Effects of a large convective storm on Saturn’s equatorial jet

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Abstract
Hubble Space Telescope observations revealed that Saturn’s equatorial jet at the cloud level blows at ∼275 m s$^{-1}$ today, approximately half the ∼470 m s$^{-1}$ wind during the Voyager flybys in 1980–1981. Radiative transfer calculations estimate the clouds to be significantly higher today than in 1980. The higher clouds make it difficult to observationally isolate any true slowdown from the vertical wind shear because Voyager and Cassini observations show that the winds become slower with altitude. Here, we test the hypothesis that the large equatorial storm in 1990 called the Great White Spot (GWS) decelerated the equatorial jet. We first use order of magnitude estimates to show: (1) if the GWS triggers vertical momentum redistribution, a minor speed change in the troposphere can lead to a substantial stratospheric wind speed change; (2) storm-triggered turbulent mixing slows a prograde equatorial jet; and (3) a prograde equatorial jet inhibits turbulent mixing in latitude. To test whether a GWS-like large storm decelerates the equatorial jet, we perform numerical experiments using the Explicit Planetary Isentropic Coordinate (EPIC) atmosphere model. Our simulation results are consistent with our order of magnitude predictions. We show that the storm excites waves, and the waves transport westward momentum from the troposphere to the stratosphere and decelerate the equatorial jet by as much as ∼40 m s$^{-1}$ at the 10-mbar level. However, our results show that the storm’s effect is too weak at the cloud levels to halve the jet’s speed from ∼470 m s$^{-1}$. Our results suggest that a combination of higher clouds and a true slowdown is necessary to explain the apparent equatorial jet slowdown. We also analyze the effect of waves on the apparent cloud motions, and show that waves can influence cloud-tracking wind speed measurements.

Keywords: Saturn, atmosphere; Atmospheres, dynamics; Meteorology

1. Introduction

Hubble Space Telescope (HST) observations of Saturn between 1994 and 2004 (Sánchez-Lavega et al., 2003, 2004) revealed that the equatorial jet at the cloud level blows at ∼275 m s$^{-1}$, about half the speed measured by Voyager in 1980–1981. The finding came as a surprise for several reasons. First, the slowdown, if real, implies a huge decrease in the jet’s angular momentum. Second, the primary energy source of atmospheric circulation, the 2-W m$^{-2}$ internal heat release (Conrath and Pirraglia, 1983), is much smaller than the solar radiation driving Earth’s circulation. Third, since the Voyager flybys in 1979, Jupiter’s zonal jets have undergone no speed change comparable in magnitude to the apparent slowdown of Saturn’s equatorial jet (smaller changes to the jovian jets are summarized in Vasavada and Showman, 2005). Jupiter and Saturn have very similar zonal jet structures; they both have a broad prograde equatorial jet and other eastward and westward zonal jets associated with the colored banding on the visible surface. Also, on both planets, the jets at higher latitudes tend to be slower than the ones closer to the equator, with a few exceptions. If the observed slower wind represents a significant slowdown of Saturn’s equatorial jet, this may hint at an atmospheric forcing unique to the planet.

There are two endpoint scenarios to explain the observed slower wind. First, the equatorial jet may have undergone a true temporal change in speed, as Sánchez-Lavega et al. (2003) interpreted. Second, the clouds may reside at higher altitudes today than during the Voyager flybys. Infrared measurements show that Saturn’s zonal jet speeds decrease with altitude (Conrath and Pirraglia, 1983; Flasar et al., 2005). Because the winds are measured by tracking discrete cloud features, higher...
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clouds move slower, and this scenario does not necessarily require a true slowdown to explain the observed slower winds. Intermediate scenarios, where the apparent slowdown results partly from a true slowdown and partly from an increase in cloud height, are also possible.

