:

$$\frac{\partial u}{\partial \log p} = \frac{R}{f} \left( \frac{\partial T}{\partial y} \right)_p$$  \hspace{1cm} (11.1)

where $u$ is the zonal wind, $p$ is pressure, $R$ is the specific gas constant, $f = 2\Omega \sin \phi$ is the Coriolis parameter, $\Omega$ is the planetary rotation rate, $\phi$ is latitude, $T$ is temperature, $y$ is northward distance, and the derivative on the righthand side is taken at constant pressure. This equation can only be applied away from the equator since geostrophy breaks down at the equator. The equation implies that, away from
the equator, vertical gradients of the zonal wind are proportional to meridional temperature gradients. Given the known winds at the cloud level from cloud tracking (Figure 11.2), and given observations of temperature and its meridional gradient in the layers above the clouds, we can therefore integrate Equation (11.1) to estimate the zonal winds above the clouds.

The top panel of Figure 11.5, from Read et al. (2009a), shows temperatures derived from Composite Infrared Spectrometer (CIRS) observations from Cassini (Fletcher et al. 2007, 2008), with an interpolation between 6 and 35 mbar where the Cassini CIRS instrument has no spectral sensitivity. The axes are latitude and log-pressure, with pressure increasing downwards. The contours show that the temperature increases upward above the 80 mbar level. The lower panel of Figure 11.5 depicts the zonal-mean zonal wind inferred for Saturn’s upper troposphere and lower stratosphere by integrating the thermal-wind equation (11.1). The integration adopts as a lower boundary condition the zonal-mean zonal winds obtained from visually tracking clouds in the 350-700 mbar pressure range. At pressures greater than several mbar, the temperatures are correlated with the winds such that the equatorward flanks of eastward jets—the anticyclonic regions—are colder than the surrounding regions (on an isobar), and the poleward flanks of eastward jets—the cyclonic regions—are warmer than the surrounding regions. Given the thermal-wind equation, this correlation implies that the change of zonal wind with altitude is opposite to the direction of the wind—the winds are decaying with altitude. Figure 11.5 indicates that they decay to an average speed of about $40 \text{ m s}^{-1}$, which is the average speed of the stratosphere at the 1-2 mbar level.

This result is not new, but it has not been fully explained. The IRIS instrument on Voyager observed the same correlation of thermal gradients with the zonal jets during the encounters with Jupiter (e.g., Conrath and Pirraglia 1983; Gierasch et al. 1986). The problem is that there is no obvious radiative or thermodynamic heat source that would produce temperature gradients that correlate with the jets. The IRIS team therefore proposed a mechanical origin. They postulated that, in the upper troposphere above the clouds, the net zonal acceleration due to eddies acts as a drag force...
in the opposite direction from the zonal jets. Such a zonal eddy force would—in statistical steady state—be balanced by the Coriolis force acting on the meridional wind. In other words, the postulated eddy accelerations would induce a meridional circulation.

Specifically, this force balance predicts that, for the Coriolis acceleration to balance the eddy acceleration, the resulting meridional circulation would be poleward across eastward jets and equatorward across westward jets. This configuration implies that the meridional flow converges on the poleward flank of eastward jets, leading to downward motion. Because the upper troposphere is stably stratified (specific entropy increases with altitude), the downwelling advects high-entropy air downward from above, leading to warmer temperature on constant-pressure surfaces. Similarly, the meridional flow diverges on the equatorward flank of eastward jets, leading to upward motion, which advects low-entropy air from below and produces cooler temperatures on constant-pressure surfaces. In other words, this scenario explains the temperature anomalies observed by Voyager and Cassini, and in thermal-wind balance, the resulting zonal jets must decay with altitude. Note that the poleward flank of eastward jets exhibit cyclonic vorticity and the equatorward flanks of eastward jets exhibit anticyclonic vorticity. Based mostly on differences in Jupiter’s cloud colors and morphology, the amateur astronomers called the cyclonic and anticyclonic regions belts and zones, respectively. With a westward eddy acceleration on the zonal jets, the belts are regions of downwelling and the zones are regions of upwelling.

### 11.2.4. Chemical Tracers

Although one cannot measure upwelling and downwelling directly—the velocities are too small—one can detect these motions using chemical tracers. If a substance is removed from the air at high altitudes, either by condensation or by chemical reactions, then finding it above that altitude means that it was brought there by an updraft. Conversely, if a substance is severely depleted below the altitude where the chemical reactions are occurring, then the depleted air was probably brought there by a downdraft. If one knows the rate of the chemical reaction, one can often estimate the speed of the updraft or downdraft.

Figure 11.6, from Fletcher et al. (2011), shows some of the tracers of vertical motion. The data are from the Cassini VIMS instrument and are from spectra in the 4.6–5.1 μm atmospheric window. A window is where the gases of the atmosphere are especially transparent, and photons in this spectral range are sampling the 1–3 bar pressure region. Clouds are the main source of opacity, and the solid and dotted curves are from two different models of the clouds. Ammonia (upper right graph) is generally abundant at low altitude and is depleted by precipitation at high altitude. Abundant ammonia usually means it was brought there by an updraft. Thus the updrafts have high ammonia and the downdrafts have low ammonia. Thus the spike at $|\phi| < 8^\circ$, where $\phi$ is latitude, is a sign of upwelling at the equator. Similarly the troughs at $8 < |\phi| < 20^\circ$ are a sign of down-

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1. For poleward motion, the Coriolis force induces an eastward acceleration, whereas for equatorward motion, it induces a westward acceleration. If there were an eastward jet with a westward eddy-induced acceleration, the Coriolis force must be eastward in order to balance the eddy acceleration. This implies a poleward meridional flow. Likewise, if there were a westward jet with an eastward eddy acceleration, the Coriolis force must be westward to achieve steady state. This implies equatorward meridional flow.

2. In principle, mass conservation requires that a meridional convergence be balanced by a vertical divergence of the flow. Thus, one could theoretically imagine that a meridional convergence could be balanced either by upward motion above the convergence level, or by downward motion below the convergence level (or a combination). Because of the strong stable stratification in the stratosphere, the magnitude of the upward flow above the convergence level should be limited, and thus we expect a significant degree of downward motion below the convergence level. Analogous arguments lead to the conclusion that upward motion should occur below locations of meridional flow divergence.

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Fig. 11.5.—Top: Saturn’s zonal-mean temperature versus latitude and pressure from Cassini CIRS nadir-viewing observations taken in 2004-2006, during Saturn’s northern spring. Analysis is from Fletcher et al. (2007). Note the temperature minimum around 80 mbar and the increase in temperatures toward the south pole at all levels above the 6-mbar level. The instrument is not sensitive to the regions of meridional flow divergence. The dotted curves are from two different models of the clouds. Ammonia (upper right graph) is generally abundant at low altitude and is depleted by precipitation at high altitude. Abundant ammonia usually means it was brought there by an updraft. Thus the updrafts have high ammonia and the downdrafts have low ammonia. Thus the spike at $|\phi| < 8^\circ$, where $\phi$ is latitude, is a sign of upwelling at the equator. Similarly the troughs at $8 < |\phi| < 20^\circ$ are a sign of down-
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Fig. 11.6.— Tracers of vertical motion. The graphs are derived from 5-µm spectra obtained by the Cassini VIMS instrument, which senses thermal emission from the 1–3 bar pressure range. The solid and dotted curves employ two different assumptions about cloud opacity. Although the optical depth of the deep cloud is relatively constant, the cloud base pressure is lower (cloud height is greater), which is suggestive of equatorial upwelling. The higher NH$_3$ mole fraction out to latitudes of ±8° is suggestive of upwelling, and the troughs at ±(8 − 20)° are suggestive of downwelling. The AsH$_3$ mole fraction shows a distribution that is almost opposite to that of NH$_3$, which is not fully understood. One possible explanation is that the pattern of upwelling and downwelling reverses at some level and the two gases are reflecting that reversal. From Fletcher et al. (2011).

welling. The broad downwelling in the cyclonic regions on either side of the equator is consistent with the expected convergence of meridional flow into a region with an eastward jet on the equatorward side and a westward jet on the poleward side.

Figure 11.7, from Janssen et al. (2013), tells a similar story. The data show thermal emission at 2.2-cm wavelength, in the microwave region, and were taken by the Cassini radar instrument acting as a radiometer. Ammonia vapor is the main source of opacity at this wavelength—clouds are unimportant—so the bright areas are places of reduced ammonia abundance, which allows radiation from the warmer, deeper levels to escape into space. The depleted bands within ±10° of the equator are evidence of downwelling. The black band on the equator is an image of Saturn’s rings, which are colder than the planet. The dotted line gives the time of ring plane crossing, and the dashed line gives the time of the closest approach of the spacecraft to the planet. As the spacecraft moved in and out of the ring plane, which it did on July 24, 2010, the rings covered a part of the opposite hemisphere, affording a brief look at the equator. The equator itself is not so bright, and is consistent with ammonia abundance near the saturation value, as it is over most of the planet outside the subtropical (±10°) band. Within that band, the ammonia must be depleted down to 1.5 bars to fit the high brightness temperatures (Laraia et al. 2013).

The distribution of NH$_3$ on Saturn resembles the distribution of H$_2$O on Earth. On both planets there is rising in the tropics and sinking in the subtropics, although on Earth the subtropics extend further out, to ±30°. On Earth the circulation is called the Hadley cell, and the band of rising motion at the equator is called the Intertropical Convergence Zone (ITCZ), a feature of which are the high cumulus clouds associated with deep convection. The clouds of Saturn show a similar pattern (lower left of Figure 11.6) in which the base pressure of the deep cloud is less, indicating displacement to higher altitude. However the Hadley cell analogy may be misleading, because Earth has equatorial easterlies (westward winds) and Saturn has equatorial westerlies (eastward winds). The eddy sources and the configuration of zonal accelerations they induce may therefore differ significantly between the two planets.

Gases other than ammonia tell a somewhat different story, and the differences are not fully understood. Figure 11.6 shows that the latitude distribution of arsine, AsH$_3$ is almost the opposite of the NH$_3$ distribution. Arsine has a dip at the equator and broad peaks on either side of the
equator out to $\pm 20^\circ$. One of the cloud models shows an extension of the peak to a latitude of $-30^\circ$, but the other model does not. The distribution of phosphine, PH$_3$ (not shown), is a lot like that of arsine. Degeneracies between the retrievals of the cloud properties and the abundances of trace species like AsH$_3$ can occur (e.g. Giles et al. 2017), so the AsH$_3$ results shown in Figure 11.6 should perhaps be viewed as tentative until more analysis has been performed; nevertheless, the different cloud models explored by Fletcher et al. (2011) all pointed toward an equatorial dip in AsH$_3$ relative to the surrounding latitudes, suggesting that it should be taken seriously. Both arsine and phosphine are in chemical equilibrium at much deeper levels than those probed in the 4.6–5.1 $\mu$m data, and they are destroyed by photolysis at high altitudes. That they don’t show the same patterns as NH$_3$ has interesting implications.

Fletcher et al. (2011) discuss the possibility of stacked meridional circulation cells, the top one with rising air at the equator, as in Earth’s Hadley circulation, and the bottom one—the reverse cell—with sinking at the equator. The NH$_3$ cloud base is around 1.5 bars, as shown in Figure 11.3, so its mixing ratio should be large up to that level in an up-
draft. Above that level it loses ammonia by condensation.
A downdraft could carry ammonia-depleted air down be-
low the 1.5-bar level until it eventually mixes with air from
the planet’s interior. Thus the NH$_3$ distribution shown in
Figure 11.6 is consistent with an Earth-like Hadley cell ex-
tending down to at least the ~2-bar level. The AsH$_3$ dis-
tribution is consistent with a reverse Hadley cell since its
chemical transformations are taking place at much deeper
levels. The base pressure of the deep cloud, shown in the
lower left panel of Figure 11.6, is about 2.0 bars at the equa-
tor and 2.8 bars on either side of the equator. This could be
interpreted in two ways. If the cloud particles are NH$_3$SH or
H$_2$O, and have been carried up from deeper levels, then the
higher altitude (lower base pressure) at the equator would
signify an updraft. However, if the cloud base pressure were
entirely dependent on the abundance of a condensing gas,
then the lower base pressure would signify a lower abun-
dance of the condensing gas and hence a downdraft. The
latter would be consistent with a reverse cell, with sinking
motion at the equator. Note however that the upper cell is
consistent with a meridional circulation that is driven by
an eddy acceleration acting in the opposite direction as the
zonal winds in the upper troposphere. The lower meridional
cell would seem to require forces that accelerate the zonal
winds, and such forces do exist, as we now describe.

11.2.5. Eddies And The Momentum Budget

The above analysis takes the zonal winds at cloud top
(Figure 11.2) as given. The eddy-induced acceleration op-
posing the jet direction was postulated to account for the
wind’s decay with height above the cloud-top level. In con-
trast, observations demonstrate that at the cloud level, ed-
dies transport momentum that acts to drive the jets. Here
we describe these observations and their interpretation.

By themselves, the zonal-mean of the eddy winds aver-
age out to zero, but the eddies can have a net effect when
the northward and eastward components act together. Con-
sider a latitude band with an eastward jet to the north and
a westward jet to the south. The eddy winds, with compo-
nents $u'$ and $v'$, induce motion of air parcels north and south
in the space between the two jets. (Here, as before, $u$ is the
zonal wind, and $v$ is the meridional wind; primes denote
deviations from the zonal average.) But if the parcels going
north have more eastward momentum than the parcels go-
ing south, there will be a net transfer of momentum to the
north. This would add to the eastward momentum of the
eastward jet and it would subtract it from the westward jet,
speeding up each jet in its respective direction.

Such a hypothesis is testable if one has good measure-
ments of the eddy velocities. Parcels moving north ($v' > 0$)
would also be moving east ($u' > 0$), and parcels mov-
ing south ($v' < 0$) would also be moving west ($u' < 0$).
In both cases the parcel trajectories would be tilted along
a northeast-southwest line, and the measured eddy winds
would have $w'v' > 0$, where the overbar denotes the zonal
mean. The eddy-momentum flux is $\rho w'v'$, where $\rho$ is the
density. The eddy-momentum flux is a stress, and it has
units of force per unit area. A positive tilt ($u'v' > 0$) im-
pies a northward transport of eastward momentum by the
eddies. A tilt the other way ($u'v' < 0$) would represent
a negative eddy-momentum flux—a northward transport
of westward momentum by the eddies. In either case, if $u'v'$
and $\partial \bar{u}/\partial y$ have the same sign, where $\bar{u}$ is the mean zonal
wind and $y$ is the northward coordinate, then the eddies are
adding momentum to the zonal jets.

Figure 11.8, from Del Genio and Barbara (2012), shows
the sign of the correlation. At almost all latitudes, $u'v'$ and
$\partial \bar{u}/\partial y$ have the same sign. Both quantities are positive
at latitudes 50-57N, 33-42N, 10-25S, 28-35S, and 45-48S.
Both quantities are negative at latitudes 42-50N, 10-33N,
35-40S, and 48-52S. The data are noisy because the eddy
winds are weak—only a few $\text{m s}^{-1}$, but the data clearly
indicate that Saturn’s eddies are acting as a force that ac-
celerates the zonal jets. Both Voyager and Cassini observed
similar behavior at Jupiter (Beebe et al. 1980; Ingersoll et al.
1981; Salyk et al. 2006).

This sounds like negative viscosity, and indeed that term
was used to describe such phenomena, which are observed
not only in Earth’s atmosphere but also the oceans, the Sun,
and laboratory experiments (Starr 1968). However, using an
eddy viscosity to relate a local stress like $\rho w'v'$ to a lo-
cal rate of strain like $\partial \bar{u}/\partial y$ is often a meaningless exer-
cise (Phillips 1969). Energy transfer from smaller to larger
scales does not violate any thermodynamic principle, and
an eddy-momentum transfer that generates a shear flow—
zonal jets—can arise naturally through the interaction of
turbulence with the planetary rotation. Since the eddies are
putting energy into the zonal flow, they must have their own
source of energy. That could come from below, as inter-
nal heat power convection currents that rise into the cloud
layer. Or it could come from the sides, as lateral tempera-
ture gradients release their potential energy into a longitudi-
nally varying wave in a process called baroclinic instability
(e.g., Vallis 2006; Holton and Hakim 2013).

This eddy force at cloud level, which acts to acceler-
ate the zonal jets, is opposite in sign to that postulated in
the upper troposphere that tends to decelerate the jets, and
it has the opposite effect on the meridional overturning.
This reversal in sign of the eddy acceleration with height
could lead to stacked meridional circulation cells, the eddy-
momentum forces that oppose the jets driving the upper cell
and the eddy-momentum forces that accelerate the jets driv-
ing the lower cell. Neither of these meridional cells how-
ever, is like the Earth’s Hadley circulation (see Vallis 2006
or Schneider 2006 for reviews). On Earth, the eddy accelera-
tions in the subtropics are westward, whereas the eddy ac-
celerations occurring within Saturn’s equatorial jet are east-
ward, leading to eddies driving a reverse meridional cell on
Saturn. The direct circulation cell on Saturn, which seems
to start at ~2 bars and extend into the stratosphere (Fig-
ure 11.5), is driven by an eddy acceleration opposing the
zonal jets, which is opposite to the eddy momentum force
at deeper levels.
Fig. 11.8.— Eddy momentum transport and zonal winds. The left panel shows the mean zonal wind profile $\overline{u}$ obtained by tracking individual cloud features. The right panel shows the number of measured wind vectors in each $1^\circ$ bin of latitude. The middle panel shows the covariance $\overline{u'v'}$ between the eastward eddy wind $u'$ and the northward eddy wind $v'$, where an eddy is defined as the residual after the mean zonal wind has been subtracted off. Multiplied by the density, this covariance is the eddy momentum transport—the northward transport of eastward momentum by the eddies. This transport tends to have the same sign as $\partial \overline{u}/\partial y$, where $y$ is the northward coordinate, which indicates that the eddies are putting energy into the mean zonal wind and not the reverse. From Del Genio and Barbara (2012).

The eddy-momentum flux driving the jets at cloud level has been observed directly on Jupiter and Saturn (Beebe et al. 1980; Ingersoll et al. 1981; Salyk et al. 2006; Del Genio and Barbara 2012). In contrast, there is no direct observational confirmation of the eddy acceleration above the clouds that is postulated to oppose the jets. The term eddy includes all motions that remain after the zonal mean has been subtracted off—waves, vortices, convective plumes and any other non-axisymmetric component of the motion. Evidence for the eddy force opposing the jets comes from the decay of the zonal winds with height as inferred from temperature gradients (Figure 11.5). Evidence for a reverse meridional circulation cell comes from the AsH$_3$ and PH$_3$ distributions, but further evidence comes from the distribution of lightning, as we discuss below.

11.2.6. Lightning And Moist Convection

Lightning is evidence of moist convection. The violent updrafts in a thunderstorm are a crucial element for electrical charge separation, which occurs when large ice particles fall through upwelling air that contains small liquid droplets (Uman 2001; MacGorman and Rust 1998). The charging mechanism is not well understood, but three phases of water—solid, liquid, and vapor—in the cloud at once seem to be necessary. On giant planets, with their three-tiered structure (Figure 11.3), it is not immediately clear which cloud produces the lightning (Yair et al. 2008). For a solar composition atmosphere, the mole fractions of H$_2$O, NH$_3$, and H$_2$S, in units of $10^{-4}$ are 9.7, 1.3, and 0.29, respectively (these values depend on the abundance of helium, which is poorly known; here we have used the latest He/H$_2$ estimate from Sromovsky et al. 2016). For Saturn, enrichment by a factor of 10 is likely (Fletcher et al. 2009; Fletcher et al. 2012), and uniform enrichment would raise all these numbers by the same factor. Thus on the basis of abundance alone, water is the condensible gas most likely to produce lightning. Also, water is the only gas where the temperatures and abundances allow three phases to exist at the same
level in the atmosphere. For Saturn the triple point of water occurs at a pressure of about 10 bars (Figure 11.3). This number depends on the temperature profile, but is relatively independent of the enrichment factor. Increasing the latter mostly affects the depth of the liquid cloud, extending it down to \( p = 20 \) bars and \( T = 330 \) K for 10 times solar abundances.

Figure 11.9, from Dyudina et al. (2013), shows the brightness distribution of a typical lightning flash as seen from the Cassini spacecraft. Every pixel in the vicinity of the flash is plotted as a function of its horizontal distance from the flash center. This removes the effects of foreshortening and shows that the flashes are roughly circular. The half width at half maximum (HWHM) of the brightness distribution is about 100 km, and that is related to the depth of the lightning relative to the tops of the clouds—the level where the photons emerge. That depth is approximately equal to 1.5 times the HWHM. Cloud parameters—both their height and their scattering properties—create a large uncertainty. Dyudina et al. (2013) estimate that the tops are probably \( \text{NH}_3 \) or \( \text{NH}_4 \text{SH} \) clouds at depths exceeding 1.2 bars, in which case the lightning is in the water cloud.

Nighttime imaging of Jupiter by the Galileo spacecraft showed that lightning is concentrated in the belts (Little et al. 1999; Gierasch et al. 2000), and that was a surprise. The traditional view, which was based on Voyager observations of the upper troposphere (Gierasch et al. 1986), was that the belts are regions of descent and the zones are regions of ascent, for several reasons: The zones exhibit a thick, bright, relatively uniform cloud deck, whereas the belts exhibit patchier, generally thinner clouds. The ammonia mixing ratios at and above the clouds are small in the belts and large in the zones, signifying descent and ascent, respectively. The belts are warm, signifying high-entropy air brought down from above, and the zones are relatively cold. Lighting therefore seemed unlikely in the belts, since moist convection requires moist air brought in from below. This led Ingersoll et al. (2000) to postulate that perhaps the belts have upwelling at the base of the cloud and downwelling at the tops. This amounted to a pair of meridional circulation cells turning in opposite directions, one on top of the other. Showman and de Pater (2005) showed that this stacked-cell scenario also best explains the deep ammonia abundances from 1–5 bars, which indicates that belts and zones are both depleted in ammonia relative to the presumed deeper atmosphere. According to this new picture, the traditional view was not wrong, but it was based only on the properties of the upper circulation cell. Lightning and ammonia had provided a view to deeper levels, at least on Jupiter. The data for Saturn are less clear, partly because lightning is such a rare event on that planet.

Figure 11.10, from Dyudina et al. (2007), shows light-
ning intensity over a 2-year period from 2004 to 2006 as recorded by the Radio and Plasma Wave Science (RPWS) instrument on Cassini. The RPWS is a radio receiver and spectrometer, and it is “on” all the time. The figure shows that there was lightning activity on days 200–270 in 2004, then a 445-day gap with no lightning, followed by activity on days 350–385 in 2005. Only one storm was active during each of the two periods, and it might have been the same storm, as indicated by the drift rate in longitude shown on the bottom part of the figure. The storm was located at −35° planetocentric latitude, at the center of the westward jet shown in Figure 11.2. This latitude band became known as “storm alley” during the first few years of the Cassini mission. The storm was a multi-armed complex, 2000 km in diameter, which spawned smaller spots at a rate of one every 1–2 days. These drifted off to the west relative to the central complex. The smaller spots began with high, thick clouds that dissipated in the first day, leaving a dark spot that consisted either of a hole in the clouds or else a deep cloud with a low reflectivity (Dyudina et al. 2007). The small bright spot in the lower left corner of Figure 11.7, the image taken on March 20, 2011, is also at −35° latitude. Its brightness at 2.2-cm wavelength indicates that it is a region of low abundance of ammonia vapor, which is consistent with a hole in the clouds. The latitude band at the center of the westward jet at −35° seems to have been active over the 10+ years of the Cassini orbital mission.

During the Cassini mission, the northern hemisphere did not have a lightning storm until December 5, 2010, when the RPWS suddenly started detecting lightning signals and the imaging system detected a new cloud at 35° latitude. As in the southern hemisphere, the latitude of the storm coincided with the center of a westward jet (Figure 11.2). However, the northern storm was a different beast from those in storm alley. Its head remained the most active part, but by late January its tail had spread around the planet to the east late January its tail had spread around the planet to the east and was encountering the head from the west (Fischer et al. 2011; Sánchez-Lavega et al. 2011). The tail was depleted in ammonia, as seen in the March 20, 2011 map at 2.2 cm (Figure 11.7). The head spawned a series of large anticyclonic vortices, 10,000 km in north-south dimension, which also drifted to the east, but more slowly than the debris in the tail. When the first of these—also the largest—encountered the head in June 2011, the head broke up and essentially ceased to exist (Sánchez-Lavega et al. 2012; Sayanagi et al. 2013). The chapter by Sanchez-Lavega et al. in this book is devoted to the giant storm, and we will only discuss a few features that are especially relevant to this dynamics chapter.