Sánchez-Lavega et al. (2000) suggested that a large convective storm called a Great White Spot (GWS), which last erupted in 1990 and disturbed the entire equatorial zone of Saturn, acted in slowing the jet. However, observationally drawing a connection between the GWS and the apparent wind slowdown is difficult due to the sporadic pre-1990 wind measurements. Sánchez-Lavega et al. (2000) analyzed the Voyager images to determine the zonal wind profile, which placed the maximum equatorial speed at $\sim 470 \text{ m s}^{-1}$. The first wind measurement after the Voyager flybys with sufficient number of wind tracers to retrieve Saturn’s equatorial jet profile was done by Barnet et al. (1992) using the HST on November 17–18th, 1990, approximately 1.5 months after the onset of the 1990 storm. Barnet et al. obtained two profiles with equatorial speeds of $\sim 400 \text{ m s}^{-1}$ at the 547 nm continuum and $\sim 300–60 \text{ m s}^{-1}$ at the 890-nm methane absorption band. Measurements in a methane absorption band and a continuum band sense the top boundaries of the tropospheric haze and the deep cloud deck, respectively (Tomasko et al., 1984). Acarreta and Sánchez-Lavega (1999) estimated that the tropospheric haze top and the cloud deck were at $\sim 60$ and 300–400 mbar, respectively, in November 1990. However, the time required for the zonal wind to respond to such a storm is not known, and whether the Barnet et al. wind speed measurements were already affected by the storm remains unclear. Records of the slower equatorial winds start in 1994 and consistently show features in the HST images in 1994 and 2003–2004 (Sánchez-Lavega et al., 1996, 1999, 2003, 2004), $100–200 \text{ m s}^{-1}$ slower than the Voyager wind at their corresponding latitudes.

Using radiative transfer calculations, Pérez-Hoyos and Sánchez-Lavega (2006) estimated that the cloud tracers in the Voyager images were located at $360 \pm 140 \text{ mbar}$ at $5^\circ \text{ N}$. The features tracked in the HST images in 1994 and 2003–2004 are both estimated to be at $\sim 50 \text{ mbar}$ (Sánchez-Lavega et al., 1996, 2004). Thus, we are comparing two different wind speeds at different times and altitudes in the comparison between the Voyager wind in 1980–1981 and the HST measurements in 1994–2004. The tracked features in Barnet et al. (1992)’s 890 nm observations and the HST data (multi-wavelengths including 890 nm) in Sánchez-Lavega et al. (2004) are both estimated to be at $\sim 50 \text{ mbar}$, and comparing their speeds suggests a $50–100 \text{ m s}^{-1}$ slowdown at this altitude.

Recent Cassini observations of Saturn do not resolve this situation. Flasar et al. (2005) determined the vertical wind shear from the CIRS measurements using thermal wind arguments. They then extrapolated the Voyager wind, assumed to be at 500 mbar, to the stratosphere to find that the wind speed reaches $\sim 275 \text{ m s}^{-1}$ only above 10-mbar level, which seems too high for discrete cloud features to exist. At $\sim 50 \text{ mbar}$, their data imply $\sim 300 \text{ m s}^{-1}$, somewhat faster than the recent HST measurements; the speed discrepancy would be larger if Flasar et al. had assumed that Voyager wind resided at 360 mbar as found by Pérez-Hoyos and Sánchez-Lavega (2006). This hints that a true speed change somewhere in the atmosphere is needed to be consistent with both the slow wind and the wind shear. The ISS cloud tracking by Porco et al. (2005) shows that speeds measured in the 727-nm methane band are virtually the same as the HST speeds measured by Sánchez-Lavega et al. (2004), but the ISS measurements in the 750 nm continuum show that the equatorial jet peaks at $\sim 400 \text{ m s}^{-1}$. Baines et al. (2005a) presented VIMS images that show shadows of clouds in the thermal background moving at close to the Voyager speed at $8^\circ \text{ S}$. These cloud tracers are presumably located deeper than any previous features used to estimate winds, but their exact altitudes remain undetermined [Baines et al. (2005b) estimate the altitude to be $\sim 2 \text{ bar}$].

Little modeling of a GWS-like storm has been carried out to date. Hueso and Sánchez-Lavega (2004) performed three-dimensional anelastic simulations of saturnian cumulus convection; however, the scale of the storms they simulated was much smaller than that of a GWS, and the domain of simulation in their study was much smaller than the planet. How a storm of the scale of a GWS would affect the global environment has not been investigated.

Here, we perform full-3D nonlinear numerical simulations to test the idea that the GWS of 1990 caused a true slowdown of Saturn’s equatorial jet.

The rest of this paper is structured as follows. In Section 2, we present order of magnitude estimates of how large-scale effects of the storm would influence the equatorial wind. The mechanisms we consider are momentum transport by waves and turbulent mixing, both of which may be triggered by a GWS-like episodic event. Our numerical model setup is presented in Section 3. The numerical experiments and their results are in Section 4. Section 5 analyzes the effect of waves on the apparent cloud motions. We discuss the implications of our results in the final section.