Figure 11.11, from Achterberg et al. (2014), shows changes from 2009 to 2011, before and after the storm, using infrared data from the Cassini CIRS instrument. Figures 11.11a and 11.11b show that the storm warmed the latitude band from 30–45°. The warming also showed up as an increase in the emitted power at that latitude (Li et al. 2015). Figures 11.11c and 11.11d show evidence of gas having been brought up from the level of the water cloud, as inferred from the relative abundance of the two states of the H₂ molecule. The observed warming made ∂T/∂y more negative on the north side of the disturbed band and more positive on the south side. The thermal-wind equation then says there should have been an increase in ∂π/∂z on the north side and a decrease on the south side. If we are seeing the tops of the clouds, and if π is constant underneath, we should see an increase in π to the north and a decrease to the south, and that is what was observed (Sayanagi et al. 2013).

Lightning on Jupiter was seen mostly in the belts, which are the regions of cyclonic shear (Little et al. 1999; Gierasch et al. 2000). Lightning on Saturn was seen near the centers of the westward jets, which have cyclonic shear on the equatorward side and anticyclonic shear on the poleward side. Thus the two planets appear to be different. However, Dyudina et al. (2013) noted that within the giant storm there were small regions of cyclonic shear, and they were where the lightning occurred. The giant storm consisted of an anticyclonic head with a chain of large anticyclones stretching off to the east, each one rolling in a clockwise direction. The narrow regions where they came close together were cyclonic. A cyclonic region has low pressure at the center, with the inward pressure gradient force balancing the outward Coriolis force. In a terrestrial hurricane, this leads to convergent flow in the boundary layer, where friction at the ocean surface slows the wind and weakens the Coriolis force, leading to an unbalanced pressure gradient force acting inward. The inward flow picks up moisture from the ocean surface, leading to intense moist convection around the center. One might think that the same process is happening on the giant planets, since moist convection occurs in the cyclonic regions. But how this might work without an ocean, or a physical surface against which friction can occur, is unclear. The deep atmosphere below the clouds would have to provide a frictional force on the cloud layer that leads to convergence. If the flow below the clouds were sufficiently turbulent, with large vertical stresses, it might do the job. But that theory needs to be worked out.

11.2.7. Stability Of the Zonal Jets

The cloud-top zonal jets on Jupiter and Saturn are remarkably constant in time. Their positions and wind speeds haven’t changed much over 100 years, although small differences between Voyager and Cassini were observed, especially at the equator (Peek 1958; Porco et al. 2003; Vasavada et al. 2006; Li et al. 2013; García-Melendo et al. 2011). This remarkable constancy motivates a consideration of the extent to which the jets are dynamically stable. The definition of a stable flow is one in which infinitesimal perturbations cannot grow in amplitude. There exist a class of theorems that provide information on jet stability for the idealized case of zonally symmetric jets, with no forcing or damping, and several of these have been evaluated for Saturn.

Most stability theorems are cast in terms of the potential vorticity (PV), which, for the case of a single-layer,
shallow-water flow is

\[ PV = \frac{\zeta + f}{h} \]  

(11.2)

where \( \zeta = k \cdot (\nabla \times \mathbf{v}) \) is the relative vorticity, \( \mathbf{v} \) is the fluid velocity, and \( k \) is the vertical unit vector. In this shallow-water form, \( h \) is the depth of the fluid. Under frictionless, adiabatic conditions, the PV is a materially conserved quantity (e.g., Vallis 2006). The inverse dependence on \( h \) reflects the fact that stretching along the rotation axis of the fluid parcel (thereby increasing \( h \)) causes a contraction toward the parcel’s rotation axis and a decrease in the moment of inertia, causing the parcel to spin faster and \( \zeta + f \) to increase. The PV of the parcel is conserved because the numerator and denominator change together. Note that, for a stratified, three-dimensional atmosphere, the PV can also be written in the form (11.2) if one uses the component of vorticity that is perpendicular to surfaces of constant specific entropy, and if one treats \( h \) as a differential mass per unit area between surfaces of constant specific entropy.

The simplest stability theorem, which dates back to Rayleigh, with modification for rotating planets, is the Charney-Stern criterion (Vallis 2006). It says that a zonal flow is stable—infinitesimal disturbances cannot grow—if PV varies monotonically in the cross-stream direction. If we ignore the stretching term by treating \( h \) as a constant, we obtain the barotropic stability criterion, which states that a steady zonal flow \( \overline{\zeta(y)} \) is stable provided

\[ \frac{d(\zeta + f)}{dy} = \beta - \overline{\nabla_{yy}} \geq 0 \]  

(11.3)

where \( \beta = df/dy = 2\Omega \cos \phi/a \) and \( d\overline{\zeta}/dy = -\overline{\nabla_{yy}}. \) Here \( \beta \) is the planetary vorticity gradient, \( \phi \) is latitude and \( a \) is the radius of the planet. Note that \( \beta \) is always positive but it goes to zero at the poles. The subscript \( y \) is a derivative, and \( \overline{\nabla_{yy}} \) is the curvature of the zonal velocity profile, which is positive at the centers of the westward jets (Figure 11.2).

If this curvature is less than \( \beta \) at every latitude, the flow is stable provided we can ignore the stretching term. Jupiter’s zonal jets violate the barotropic stability criterion (Ingersoll et al. 1981; Limaye 1986; Li et al. 2004). Here we describe an application of stability criteria to Saturn, starting with the barotropic stability criterion and moving on to the more general Ertel and quasi-geostrophic versions.

Figure 11.12, from Read et al. (2009a), shows the zonal-mean velocity \( \overline{\zeta(y)} \), the vorticity \( \zeta = -\overline{\nabla_y} \), and the curvature \( \overline{\nabla_{yy}} = -\overline{\zeta_y} \) in the left, center, and right panels, respectively. The smooth curve in the right panel is \( \beta \), which varies with latitude as \( \cos \phi \). One can see that the curvature exceeds \( \beta \) at the latitudes of the westward jets, indicating a violation of the barotropic stability criterion. The violation is more evident near the poles, i.e., the difference \( \beta - \overline{\nabla_{yy}} \) is more negative there, partly because the jets have more curvature there, but also because \( \beta \) is approaching zero.

Read et al. (2009a) also use temperature profiles derived from infrared observations by the Cassini CIRS instrument to estimate the magnitude of the stretching term. They calculate Ertel PV and quasi-geostrophic PV, including the stretching term, as functions of latitude and show that the stretching term has little effect, at least in the upper troposphere and lower stratosphere where the analysis was done. At these altitudes the atmosphere is so stably stratified that the stretching is small. Although Ertel PV is the right quantity to use, the barotropic stability criterion gives the same result, that the centers of the westward jets are where the stability criterion is violated. One might get a different result if one could measure Ertel PV within and below the...
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Fig. 11.12.— Stability of Saturn’s zonal jets. The barotropic stability criterion states that a zonal flow is stable provided the curvature $u_{yy}$ of the zonal velocity profile does not exceed the planetary vorticity gradient $\beta$. Stable means that disturbances to the profile cannot grow, but violation of the criterion does not mean that they will grow. The left panel shows the zonal velocity profile $\Pi$. The middle panel shows the relative vorticity $-u_y$ associated with the zonal flow, where the subscript means differentiation of $\Pi$ with respect to the northward coordinate $y$. The right panel shows the curvature $u_{yy}$ (jagged curve) and $\beta$ (smooth curve, given by $2\Omega \cos \phi / a$), along with the zero-curvature line for reference (straight vertical line). At the latitudes where the curvature is positive, it clearly exceeds $\beta$, indicating that the barotropic stability criterion is violated. The barotropic stability criterion belongs to a more general class of stability theorems that use potential vorticity (PV) as the diagnostic quantity. The rigorous form uses Ertel’s PV, but the results are the same, at least for the upper troposphere where the velocity profile is measured. From Read et al. (2009a).

clouds, but that is not possible with the remote sensing data that we have.

Violation of the barotropic stability criterion or any of the related stability criteria does not mean that the flow is unstable. Further, instability does not mean that the flow cannot exist. The hexagon that surrounds the pole at mean planetocentric latitude of 76° marks the path of a meandering jet that violates the barotropic stability criterion (Antuñano et al. 2015). The jet may be unstable, but the disturbance has grown into a steady finite-amplitude wave. Presumably the wave stopped growing due to non-linear effects that are not included in the linear stability analysis. Ingersoll and Pollard (1982) describe a stability criterion that involves interior flow, in which the zonal winds are aligned on cylinders concentric with the planet’s axis of rotation. Dowling (1995a,b) has argued that Jupiter’s jets could be stable according to Arnol’d’s second stability criterion, which states that the flow will be stable to nonlinear perturbations if there is a reference frame in which the reversals of the PV gradient coincide with reversals in the sign of velocity. Read et al. (2009b) found that the upper tropospheric and stratospheric winds appear to be close to neutral stability according to this criterion; see Dowling (2014) for a detailed discussion. The reference period they derived is ~5 minutes shorter than the System III reference period, and they propose that the period derived from Arnol’d second stability criterion is the rotation period of Saturn’s interior.

11.3 Observations of the Stratosphere

In the absence of visible tracers, the primary source of data constraining the winds and circulation in Saturn’s
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The stratosphere is from observations in the thermal infrared. Observations in the collision-induced hydrogen continuum longward of 16 $\mu$m provide information on the temperatures in the upper troposphere between roughly 50 to 500 mbar. Observations in the $\nu_4$ band of methane near 7.7 $\mu$m provide information on temperatures in the middle and upper stratosphere, between 0.5 and 5 mbar for typical observations, extending to about 0.01 mbar for observations with very high spectral resolution, or limb viewing observations from Cassini. However, these thermal observations have vertical resolution of a pressure scale height, and are thus not sensitive to wave structure on small vertical scales. Temperature profiles with vertical resolution of a few kilometers can be obtained through radio occultation observations, in which the radio signal from a spacecraft is observed as the spacecraft passes behind a planet at seen from Earth, and from stellar occultations. Radio occultations are generally sensitive to the lower stratosphere and upper troposphere between about 0.1 mbar and 1 bar, while stellar occultations generally sense the upper stratosphere or higher.

The first spatially resolved thermal observations of Saturn were made in the $\nu_9$ band of ethane at 12 $\mu$m (Gillett and Orton 1975; Rieke 1975) during southern summer, and showed emission increasing from north to south, peaking at the south pole. Subsequent observations in the $\nu_4$ methane band at 8 $\mu$m (Tokunaga et al. 1978; Sinton et al. 1980) showed similar variations. Since methane is expected to be meridionally mixed, these observations indicated that the emission variations are caused by a north-to-south temperature gradient, with stratospheric temperatures warmest at the south (summer) pole.

The first spacecraft observation in the thermal infrared was made by the Pioneer 11 infrared radiometer (IRR) in 1979, just prior to northern spring equinox. IRR observed in two bands at 20 $\mu$m and 45 $\mu$m, sensitive to the upper troposphere, between 10$^\circ$N and 30$^\circ$S (Ingersoll et al. 1980; Orton and Ingersoll 1980), finding the temperatures within 10$^\circ$ of the equator $\sim$2.5 K colder than higher latitudes. Soon afterward, the Infrared Interferometer Spectrometer (IRIS) on Voyagers 1 (1980) and 2 (1981) mapped Saturn at 5 to 45 $\mu$m. Retrievals of upper tropospheric temperatures (Hanel et al. 1981, 1982; Conrath and Pirraglia 1983) showed no equator to pole temperature gradient in the southern hemisphere, and north polar temperatures roughly 5 K colder than the equator in the upper troposphere at pressures less than about 300 mbar. In addition to the large-scale gradient, the upper tropospheric temperatures showed variations of around 2 K, with the gradient anticorrelated with the zonal winds (Conrath and Pirraglia 1983). A limited set of temperature profiles, covering roughly between 1 bar and 1 mbar, were also obtained by the radio occultation experiments on Pioneer 11 (Kliore et al. 1980) and Voyagers 1 and 2 (Lindal et al. 1985). These profiles showed a broad tropopause with the temperature minimum at about 50 mbar, with mid-stratospheric temperatures of $\sim$140 K at midlatitudes and $\sim$120 K near the equator. The near equatorial profiles also showed small-scale vertical oscillations that were interpreted as vertically propagating waves.

Subsequent ground-based observations taken during northern summer showed enhanced emission at the north pole in the stratospheric methane and ethane bands (Gezari et al. 1989), but not at wavelengths sensitive to the troposphere (Ollivier et al. 2000). These observations, when compared to earlier observations, were interpreted as evidence for seasonal evolution of the temperatures as predicted by radiative models (Cess and Caldwell 1979; Bezard and Gautier 1985), with warm temperatures at the summer pole and the strength of the radiative response weaker at higher pressures where the radiative time constants are longer.

Just prior to the arrival of Cassini at Saturn, Greenhouse et al. (2005) observed Saturn at high spectral resolution in 2002 near southern summer solstice, using the methane $\nu_4$ band to retrieve southern hemisphere temperatures between 0.01 and 10 mbar. They found a general equator to pole temperature gradient, with the south pole 10 K warmer than the equator. In early 2004, Orton and Yanamandra-Fisher (2005) used the Keck telescope to image Saturn in the CH$_4$ and H$_2$ bands, allowing the retrieval of temperatures at 100 mbar and 3 mbar at 3000 km spatial resolution. They also found a 10 K temperature increase between the equator and the summer pole in the stratosphere, with a sharp $\sim$5 K increase between 60$^\circ$S and 74$^\circ$S planetocentric latitude.

The arrival of Cassini at Saturn in mid-2004, near southern mid-summer, began a period of quasi-continuous monitoring of Saturn’s thermal emission by the Cassini Composite Infrared Spectrometer (CIRS) which is planned to continue until northern summer solstice in 2017. Flasar et al. (2005) showed temperature retrievals of the southern hemisphere from data taken during the approach of Cassini to Saturn. They found a 15 K temperature gradient between the equator and south pole at 1 mbar, decreasing with increasing pressure. This temperature gradient is larger than the 5 K expected from radiative models for the season, which Flasar et al. (2005) suggested may be caused by adiabatic heating from subsidence at the pole. The first global temperature cross-sections were made by Fletcher et al. (2007, 2008), using CIRS data to retrieve zonal-mean temperatures between 0.5 and 5 mbar in the stratosphere and 50 to 800 mbar in the troposphere; an updated version of their results is shown in Figure 11.5. At the 1-mbar level,
they found a 30 K temperature difference between the warm southern summer pole and the cold northern winter pole; the temperature gradient is nearly monotonic from pole to pole except for a local temperature maximum at the equator, and a roughly 3 K warming between 80°N and the north pole. At higher pressures, the global temperature gradient becomes weaker, and small-scale variations correlated with the zonal winds appear. CIRS limb viewing observations allowed Fouchet et al. (2008) and Guerlet et al. (2009) to extend temperature retrievals up to the 0.01 mbar pressure level. Surprisingly, they found that the large-scale pole-to-pole temperature gradient was weaker at 0.1 mbar and 0.01 mbar than at 1 mbar, despite the thermal timescales becoming shorter at lower pressures. They also found that the temperature profile oscillates with altitude at equatorial and low latitudes, with equatorial temperature alternately warmer and colder than temperature at ±15°. Long-term monitoring from ground-based observations by Orton et al. (2008) showed that the difference between equatorial and low-latitude temperatures also oscillates in time, with a period near 15 Earth years. This equatorial oscillation will be discussed in more detail in section 11.3.3.

Cassini CIRS has continued to monitor Saturn’s thermal structure through equinox and into northern spring (Fletcher et al. 2010; Fletcher et al. 2015; Guerlet et al. 2011; Sinclair et al. 2013; Sylvestre et al. 2015). Outside of the equatorial region, the seasonal temperature variations are broadly consistent with models (Fletcher et al. 2010; Friedson and Moses 2012; Guerlet et al. 2014), with a warming of the northern hemisphere and cooling of the southern hemisphere. The seasonal variations are discussed in detail in the chapter by Fletcher et al. in this volume.

11.3.1. Zonal Mean Winds

As described in Section 11.2, the meridional temperature gradients observed in the upper troposphere are positively correlated with the measured cloud-top zonal winds at many latitudes. Using the thermal-wind equation (11.1), this result implies that the zonal winds observed at cloud level decay with altitude into the stratosphere, over a vertical scale of about six pressure scale heights. Furthermore, because the eastward and westward wind gradients are of similar magnitude, while the cloud top winds are predominately eastward, the decay of the jets is to a finite eastward velocity and not zero, as measured in the System III reference frame. Further evidence for net eastward stratospheric winds was provided by the analysis of the 1989 occultation of 28 Sgr by Nicholson et al. (1995), who found that winds of around 40 m s⁻¹ were needed near the 2.5 mbar level between 25°N and 70°N to fit the observed timings of the central flash. As can be seen in Figure 11.5, many—though not all—of the off-equatorial jets decay in speed with decreasing pressure. Figure 11.5 also makes plain that the middle stratosphere winds are generally eastward as measured System III, except at high southern latitudes. As mentioned previously, however, the System III reference frame may not represent the true interior rotation rate, and other possible interior rotation rates would imply different values for the overall stratospheric winds (in particular, several proposals for the interior rotation rate are faster than the System III rotation rate, which would imply stratospheric winds weaker—that is, more westward—than shown in Figure 11.5). That said, there is no dynamical reason to expect that the mean zonal velocity in the troposphere or stratosphere is the same as in the deep interior.

Observations of seasonal changes in the zonal mean stratospheric temperatures into early northern spring (Fletcher et al. 2010; Sinclair et al. 2013; Fletcher et al. 2015) show a weakening of the equator-to-pole temperature gradients, with the hemispheric average stratospheric ∂T/∂y becoming less negative. This implies corresponding changes in the stratospheric zonal winds, with northern hemispheric winds becoming more westward and southern hemispheric winds more eastward, as can be seen in the results of Fletcher et al. (2015), who used the thermal wind approximation to estimate the winds at 100, 1 and 0.5 mbar poleward of 60° from 2005 to 2013.

11.3.2. Meridional Circulation

The zonal-mean circulation in the stratosphere can be inferred from the observed temperature field using the thermodynamic energy equation

$$\frac{\partial T}{\partial t} + \nabla T + w \left( \frac{\partial T}{\partial z} + g/c_p \right) = \frac{q}{c_p},$$

(11.4)

where v and w are the meridional and vertical velocities, $c_p$ is the specific heat, and q is the specific radiative heating/cooling rate. Due to the strong stratification in Saturn’s stratosphere, the vertical advection term is much larger than the meridional advection term, and so the meridional advection is often ignored when estimating the vertical velocity. Observations of stratospheric temperature allow estimates of the time derivative term and the factor in parentheses. Given estimates of the radiative heating/cooling $q/c_p$ from radiative-transfer models and observations, Equation (11.4) can be used to estimate the vertical velocity. The velocities v, w in (11.4) are the so-called "residual mean velocities" (see e.g., Andrews et al. 1987, section 3.5) which include eddy fluxes as well as the advective transport. In the Earth’s stratosphere, the residual mean circulation is a good

$\Delta T_{\text{merid}}$ is the characteristic meridional temperature contrast, which occurs over a meridional scale L. Considering a vertically isothermal temperature profile for simplicity, the vertical advection term scales simply as $\nabla q/c_p$. The continuity equation implies that $\nabla L \sim \nabla D$, where D is the characteristic vertical scale of the circulation. The ratio of the meridional to the vertical advection term is therefore $c_p \Delta T_{\text{merid}}/q L$. We insert numbers for the global-scale seasonal meridional circulation, for which $\Delta T_{\text{merid}} \sim 20$ K (see Figure 11.5). Noting that the circulation is coherent over many scale heights vertically (Figure 11.13), and that the scale height is $H \sim 50$ km at these temperatures, we adopt $D \sim 300$ km. Using $c_p \approx 1.3 \times 10^6$ J kg⁻¹ K⁻¹ and $g \approx 10$ m s⁻² shows that the ratio of horizontal to vertical advection terms is ~0.1.
approximation to the net transport of conserved constituents (Dunkerton 1978).

Flasar et al. (2005) estimated the vertical velocities needed to produce the observed warm temperatures at the south pole by balancing the vertical advection with the time derivative of the temperature, assuming temperature variations on the order of the observed equator-to-pole gradient on a seasonal timescale, finding $w \sim 0.1 \text{ mm s}^{-1}$. Fletcher et al. (2015) used the thermodynamic energy equation to estimate the vertical velocity at latitudes poleward of 60° averaged over the first ten years of the Cassini mission (2004-2014), using temperatures from Cassini CIRS and net radiative heating from the model of Guerlet et al. (2014). They also found velocities of the order of 0.1 mm s$^{-1}$ near 1 mbar, becoming weaker at higher pressures, with rising motion in the southern hemisphere and subsidence in the north (Figure 11.13). At this speed, it would take an air parcel 15 years to rise 50 km, which is about one scale height in Saturn’s stratosphere.

Information on the stratospheric meridional circulation can also be obtained from the observed distribution of disequilibrium chemical species. The most useful species are ethane (C$_2$H$_6$) and acetylene (C$_2$H$_2$), which have photochemical timescales in excess of a Saturn year in the middle and lower stratosphere, with ethane having a somewhat longer timescale. Photochemical models predict that in the absence of meridional transport, ethane and acetylene will have a meridional distribution that follows the yearly mean solar insolation at pressures of 1 mbar and greater, with the abundance a maximum at the equator and decreasing toward the poles; at lower pressures the chemical timescales become shorter and the expected profiles follow the seasonally varying insolation (Moses and Greathouse 2005; Fouchet et al. 2009, see also the chapter by Fletcher et al. in this volume). Observations of ethane and acetylene from ground-based (Greathouse et al. 2005) and Cassini CIRS data (Howett et al. 2007; Guerlet et al. 2009; Hesman et al. 2009; Sinclair et al. 2013; Sylvestre et al. 2015) found the 1–2-mbar acetylene abundance decreased from the equator towards the poles as predicted by photochemical models, while the ethane abundance was approximately constant with latitude. These observations have been interpreted as evidence that meridional mixing occurs on a timescale intermediate between the photochemical timescale of acetylene and ethane (Greathouse et al. 2005; Guerlet et al. 2009; Hesman et al. 2009). However, Moses et al. (2007) have pointed out that the ethane and acetylene photochemistry are strongly coupled, such that the meridional profile of acetylene should track that of ethane even in the presence of transport. Furthermore, CIRS observations by Sinclair et al. (2013) and Fletcher et al. (2015) show an increase in the 2-mbar abundances of ethane and acetylene in the northern hemisphere and decrease in the southern hemisphere, which they interpret as evidence of hemispheric transport on seasonal timescales. Thus, the global meridional distribution of ethane and acetylene is poorly understood.

In addition to the large-scale variations, acetylene and ethane both show a localized enhancement at the south pole at millibar pressures (Hesman et al. 2009; Sinclair et al. 2013; Fletcher et al. 2015). As the acetylene and ethane abundances increase with altitude, this is consistent with downward transport at the south pole indicated by temperature data. Using CIRS limb data, Guerlet et al. (2009) found a local maximum in the acetylene and ethane abundance at 25°N at 0.1 and 0.01 mbar, which they attribute to subsidence in the downward branch of a meridional circulation. This interpretation is consistent with the general circulation model of Friedson and Moses (2012) which produces a seasonally varying low-latitude circulation with rising motion in the summer hemisphere and subsidence in the winter hemisphere. The Equatorial Semi-Annual Oscillation (see Section 11.3.3) may also contribute to the subsidence at this
11.3 OBSERVATIONS OF THE STRATOSPHERE

The earliest, pre-Cassini, models of the meridional circulation (Conrath and Pirraglia 1983; Conrath et al. 1990) assumed a circulation that is mechanically forced in the troposphere, modeled by imposing the observed zonal winds at the lower boundary, and assumed zonal momentum balance between the Coriolis acceleration of the meridional wind and damping of the zonal wind by eddies. Parameterizing the zonal acceleration due to the eddies as $-\overline{u}/\tau_f$, where $\tau_f$ is the eddy-damping timescale, this balance can be written $f\overline{v} = \overline{u}/\tau_f$. Conrath et al. (1990) also included radiative forcing from solar heating and radiative cooling. This model produces upward motion and cold temperatures on the equatorward side of eastward jets, and downward motion and warm temperatures on the poleward side, as well as the observed decay of the zonal jet with altitude. The resulting upper tropospheric temperature perturbations are consistent with the Voyager IRIS observations when the frictional and radiative timescales are comparable, $\tau_f \sim \tau_r$.

A more detailed model of the stratospheric circulation was produced by Friedson and Moses (2012), using a fully 3D general circulation model adapted to Saturn. Their model also uses a mechanical forcing at the lower boundary to produce the observed tropospheric zonal winds, along with random thermal perturbations in the troposphere to simulate generation of waves by convection, and includes realistic radiative forcing. Their model produced global scale temperature gradients generally consistent with observations, though temperatures were $\sim 5$ K too warm in the summer stratosphere. The modeled meridional transport was dominated at low latitudes by a seasonally reversing Hadley circulation, with broad upwelling at low latitudes and strong subsidence near $25^\circ$ in the winter hemisphere, consistent with the circulation inferred from the ethane and acetylene measurements of Guerlet et al. (2009).

At mid-latitudes they found 1-mbar eddy mixing timescales of just over 100 years, slightly longer than the photochemical timescale for acetylene and shorter than for ethane, consistent with the observations. Note, however, that the model was unable to reproduce the Semi-Annual Oscillation (see next section), which compromised the model’s ability to capture the observed variations in temperature and constituents at low latitudes associated with the oscillation.

11.3.3. Equatorial Semi-Annual Oscillation

A common feature of equatorial stratospheres in rapidly rotating atmospheres is the presence of quasi-periodic oscillations, in both time and altitude, of the zonally averaged temperature and wind fields. The best studied of these are the Quasi-Biennial Oscillation (QBO) in the Earth’s lower stratosphere, in which alternating layers of eastward and westward zonal mean winds, with associated warm and cold temperatures, slowly descend with a variable periodicity averaging approximately 28 months (Baldwin et al. 2001). A similar oscillation, with a more regular semi-annual period, is also seen in the upper stratosphere and mesosphere. An approximately four year periodicity in Jupiter’s equatorial temperatures was found in an analysis of ground-based data by Orton et al. (1991). Leovy et al. (1991) proposed that this Quasi-Quadrennial Oscillation (QQO) was analogous to the terrestrial QBO, and calculation of the stratospheric winds from CIRS temperature measurements by Flasar et al.
(2004) showed the presence of a vertical oscillation in the zonal winds.

Using 24 years of ground-based data, Orton et al. (2008) found an oscillation of Saturn’s equatorial brightness temperatures, in both 7.8µm methane emission and 12.2 µm ethane emission, with a period of roughly 15 Earth years, which has been labeled the Saturn Semi-Annual Oscillation (SSAO). Using CIRS limb observations from 2005 to 2006, Fouchet et al. (2008) retrieved temperatures between 20 and 0.003 mbar and found equatorial temperature oscillations similar to those seen in the QBO and QQO (Figure 11.14), confined within ~10° of the equator and oscillating in height with a wavelength of several scale heights. These equatorial perturbations are flanked by temperature perturbations of opposite sign from ~10–20° latitude. Calculation of the zonal winds from the thermal wind equation reveals vertically alternating eastward and westward jets (Figure 11.15); the presence of a stratospheric jet was also inferred from CIRS nadir viewing data (Li et al. 2008). CIRS limb observations from 2010 (Guerlet et al. 2011) showed that the pattern of winds and temperatures had descended in altitude over the intervening five years, as is seen in the QBO. The descent of the temperature field was also observed in equatorial temperature profiles from Cassini radio occultations (Schinder et al. 2011), and cloud tracking measurements from Cassini imaging have shown variations in the velocity of the equatorial jet at the tropopause (Li et al. 2011). Comparisons of Cassini CIRS temperatures from 2009 with a reanalysis of Voyager IRIS thermal data from one Saturn year earlier (Sinclair et al. 2013) found that temperatures at 1–2 mbar were ~7 K warmer in 2010 than 1980 at the equator, and ~4 K cooler at ±10° latitude, indicating that the period of the SSAO is not exactly one-half of a Saturnian year.

The terrestrial QBO is driven by the interaction of a spectrum of vertically propagating waves—of both eastward and westward phase velocities—with the zonal wind. Absorption of the waves by the zonal flow, through radiative or frictional damping, produces a momentum transfer between the wave and zonal flow that accelerates the flow toward the zonal phase velocity of the wave (Lindzen and Holton 1968). The momentum transfer is most effective as the wave approaches a critical level where the zonal phase velocity of the wave matches the zonal wind velocity and the vertical wavelength and vertical velocity go to zero. In the presence of a jet with vertical wind shear, this results in the waves being absorbed at altitudes slightly below where the wind speed matches the wave phase velocity—and below the altitude where the jet speed peaks. This causes the pattern of zonal winds to slowly migrate downward over time (note that, as contours of zonal wind are not material surfaces, this does not imply downward transport of air parcels themselves). As the jet descends, its lower boundary becomes sharper and as the jet descends to the troposphere it dissipates, allowing the waves to now propagate to higher altitudes, where they accelerate a new jet. Equatorial confinement of the oscillation can be explained by two possible mechanisms. First, the waves driving the oscillation may be equatorially confined. Secondly, at latitudes away from the equator, Coriolis forces become important, and the wave induced momentum convergences may be partially balanced by Coriolis forces on the meridional circulation induced by the waves, instead of by the momentum changes to the winds. Full details on the terrestrial QBO

![Thermal wind map](image)

**Fig. 11.15.** Low-latitude, stratospheric zonal winds on Saturn calculated based on the observed meridional temperature gradients using the thermal-wind equation (11.1). The equation is integrated from the 20-mbar level and so the plotted winds are differences relative to those at 20 mbar. The thermal-wind equation breaks down near the equator, and the winds equatorward of the dashed lines are interpolated on isobars from regions outside the dashed lines. Note the existence of an oscillatory structure in the magnitude of the zonal wind with height, which has shifted downward in 2010 relative to its phase in 2005/2006. The difference is plotted in the bottom row. From Guerlet et al. (2011).
can be found in the review by Baldwin et al. (2001). The wave modes responsible for the observed oscillations are poorly constrained, even on the Earth, although Kelvin and eastward-propagating gravity waves are most likely for contributing the eastward jet accelerations, and mixed Rossby-gravity waves and westward-propagating gravity waves are most likely for contributing the westward jet accelerations.

In case of Saturn’s SAO, the amplitude and period of the oscillation likely contain information about the types, spectra, and amplitudes of the wave modes that drive it, and detailed dynamical modeling may thus provide information about the wave properties in Saturn’s stratosphere. Analysis with the Earth would suggest that such waves might arise from convective forcing in the troposphere, so in principle the properties of the SSAO might lead to improved understanding of the extent to which tropospheric convection on Saturn couples to the overlying stratified atmosphere. Moreover, although primarily an equatorial phenomenon, Earth’s QBO exerts a global influence, and this is also likely to be the case on Saturn, though detailed observations and modeling will be necessary to quantify this possibility.

11.3.4. Stratospheric Response To the 2010 Storm

One of the most surprising consequences of the large convective storm, known as a Great White Spot (GWS) that began in December 2010 (see chapter by Sanchez-Lavega et al.), was the observation of large changes in the thermal structure of the stratosphere at northern midlatitudes. Observations taken in January 2011 by Cassini CIRS and ground-based telescopes (Fletcher et al. 2011) showed a cool region in the stratosphere above the anticyclonic vortex associated with the storm, flanked by two warmer regions nicknamed the “beacons,” 16 K warmer than the cold region at roughly 0.5 mbar, much larger than any zonal contrasts previously seen in thermal observations of Saturn’s stratosphere. Subsequent observations from CIRS and ground-based telescopes (Fletcher et al. 2012) showed the beacons moving westward at differing rates, with one of the beacons remaining located above the convective plume associated with the storm, while increasing in temperature. At the end of April 2011, the two beacons merged, resulting in a single warm oval, with a longitudinal extent of 70°, a latitudinal extent of 30° and a peak temperature of 220 K at 2 mbar, 80 K warmer than the background atmosphere and the warmest stratospheric temperatures ever observed on Saturn. This new warm oval moved westward at a velocity intermediate between the two original beacons, and was centered about 1.5 scale heights lower in the atmosphere. The temperature of the beacon dropped by 20 K over the next two months, after which the temperature slowly declined by 0.11 ± 0.01 K day$^{-1}$ and the width of the warm spot shrank by 0.16 ± 0.01 deg day$^{-1}$. Fletcher et al. (2012) used the thermal-wind equation to calculate the wind fields for the initial beacons in March 2011, and for the merged beacon in August 2011. The winds reveal that the hot spots are coherent anticyclonic vortices, with tangential winds of 70–140 m s$^{-1}$ for the pre-merger vortices, and ~200 m s$^{-1}$ for the final merged vortex.

The beacons were associated with perturbations of chemistry as well as temperature, providing additional insight into the dynamics. The abundances of ethylene (C$_2$H$_4$) and acetylene increased significantly in the hot beacon core (e.g., at pressures of order ~1 mbar) relative to pre-storm conditions, and ethane apparently also increased in abundance, though more modestly (Fletcher et al. 2012; Hesman et al. 2012; Moses et al. 2015). One-dimensional photochemical models show that the temperature increase alone, while modifying the chemistry, is insufficient to explain the observed increases (Cavalié et al. 2015; Moses et al. 2015). Therefore, dynamics likely played a role. In particular, all three species increase with altitude at pressures of ~1–10 mbar, so a likely explanation is that subsidence occurred inside the beacon, advecting the greater high-altitude abundances of these species downward into the beacon core region. The compressive temperature increase associated with such descent would also naturally explain the high beacon temperatures. Nevertheless, the descent velocity required to explain the chemical abundances in photochemical models is surprisingly large—Moses et al. (2015) suggest that ~10 cm s$^{-1}$ is required. It is not clear whether this is consistent with the magnitude of descent needed to explain the temperature increase, nor how it would be generated.

Models of moist convection on Saturn indicate that the convective plumes will not penetrate above the upper troposphere, 150 mbar or so (Hueso and Sánchez-Lavega 2004), and modeling of Cassini imaging data of the plume indicated that the cloud tops were at 400 mbar or deeper. It is thus unlikely that the stratospheric perturbations were produced directly by the convective plumes associated with the GWS (García-Melendo et al. 2013). Sayanagi and Showman (2007) showed that tropospheric storms can cause significant wave generation, and that such waves can propagate into the stratosphere where they exert a significant dynamical effect when they break or become absorbed. Fletcher et al. (2012) noted that the storm was located at a latitude where the zonal winds have been suggested to be barotropically unstable (Achterberg and Flasar 1996; Read et al. 2009a), and where quasi-stationary Rossby waves may potentially propagate upward from the troposphere to the stratosphere. They therefore suggested that the strong perturbations in the stratosphere were the result of waves, either gravity or Rossby waves, generated by the convection impinging on the statically stable tropopause, propagating into the stratosphere. This conjecture still needs to be tested with additional numerical modeling.

11.4. Dynamics of the Zonal Jets

11.4.1. Jet Structure

As described in Section 11.2, Saturn rotates rapidly, and the large-scale dynamics is in approximate geostrophic balance. Although we lack detailed observations of the deep
interior, dynamical balance arguments can be used to con-
strain the structure of the jets there. In particular, for a zonal
flow at low Rossby number, it can be shown that, to good
approximation (Showman et al. 2010)
\[ 2\Omega \frac{\partial u}{\partial z_\ast} \approx \frac{g}{\rho} \left( \frac{\partial \rho}{\partial y} \right)_p, \]  
where \( z_\ast \) is the coordinate parallel to the rotation axis, \( g \) is
gravity, and \( \rho \) is density. This is a generalized thermal-wind
equation that is valid in the deep molecular interior even
in the presence of non-hydrostatic motions. If the density
is constant on isobars—that is, if the fluid is barotropic—
then Eq. (11.5) implies that the zonal wind is constant on
surfaces parallel to the rotation axis, i.e.,
\[ \frac{\partial u}{\partial z_\ast} \approx 0. \]  
This result is analogous to the standard Taylor-Proudman
theorem (Pedlosky 1987), except that it does not make
any assumption of constant density. It implies that, for a
barotropic flow, the zonal wind is constant along lines par-
allel to the rotation axis, and is valid even if the mean den-
sity varies by orders of magnitude from the atmosphere to
the deep interior, as it does on Saturn. On the other hand, if
density varies on isobars—that is, if the fluid is baroclinic—
then (11.5) implies that the zonal wind varies along surfaces
parallel to the rotation axis.
These results have been used to argue for several end-
point scenarios regarding the interior wind structure. Con-
vective mixing is normally thought to homogenize the in-
terior entropy, which would suggest a nearly barotropic in-
terior in which variations of density on isobars are small.
Convective plumes of course involve thermal perturbations,
but simple mixing-length estimates suggest that at Saturn’s
heat flux these convective density variations induce only
slight deviations from a barotropic state (Showman et al.
2010). Based on arguments of this type, Busse (1976) sug-
gested that the observed zonal winds on Jupiter and Sat-
urn extend downward into the interior on surfaces paral-
lel to the rotation axis following Equation (11.6); the ob-
served zonal winds would then be the surface manifesta-
tion of concentric, differentially rotating cylinders of fluid
centered about—and parallel to—the rotation axis. On the
other hand, other authors have suggested that perhaps the
jets on Jupiter and Saturn extend only a few scale heights
into the interior (e.g. Ingersoll and Cuzzi 1969; Stone 1973;
Ingersoll 1976). In this case, Eq. (11.5) demands the ex-
istence of latitudinal thermal variations, and would imply
that, just below the cloud deck, the cyclonic bands are
cooler (denser) than the anticyclonic bands. The required
temperature differences, \( \sim 5-10 \text{~K} \), are modest, and could
plausibly be supplied by variations in latent or radiative
heating between latitude bands. Intermediate scenarios are
also possible, with horizontal temperature gradients and
vertical variation of the jets in the cloud layer superposed
on deep, nearly barotropic jets in the interior (Vasavada and
Showman 2005). Liu et al. (2013, 2014) suggested yet an-
other scenario based on the assumption that the interior en-
tropy is only homogenized in the direction of \( z_\ast \), while al-
lowing for entropy variations in the direction toward/away
from the rotation axis. Via Equation (11.5), this scenario
also implies significant thermal wind shear and a gradual
decay of the jets with depth.
It has long been recognized that the above arguments
may break down in the metallic hydrogen interior at pres-
sures \( \geq 10^6 \text{~bars} \). There, the high electrical conductivity
allows sufficiently strong Lorentz forces to alter the forces
balances, causing a breakdown of geostrophy. It has been
suggested that the Lorentz forces will act to brake the zonal
flows, leading to weak winds in the metallic region (e.g.
Busse 1976, 2002). The transition from molecular to met-
allc hydrogen is gradual (Nellis 2000), and several authors
have pointed out that significant magnetic effects on the
flow can occur even in the extended semiconducting re-
gion between the metallic interior and the overlying molec-
ular (electrically insulating) envelope (Kirk and Stevenson
1987; Liu et al. 2008). In particular, Liu et al. (2008) ar-
gued that the observed zonal winds cannot penetrate deeper
than 96% and 86% of the radius on Jupiter and Saturn, re-
spectively, because otherwise the Ohmic dissipation would
exceed the planets’ observed luminosities. If such Lorentz-
force braking occurs at the base of the molecular region
and leads to weak winds there, and if the overlying jets are
nearly barotropic, then Equation (11.6) would suggest that
the winds may be weak throughout much of the molecular
interior—and not just in the metallic and semiconducting
regions (Liu et al. 2008). Nevertheless, the models to date
have been largely kinematic, and more work is needed to
determine self-consistent solutions to the full magneto-
hydrodynamic problem (Glatzmaier 2008). The main region
that can escape such coupling is at low latitudes—there, it is
possible for Taylor columns to extend throughout the planet
without ever encountering electrically conducting layers.
Interestingly, the width of this region is similar to the ob-
served width of the equatorial jets on Jupiter and Saturn.
An attractive scenario is therefore that the equatorial jets
penetrate throughout the planet along surfaces parallel to
the rotation axis (approximately following Equation 11.6 in
the deep atmosphere), while the higher-latitude jets truncate
at some as-yet poorly determined depth. Juno and Cassini
will provide observational constraints on this question in the
next year.

11.4.2. Models Of Jet Formation

Thorough reviews of models for jet formation on Jupiter
and Saturn were provided by Vasavada and Showman
(2005) and Del Genio et al. (2009); here, we summarize
only the highlights, emphasizing developments within the
last ten years. Two classes of model have been introduced
to explore the zonal jets on Jupiter and Saturn. In one ap-
proach, which we dub the “shallow forcing” scenario, it
is assumed that baroclinic instabilities, moist convection,
and other processes within the cloud layer are responsible for driving the jets. This scenario has mostly been explored with thin-shell, one-layer and multi-layer atmospheric models from the terrestrial atmospheric dynamics community. In another approach, which we dub the “deep forcing” scenario, it is assumed that convection throughout the interior leads to differential rotation in the molecular envelope that manifests as the zonal jets at the surface. This scenario has primarily been explored with Boussinesq and anelastic models of convection in spherical shells that are derived from the geodynamo and stellar convection communities. The distinction between the approaches is artificial, but because of the different communities involved, and the difficulty of constructing a truly coupled atmosphere-interior circulation model, the distinction has persisted over the ~40-year history of the field. Nevertheless, new attempts are being made to bridge this gap, an area that will see major progress in the next decade.

It is important to emphasize that the issue of whether the jets exhibit deep or shallow structure is distinct from whether the jet driving occurs primarily in the atmosphere versus the interior (Vasavada and Showman 2005; Showman et al. 2006). Because of the nonlocal nature of the atmospheric response to eddy driving, zonal jets that extend deeply into the interior can result from eddy accelerations that are primarily confined to the atmosphere (Showman et al. 2006; Lian and Showman 2008; for theory, see Haynes et al. 1991). Likewise, under some conditions, convection throughout the interior can produce jets that may exhibit confinement near the outer margin of the planet (Kaspi et al. 2009).

Models for the atmospheric jet formation generally assume that the zonal jets result from the interaction of turbulence with the $\beta$ effect that is associated with latitudinal variations in the Coriolis parameter. Early work emphasized two-dimensional, horizontally non-divergent models and one-layer shallow-water models. More recently, several three-dimensional models have been performed.

One-layer models are intended to represent an appropriate vertical mean of the atmospheric flow. They are motivated by the observation that the atmospheric winds on Jupiter and Saturn are approximately horizontal—with generally small horizontal divergence—and were originally intended to capture the flow in a presumed shallow weather layer below the clouds. More broadly, however, they serve as an ideal process model to explore the dynamics of zonal jets and vortices in the simplest possible context, thereby shedding insights into dynamical mechanisms that may also occur (but be harder to identify) in more realistic systems. In one-layer models, the effect of convection must be parameterized, typically by introducing small-scale vorticity sources and sinks that fluctuate in time. Damping is commonly represented using a frictional drag.

Most work to date has been performed with the two-dimensional, horizontally non-divergent model. This system captures the interaction of turbulence with the $\beta$ effect, but neglects buoyancy, gravity waves, and any effects associated with a finite Rossby deformation radius (an assumption that is not strictly valid on Jupiter and Saturn, where the deformation radius is comparable to or smaller than the meridional jet width). When forced with small-scale turbulence under conditions of fast rotation and relatively weak frictional drag, these models generally produce multiple zonal jets whose meridional widths scale with the Rhines scale $(U/\beta)^{1/2}$, where $U$ is a characteristic wind speed$^6$(e.g. Williams 1978; Nozawa and Yoden 1997; Huang and Robinson 1998; Sukoriansky et al. 2007, and many others). Interestingly, these models lack an unusually strong equatorial jet as occurs on Jupiter and Saturn—instead, the equatorial jet tends to resemble the higher-latitude jets in speed and structure, and it often is not centered precisely around the equator. These models also tend not to produce large, long-lived coherent vortices that coexist with the jets, such as Jupiter’s Great Red Spot or smaller but analogous vortices on Saturn. Both of these discrepancies likely result from the lack of a finite deformation radius in the 2D non-divergent model. Despite these failures, the ease of analyzing this model has led to crucial insights into the workings of inverse energy cascades (Sukoriansky et al. 2007) and the physical mechanisms for jet formation (e.g. Dritschel and McIntyre 2008).

The one-layer shallow-water model represents a more realistic system because it includes the effect of buoyancy (via a horizontally variable vertical layer thickness), gravity waves, and a finite deformation radius, albeit still in a context that does not capture detailed vertical structure. In the context of giant planets, the model captures the behavior of an active atmospheric weather layer that overlies an abyssal layer (representing the deep planetary interior) whose winds are specified, usually to be zero. The related, one-layer quasi-geostrophic (QG) model provides a simplification by restricting the flow to be nearly geostrophic, which filters gravity waves while retaining the effect of a finite deformation radius. Showman (2007) and Scott and Polvani (2007) presented the first forced turbulence calculations of jet formation in the shallow water system, and Li et al. (2006) performed a similar study in the QG system. These authors showed that multiple zonal jets can result from small-scale forcing, and that in many cases, the jet widths are comparable to the Rhines scale $(U/\beta)^{1/2}$. Interestingly, the deformation radius influences jet formation, and when it is sufficiently small, can suppress the formation of jets entirely, leading to a flow dominated by vortices—an effect first described in the QG system (Okuno and Masuda 2003; Smith 2004) before being extended to the shallow-water system (Showman 2007; Scott and Polvani 2007). Thomson and McIntyre (2016) showed in a QG model that the presence of specified, zonally symmetric jets in the abyssal layer can help to increase the straightness of the weather-layer.

$^6$In Rhines’ original theory, this corresponds to an eddy velocity associated with the turbulence, but in other contexts other velocity scales such as the zonal-mean zonal wind can also be appropriate (see, e.g., Scott and Dritschel 2012).
jets, causing the flow to more closely follow latitude circles. Using a simple parameterization of convective forcing, their model also naturally produces a preference for convection in belts rather than zones, as observed on Jupiter. In some cases, especially when jets in the abyssal layer exist, the weather-layer jets in one-layer shallow-water and QG models can violate the Charney-Stern stability theorem, in agreement with Jupiter and Saturn (Showman 2007; Scott and Polvani 2007; Thomson and McIntyre 2016).

Under conditions appropriate to Jupiter and Saturn—namely, small Rossby number and a deformation radius a few percent of the planetary radius—shallow-water models typically produce a broad, fast westward equatorial jet (Cho and Polvani 1996; Iacono et al. 1999; Showman 2007; Scott and Polvani 2007). This equatorial flow intensification could be considered a step forward from the 2D non-divergent model, since the equatorial jets of Jupiter and Saturn are faster and wider than the jets at higher latitudes. However, although Uranus and Neptune exhibit westward equatorial flow, this result represents a major failing for Jupiter and Saturn, where the equatorial flow is eastward. Nevertheless, Scott and Polvani (2008) showed that under certain conditions, shallow-water models can generate eastward equatorial jets resembling those on Jupiter and Saturn. They speculate that the key property allowing emergence of eastward (rather than westward) equatorial flow is the usage of radiative rather than frictional damping. However, this is inconsistent with the findings of Showman (2007), who adopted radiative damping and yet always obtained westward equatorial jets. Most likely, specific types of both forcing and damping are necessary. A similar “thermal shallow water” model presented by Warford and Dellar (2014) exhibits equatorial superrotation at short radiative time constant but subrotation at long radiative time constant, although the value of the radiative time constant at the transition between these regimes is too short to explain the transition from the Jupiter/Saturn regime to the Uranus/Neptune regime. The dynamical mechanisms controlling the equatorial jet properties in all these shallow-water-type models remain poorly understood and deserve further study.

Over the past 15 years, several three-dimensional models have been published showing how zonal jets can develop in the atmosphere. Following on earlier work by Williams (2003), Lian and Showman (2008) showed that baroclinic instabilities in the weather layer (induced by meridional temperature differences associated with the latitudinal gradients in radiative heating) can generate multiple zonal jets. These jets do not remain confined to the cloud layer where they are driven, but rather can extend deep into the interior as long as friction there is weak. The model produces statistical distributions of eddy momentum fluxes that match observations on Jupiter and Saturn, as well as jets that remain stable while violating the Charney-Stern stability criterion. Both Williams (2003) and Lian and Showman (2008) showed that sharp latitudinal temperature gradients near the equator can lead to equatorial superrotation.

Schneider and Liu (2009) presented a shallow 3D model (truncated at 3 bars), which, as in Williams (2003) and Lian and Showman (2008), produces multiple mid-to-high latitude jets in response to baroclinic instabilities associated with latitudinal temperature gradients. Moreover, they introduced a simple convective parameterization which allows the emergence of Jupiter-like equatorial superrotation in response to equatorial convection in the model.

Several 3D models now exist that explain the overall features of the circulation on all four giant planets, including the transition from equatorial superrotation on Jupiter and Saturn to equatorial subrotation on Uranus and Neptune. Lian and Showman (2010) incorporated a hydrological cycle that captures the advection, condensation, rain out, and latent heating associated with water vapor, providing the first test of the long-standing hypothesis that such latent heat transport near the equator can lead to equatorial superrotation.

**Fig. 11.37.**—3D atmosphere simulations demonstrating that latent-heat release associated with condensation of water can lead to zonal jet patterns qualitatively resembling all four giant planets. Top, middle, and bottom rows are three models representing Jupiter, Saturn, and Uranus/Neptune, which are assumed to have water abundances of 3, 5, and 30 times solar, respectively. Color scale represents zonal wind in m s$^{-1}$.

**Fig. 11.16.**—3D atmosphere simulations demonstrating that latent-heat release associated with condensation of water can lead to zonal jet patterns qualitatively resembling all four giant planets. Top, middle, and bottom rows are three models representing Jupiter, Saturn, and Uranus/Neptune, which are assumed to have water abundances of 3, 5, and 30 times solar, respectively. Color scale represents zonal wind in m s$^{-1}$.

**Fig. 11.16.**—3D atmosphere simulations demonstrating that latent-heat release associated with condensation of water can lead to zonal jet patterns qualitatively resembling all four giant planets. Top, middle, and bottom rows are three models representing Jupiter, Saturn, and Uranus/Neptune, which are assumed to have water abundances of 3, 5, and 30 times solar, respectively. Color scale represents zonal wind in m s$^{-1}$.

**Fig. 11.16.**—3D atmosphere simulations demonstrating that latent-heat release associated with condensation of water can lead to zonal jet patterns qualitatively resembling all four giant planets. Top, middle, and bottom rows are three models representing Jupiter, Saturn, and Uranus/Neptune, which are assumed to have water abundances of 3, 5, and 30 times solar, respectively. Color scale represents zonal wind in m s$^{-1}$.
heating is crucial in driving the circulation (e.g. Barcilon and Gierasch 1970; Ingersoll et al. 2000). Their results are shown in Figure 11.16. In addition to explaining the observed transition from superrotation to subrotation, as well as mid-to-high latitude jets qualitatively similar to those observed on the giant planets, their model also produces local “storm”-like features that crudely resemble convective thunderstorm events observed on Jupiter and Saturn. In some cases, their model exhibits large storm events, which bear resemblance to Saturn’s recent Great White Spot of 2010-2011. Liu and Schneider (2010) extended the work of Schneider and Liu (2009) to all four planets, likewise showing a transition from equatorial superrotation on Jupiter and Saturn to equatorial subrotation on Uranus and Neptune, as well as other jet properties similar to those on the giant planets. All of these models represent a major step forward. Nevertheless, the equatorial jet under Saturn conditions is insufficiently fast and broad, and more work is warranted to determine whether equatorial jets better resembling Saturn can be produced in this class of model.

We now turn to discuss models of jet formation via convection in the interior. Busse (1976) first envisioned that the convection in the interior could organize into the form of Taylor columns aligned with the rotation axis, and that the Reynolds stresses associated with this convection would drive jets that would manifest as concentric, differentially rotating cylinders centered on the rotation axis. The observed zonal jets would then represent the outcropping of this differential rotation at the cloud level. Early investigations of this hypothesis involved analytical calculations, laboratory studies, and numerical simulations in the linear and weakly nonlinear regime. These studies generally confirmed the columnar nature of the convection at low amplitude, and showed how the spherical planetary shape would promote the emergence of equatorial superrotation (see Busse 1994 and Busse 2002 for reviews). Only in recent years, however, have computational resources become sufficient to investigate the dynamics in the more strongly nonlinear regime relevant to the giant planets.

The first such models adopted the Boussinesq approximation wherein the background density is assumed constant (i.e., the continuity equation is \( \nabla \cdot \mathbf{v} = 0 \)). This assumption is not valid for the giant planets, but the dynamics are nevertheless rich and provide significant insights. Generally, these models explore convection in a spherical shell with boundaries that are free-slip in horizontal velocity, and that are maintained at constant temperature (with the interior boundary being hotter than the outer boundary, allowing convection to occur). The earliest models explored relatively thick shells, with an inner-to-outer radius ratio less than 0.7. These simulations showed that when the convection is sufficiently vigorous and friction sufficiently weak (i.e. the Rayleigh number sufficiently high and Ekman number sufficiently low), the convection can produce zonal jets whose speeds are significantly faster than the convective velocities (Aurnou and Olson 2001; Christensen 2001, 2002). The jets take the form of a broad equatorial superrotating jet and, in some cases, a small number of weaker jets at high latitudes. Despite the existence of the superrotation, the zonal jet patterns in these models do not resemble that on Jupiter and Saturn. Subsequent studies considered thinner shells, with an inner-to-outer radius ratio of \( \sim 0.9 \), and when appropriately tuned, these models are able to produce superrotation comparable to that of Jupiter and Saturn, along with multiple high-latitude jets (Heimpel et al. 2005; Heimpel and Aurnou 2007; Aurnou et al. 2008, see Figure 11.17). An issue is that, in all of these models, the meridional width of the superrotation is controlled by the depth of the inner boundary, and yet such a boundary is artificial in the context of a giant planet (which lack any such impermeable boundary in their molecular/metallic envelopes).

In Jupiter and Saturn, the density varies by a factor of \( \sim 10^4 \) from the cloud deck to the deep interior, and recently several models have started to include such radial density variations via the anelastic approximation, which allows the background density to vary radially, while (like the Boussinesq system) filtering sound waves, which is a reasonable approximation for the giant planet interiors. The first such models were two-dimensional, considering convection in the equatorial plane and investigating the emergence of differential rotation (Evonuk and Glatzmaier 2004, 2006, 2007; Glatzmaier et al. 2009). These models demon

Fig. 11.17.— The zonal flow on the outer and inner surface of a 3D convection simulation with an aspect ratio of 0.9, showing the eastward (red) and westward (blue) zonal velocities. The Boussinesq equations were solved, in which the background (reference) density is independent of radius. Free-slip boundary conditions are used on both the inner and outer boundaries. Equatorial superrotation, and numerous higher-latitude eastward and westward jet, develop. Adapted from Heimpel et al. (2005).
strated a new mechanism for generating equatorial superrotation that does not exist in the Boussinesq system; as convective plumes rise and sink, the compressibility alters their vorticity in such a way as to promote Reynolds stresses that cause superrotation (Glatzmaier et al. 2009).

More recently, several 3D anelastic convection models have been developed (Kaspi 2008; Kaspi et al. 2009; Jones and Kuzmany 2009; Jones et al. 2011; Gastine and Wicht 2012, and others). Like the Boussinesq models, these anelastic simulations generally exhibit equatorial superrotation and (in some cases) several higher-latitude zonal jets. If the spherical shell is thin, then the shell thickness controls the meridional width of the equatorial jet—just as in Boussinesq simulations. However, if the shell thickness becomes sufficiently deep, then the equatorial jet width tends to become invariant to the location of the bottom boundary, which is an improvement over the situation in Boussinesq models. Nevertheless, in this situation, the superrotation in the simulations tends to be too broad and the higher latitude jets fewer in number compared to Jupiter and Saturn. In the models of Kaspi (2008) and Kaspi et al. (2009), the jets exhibit significant shear, becoming weaker with depth; in contrast, some other models exhibit weaker shear, with a more barotropic structure (Jones and Kuzmany 2009; Gastine and Wicht 2012; Cai and Chan 2012; Gastine et al. 2013; Chan and Mayr 2013). The difference likely results from the different treatment of the fluid’s thermodynamic properties, including the radial variation of density and especially the entropy expansion coefficient, in these different models.

A concern with these models is that the simulated parameter regime is far from that of the giant planets; the simulated heat fluxes and viscosities are both too large by many orders of magnitude. Even within the simulated range, the jet speeds vary strongly with Rayleigh number and Ekman number. Therefore, even though it is possible to choose values of the Rayleigh number and Ekman number that yield realistic jet speeds, it is unknown whether such models would produce realistic jet speeds at the values of Rayleigh and Ekman number relevant to Jupiter and Saturn. Unfortunately, computational resources will be insufficient for directly simulating the planetary regime for the foreseeable future. Therefore, answering this question requires the development of scaling laws that can be extrapolated from the simulated regime to the planetary regime. Christensen (2002) made a first attempt at this, suggesting a scaling law that could bridge the gap based on an empirical assessment of his Boussinesq simulations. Showman et al. (2011) used anelastic simulations to show that, within the simulated range of Rayleigh and Ekman numbers, two distinct regimes exist, leading to distinct dependences of jet speeds on convective heat flux and viscosity. Christensen (2002) and Showman et al. (2011) both suggested an additional regime may exist wherein jet speeds become independent of viscosity when the viscosity is sufficiently small. More recent work, however, challenges the existence of a regime independent of viscosity and thermal diffusivity (Gastine et al. 2013, 2014). Additional work is warranted.

All of the above models neglect any coupling to the magnetic field. Recently, however, several models have included the effect of an electrically conducting interior and its coupling to the dynamics, allowing a joint investigation of zonal-jet formation and dynamo generation of a magnetic field. This is a challenging problem because the electrical conductivity varies by many orders of magnitude with radius, a situation different from that encountered in the geodynamo problem. This situation was first investigated in Boussinesq models (Heimpel and Gómez Pérez 2011) and more recently in anelastic models (Duarte et al. 2013; Yadav et al. 2013; Jones 2014) that allow both density and electrical conductivity to vary strongly with radius. These models show that the equatorial jet penetrates to a depth where the Lorentz force becomes comparable to the Reynolds stress associated with the convection (which effectively means the jet is confined to the electrically insulating outer envelope). Jets poleward of the equatorial superrotation tend to be suppressed, because their bottoms penetrate into the electrically conducting region where Lorentz forces can act to brake the flows—an effect that extends throughout the molecular envelope due to the nearly barotropic nature of these jets (cf Equation 11.6). This suggests the possibility that the equatorial jet results from interior convection, but the higher latitude jets result from atmospheric processes including baroclinic instability and moist convection (Vasavada and Showman 2005).

### 11.4.3. Detection Of Deep Dynamics By Gravity Measurements

Information to date on Saturn’s deep winds has been indirect. This is likely to change in 2017, as towards the culmination of its 13-year-long survey of the Saturnian system, the Cassini spacecraft will shift into a highly inclined orbit with a periapse between the planet and its rings—an orbit ideally suited to measuring the small-scale structure of Saturn’s gravitational field. During this phase, known as the Cassini Grand Finale, the spacecraft will complete 22 orbits, six of which will be dedicated to gravity science. This will be the final maneuver of Cassini before it descends into the planet, terminating the mission. These gravity measurements will allow the determination of Saturn’s gravity field to much higher accuracy than exists today (Jacobson et al. 2006). The gravitational field is commonly represented using a spherical harmonic expansion (Hubbard 1984), and so far only the lowest zonal harmonics $J_2$, $J_4$, and $J_6$ have been measured. These reflect the long-wavelength gravitational perturbations associated primarily with the planet’s rotational bulge, and provide essentially no information on interior flows. Cassini’s proximal orbits will allow the measurement of the gravity field at least up to $J_{10}$, including the possibility of measuring the odd gravity harmonics for the first time (Kaspi 2013).

If the strong cloud-level winds extend sufficiently deep
Fig. 11.18.— Anelastic 3D simulations of convection in a giant planet interior including coupling to the magnetic field, from Duarte et al. (2013). Each image shows the zonal-mean structure of a distinct simulation; the color scale shows the zonal-mean zonal wind (expressed as a Rossby number) and the contours show poloidal magnetic field. $N_p$ denotes the number of density scale heights spanned from the outer to the inner boundary; the left column ($N_p = 0$) denotes Boussinesq simulations, while the right column ($N_p = 5$) presents simulations with five density scale heights (inner density 148 times the outer density). The top row represent hydrodynamic models (no MHD effects). The middle and bottom rows represent MHD simulations, with electrically insulating outer regions and electrically conducting inner regions. The transition occurs at a fractional radius of 0.95 and 0.8 for simulations in the middle and bottom rows, respectively. Robust equatorial superrotation can be seen in all models, which is confined to the electrically insulating region when MHD is used. The inclusion of compressibility and magnetic effects suppress the formation of zonal jets at mid-to-high latitudes.
2013). Given some assumed structure of the zonal winds, the thermal-wind relation (Equation 11.5) can be used to calculate the dynamically induced density gradients in the interior. These density perturbations in turn cause perturbations to the gravity field. The dynamical contributions to the gravity harmonics are defined as (Kaspi et al. 2010)

\[ \Delta J_n^{\text{dyn}} = - (M a^n)^{-1} \int P_n \rho^\prime r^n d^3r, \]  

(11.7)

where \( \rho^\prime \) is the deviation of the density caused by dynamics (i.e., the deviation from some static density distribution, \( \rho_{\text{static}} \), that would exist in the absence of dynamics), \( P_n \) is the \( n \)-th Legendre polynomial, \( M \) is the planetary mass and \( a \) is the mean planetary radius. Kaspi et al. (2016) showed that in the barotropic limit the thermal-wind and potential-theory approaches give very similar results. However, the thermal-wind models have typically been limited to spherical symmetry, resulting in inability to calculate the static (solid-body) gravity spectrum and neglecting the effect of the oblateness of the planet on the dynamic gravity moments. Zhang et al. (2015) suggested that such effects, namely the self gravity due to the dynamical variation itself \( g(r, \theta) \), can be important, and Cao and Stevenson (2015) suggested that the effect of the oblateness on the background mean state can be important for the gravity moments. However, solving self-consistently for the full oblate system, Galanti and Kaspi (2017) have shown that these oblateness effects give a very small contribution to the gravity moments, meaning that the spherical thermal wind captures well the leading order balance, and therefore can be used for calculating the gravity moments.

Figure 11.19 shows the gravity field resulting from a model using the thermal-wind approach, where the gravity harmonics are shown as a function of the depth that the cloud-level winds extend beneath the cloud-level. It shows that beyond \( n = 10 \) the gravity signal induced by the wind structure becomes larger than the solid-body (static) gravity harmonics.

Because Jupiter and Saturn are gaseous, aside from the cloud-level winds there is no apparent asymmetry between the northern and southern hemispheres. Therefore, the gravitational moments resulting from the shape and vertical structure of the planets have identically zero odd moments (Hubbard 1984, 1999). Unlike the even gravity moments that have a contribution both from the static density distribution and the dynamics, the odd moments are caused purely by dynamics. Thus, any odd signal detected \( (J_3, J_5, J_7, \ldots) \) will be a sign of a dynamical contribution to the gravity. Indeed, the cloud-level winds are not precisely symmetrical around the equator, and if these hemispheric differences extend into the interior, they could produce measurable odd harmonics (Figure 11.19) (Kaspi 2013). Using also the thermal-wind approach, Liu et al. (2013) calculated the penetration depth of the winds on Jupiter with the additional assumption that the entropy gradient in the direction of the spin axis must be zero. This requirement sets the penetration depth of the winds, and they have also found that such a wind structure (which extends deep throughout the molecular envelope) should be detectable by Juno and Cassini (Liu et al. 2013, 2014).

The main challenge of interpreting the gravity data will be to invert the gravity measurements into meaningful wind fields. Until recently all studies have been only in the direction of forward modeling; thus, given a hypothetical wind structure (based on the observed surface winds and some assumption regarding the penetration depth) the gravity moments are calculated via the effect of the winds on the density structure (e.g., Hubbard 1999; Kaspi et al. 2010; Liu et al. 2013; Kong et al. 2012). However, in order to analyze the gravity field that will be detected by Juno and Cassini the inverse problem needs to be solved—calculating the zonal wind profile given the gravity field. This causes a difficulty since a gravity field is not necessarily invertible and a given gravity field might not have a unique corresponding wind structure. Galanti and Kaspi (2016, 2017) propose to address this using an adjoint based inverse method that will allow the investigation of the giant planet dynamics using the observed measured gravity field. This method has been used extensively in the study of oceanic and atmospheric fluid dynamics (e.g., Tziperman and Thacker 1989; Moore et al. 2011). This method can be applied to any forward relation between the gravity and the wind structure (e.g., thermal-wind, potential-theory, etc.), and to 3D wind struc-

![Fig. 11.19.— The gravity harmonics of Saturn corresponding to the static, rotationally flattened contribution (squares) and the contribution from dynamics (circles) calculated using the thermal-wind model of Kaspi (2013). The dynamical contributions to the gravity harmonics (\( \Delta J_n \)) are shown for four different assumptions about the characteristic length scale, \( H \), over which the jets decay with depth into the interior: \( H = 300 \) km (purple), \( H = 1000 \) km (orange), \( H = 3000 \) km (blue) and \( H = 10000 \) km (maroon). Filled (open) symbols indicate positive (negative) zonal harmonics. Black plus signs show the observed values of \( J_n \), and the dashed line is the expected detectability limit of the Cassini proximal orbits. The static values (Hubbard 1999) have only even components (odd harmonics are identically zero). From Kaspi (2013).]
ture which give the full 3D gravity fields including contributions from longitudinal variations in the wind structure and meridional winds (Parisi et al. 2016).

The coincidental timing of the Cassini proximal orbits with the Juno gravity experiment will provide an opportunity to simultaneously probe the deep dynamics of both Saturn and Jupiter. Comparing the atmospheric dynamics on the two planets shows that the equatorial jet is much stronger and wider on Saturn, extending to latitude 30°, with fewer high-latitude jets—but the jets are much stronger on Saturn reaching about 400 m s$^{-1}$ depending on the reference frame chosen (see section 11.2). The fact that the jets are stronger on Saturn means that the induced gravity anomalies will be generally larger as well. However, as Saturn is more oblate than Jupiter (Jupiter has an oblateness of 6.5%), the static gravity harmonics due to the non-dynamic density distribution will likewise be larger than on Jupiter. These two factors imply that the ratio between the static and the dynamic gravity harmonics are roughly similar on both planets (Kaspi 2013).

Despite the fact that no spacecraft will be visiting Uranus and Neptune in the near future, the strong and broad wind structure and smaller mass on these planets allows the use of these same techniques to infer the depth of the winds using gravity moments that have already been measured. In particular, the speed and meridional breadth of the jets on Uranus and Neptune allow the possibility that—unlike the case of Jupiter and Saturn—even the low harmonics such as $J_4$ can be significantly affected by dynamics. Using the gravity data then available, Hubbard et al. (1991) first showed that the fast winds must be confined to the outermost few % of these planets’ masses. Exploiting recent updates on the gravity harmonic $J_4$ and its uncertainty (Jacobson 2007, 2009), Kaspi et al. (2013) were able to refine this constraint considerably, showing that the observed flows must be constrained to the outermost 1000 km of the planetary radius—corresponding to no more than 0.15% and 0.2% of the mass of Uranus and Neptune, respectively. Thus, the fast jets on Uranus and Neptune are relatively shallow.

In summary, within the next few years, using the fact that if deep flows exist on giant planets they will induce a measurable gravity signal, we will finally address one of the longest-standing debates in planetary science regarding the depth of the observed zonal flows on all four giant planets (Vasavada and Showman 2005). Understanding if the flows are deep or shallow will allow tests between the theories of jet formation and energetics as discussed in Section 11.4.1–11.4.2. The upcoming few years therefore provide an exciting opportunity in the study of the atmospheric dynamics of Saturn and the other giant planets.

11.5 CONCLUSIONS

Saturn’s atmosphere exhibits a wealth of interacting dynamical processes that present an interesting contrast with Jupiter. The dynamics are dominated by fast zonal jets whose meridional width, latitudinal positions, and speeds have remained relatively constant over many decades. Saturn’s clouds are organized into zonal bands that are modulated by (and perhaps affect) the zonal jets, but the relationship of the cloud bands to the jets themselves differs from that on Jupiter and remains poorly understood. Observations indicate that eddies transport momentum up-gradient into the cores of the zonal jets at cloud level, thereby maintaining the jets. This points toward the importance of cloud level processes in jet dynamics, but the exact processes that power these jet-pumping eddies—baroclinic instability, moist convection, or interaction of dry convection with the stratified atmosphere—remain poorly understood. LIGHT-}

ning and storm activity—including small storms as well as the Great White Spot of 2010-2011—provide constraints on the thermal structure and vertical motion below the clouds.

There also exist a variety of vortices, waves, and other local features at cloud level, with lifetimes of months to decades, whose relationship to the jet dynamics remains enigmatic. Notable features include the string of pearls (Sayanagi et al. 2014), the Ribbon, the polar hexagon (Baines et al. 2009), and a significant population of small vortices that tend to cluster predominantly at the latitudes of the westward jets (e.g., Vasavada et al. 2006; Choi et al. 2009). Interestingly, at and above the cloud level, the jets exhibit reversals in the meridional gradient of potential vorticity, and the relative stability of the jets despite these reversals has important implications for the flow below the clouds.

A variety of observations place constraints on the circulation in the upper troposphere and stratosphere, which has important analogies to the stratospheric circulation on the other giant planets as well as Earth. Saturn’s obliquity of 27° causes seasonal temperature changes in the stratosphere, but there also exist a wealth of temperature variations at a variety of length and timescales that are likely dynamical in origin. The temperature patterns above the clouds suggest, through thermal-wind balance, that most of the zonal jets decay with altitude in the upper troposphere. These temperature patterns, along with the ammonia abundance near the clouds, are best explained by a meridional circulation comprising ascent in the anticyclonic regions, producing the cold temperatures and high ammonia abundances, and descent in the cyclonic regions, producing the warm temperatures and low ammonia abundances. Interestingly, such a circulation is thermally indirect and is likely driven by absorption of waves propagating from below. This is directly analogous to the wave-driven circulations that exist in Earth’s stratosphere and that have been inferred in the stratospheres of Jupiter, Uranus, and Neptune. The source of the waves and the specific details of this circulation in the Saturn context remain poorly understood. Additionally, stratospheric thermal perturbations have been observed at low latitudes that suggest an oscillation of equatorial winds and temperatures with a period close to 15 years, which seems to be analogous to the terrestrial Quasi-Biennial Oscillation, which is driven by a population of upward propagating waves generated by convect-
tion and other processes in the troposphere.

Dynamical models do not yet come close to explaining this wealth of phenomena; nevertheless, significant progress in understanding the dynamics of zonal jets and local features has been made over the past ~15 years. Over the past decade, highly idealized models have greatly improved our understanding of the processes that cause the flow to self-organize into zonal jets. More realistic three-dimensional models of the weather-layer flow are now capable of explaining the equatorial superrotation and multiple high-latitude jets on Jupiter and Saturn, although the superrotation in these models tends to be narrower and slower than that on Saturn. Models of convection throughout the planetary interior now include the multi-order-of-magnitude variation of density with radius and are likewise starting to incorporate the coupling to magnetic fields, including the transition from electrically insulating in the molecular envelope to electrically conducting in the deeper interior. These models are now elucidating how the compressibility and coupling to magnetic fields influences the zonal jets and other aspects of the dynamics. These models show that convection in the interior can naturally produce a Saturn-like equatorial jet, but the coupling to magnetic effects tends to suppress any higher latitude jets. An attractive hybrid scenario, which combines the strengths of the shallow and deep models, might suggest that convection in the molecular envelope outside some tangent cylinder causes the broad equatorial jet, but that the mid-to-high-latitude jets result from atmospheric processes such as baroclinic instabilities or moist convection in the cloud layer (Vasavada and Showman 2005).

Future progress in modeling and theory will occur on several fronts. Investigations of atmospheric, weather-layer processes and of deep, convective dynamics are usually conducted separately, using distinct classes of dynamical model that solve distinct equation sets and use distinct numerical codes. The distinction between “shallow” and “deep” processes is artificial, however, and in reality atmospheric and interior processes may interact in a variety of ways not included in current models. A new generation of GCMs that spans both atmosphere and interior processes is needed to explore this coupling, and this will be a major area of progress over the next decade. Global atmospheric models of jet formation do not yet include clouds, and yet clouds represent a significant fraction of the observational database for both Jupiter and Saturn, as well as influencing the dynamics through their mass loading, interaction with radiation, and latent heating and cooling due to their formation and sublimation. Thus, significant progress is possible by including clouds in models, both in widening the range of observations against which the models can be tested, and by illuminating the extent to which clouds influence the atmospheric circulation. Inclusion of realistic radiative transfer and chemistry into stratospheric GCMs will be a growth area, and will help shed light on the implications of tracers such as ethane and acetylene for the dynamics. The nature of the eddies or waves that apparently cause a deceleration of the upper tropospheric and lower stratospheric zonal jets, and that cause the SSAO, are poorly understood. These and many related topics concerning the zonal jets are amenable to continued idealized theory and modeling.

Future progress will also depend crucially on improved observations. Due to the Cassini mission, Saturn is perhaps now the best-observed giant planet—outpacing Jupiter—but many of the Cassini observations have yet to be analyzed, and significant progress is possible with the existing data. For all four giant planets, improvements over the past decade in ground-based astronomical facilities have led to a renaissance in observations of cloud features and upper tropospheric temperature structure from the ground—by professional and amateur astronomers alike. These ground-based observations will continue to play a crucial role in filling in the temporal gaps between higher-quality, but episodic, coverage from spacecraft missions. Moreover, over the next several years, the Juno mission to Jupiter and the Cassini Grand Finale at Saturn will provide gravity measurements that promise to constrain the depth to which the zonal jets extend below the visible cloud decks. Close-in microwave and IR sounding of Jupiter by Juno, as well as close-in sensing of Saturn with Cassini’s suite of instruments during the proximal orbits, may produce qualitatively new constraints on the cloud-level dynamics of both planets. The Juno microwave radiometer will measure the global water abundance and the latitudinal variation of water and ammonia below cloud base. Future missions to the giant planets include the NASA Europa Multiple Flyby Mission and the European Jupiter Icy Moon Explorer (JUICE) mission, which will obtain observations of Jupiter in addition to their prime targets of icy satellites starting around 2030. Prospects for giant-planet missions beyond the 2030s are currently nebulous, but there is strong scientific merit—and interest—in missions to Uranus and/or Neptune, as well as an entry probe mission to Saturn. This activity will not only revolutionize our understanding of the grandest planets in our solar system, but will also provide a foundation for understanding the many brown dwarfs and extrasolar giant planets that are being discovered and characterized light years away.

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Icarus, 555.


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