2. Possible slowdown mechanisms

2.1. Slow deep winds

If the equatorial jet speed decreases rapidly with depth below the $\sim 1$-bar level, then the GWS could cause mixing between high-speed air aloft and low-speed air at depth, inducing a net deceleration at the cloud level. We expect that moist convection initiates near the $\sim 10$-bar water condensation level, so this scenario requires the jet to approach zero by depths of $\sim 10$ bar. However, if the equatorial jet reduces to zero at $\sim 10$ bar from the $\sim 400 \text{ m s}^{-1}$ wind revealed by Baines et al. (2005b) at the 2-bar level, the thermal wind equation (e.g., Holton, 1992) requires that there must be at least 20 K latitudinal temperature variation over the equatorial jet width. Observations show only $\sim 5$ K latitudinal temperature variation in the troposphere (Conrath and Pirraglia, 1983; Flasar et al., 2005; Orton and Yanamandra-Fisher, 2005). Ingersoll et al. (1984) summarize various estimates of Saturn’s zonal flow depth and argue that the level of no motion (if any) is at least several hundred bars deep. Furthermore, even if the deep winds are weak,
convective mass transport must dilute the momentum enough to halve the cloud level speed, which is highly implausible.

2.2. Vertical momentum redistribution

Another possibility is that the GWS disturbance vertically redistributed the momentum via waves. Such waves are known to vertically transport momenta to alter the equatorial zonal wind in the stratosphere on Earth (Baldwin et al., 2001) and on Jupiter (Friedson, 1999; Li and Read, 2000).

Consider a case in which the disturbance causes a speedup of \( \Delta u_{\text{lower}} \) at a lower atmospheric level across a pressure thickness \( \Delta p_{\text{lower}} \), and to conserve the angular momentum, a slowdown of \( \Delta u_{\text{upper}} \) occurs in an upper level across a pressure thickness \( \Delta p_{\text{upper}} \). We can write

\[
\frac{\Delta u_{\text{upper}}}{\Delta u_{\text{lower}}} \approx \frac{\Delta p_{\text{lower}}}{\Delta p_{\text{upper}}}. \tag{1}
\]

Thus, a \( \sim 10 \text{ m s}^{-1} \) speedup in \( \Delta p_{\text{lower}} \approx 1 \text{ bar} \) would lead to a \( \sim 100 \text{ m s}^{-1} \) slowdown above \( p \approx 100 \text{ mbar} \). It illustrates that a slight change in the deep wind can have a significant effect on the wind aloft although this argument does not constrain the direction of the speed change.

2.3. Potential vorticity homogenization

Turbulent homogenization triggered by a storm may also play an important role. Homogenization of a conserved dynamical quantity such as potential vorticity (PV) puts a constraint on the resulting wind profile. Below, we show that homogenization of PV in the equatorial region must be accompanied by a westward acceleration of equatorial wind.

Here, we estimate the effect of PV homogenization on a prograde equatorial jet using a simple shallow-water calculation. We start by taking an initial zonal wind profile

\[
u_i(y) = u_0 - ky^2, \tag{2}
\]

where \( u_0 \) is the peak wind speed at the equator, \( k \) is a positive constant, and \( y \) is northward distance on the sphere. We assume that both the initial and the PV-homogenized wind profiles have no meridional component and are in geostrophic balance. Geostrophy is usually a poor approximation near the equator; however, it is adequate here in showing the general tendency of the wind speed change because the assumed final state is zonally symmetric. For such a zonally symmetric state, the curvature terms in the shallow-water equations are two orders of magnitude smaller than the Coriolis term over the range of \( y \) and \( u \) we consider, so it is sufficient to proceed with geostrophic approximation. We solve the meridional component of the shallow-water momentum equations in steady state [see, for example, Pedlosky (1987, Eq. (3.3.15b))]

\[-g \frac{dh_i(y)}{dy} - fu_i(y) = 0 \tag{3}
\]

to obtain the initial shallow-water thickness as a function of \( y \),

\[h_i(y) = \frac{\Omega}{ag} \left( \frac{k}{2} y^4 - u_0 y^2 \right) + h_0, \tag{4}\]

where \( f \equiv 2\Omega \sin(y/a) \) is the Coriolis parameter, \( \Omega \) is the planetary rotation angular velocity, and \( h_0 \) is the integration constant.

Taking the shallow-water definition of potential vorticity \( q \equiv (\zeta + f)h^{-1} \) where \( \zeta = -\partial u/\partial y \) is the vertical component of the relative vorticity with zero meridional wind, we find that the spatial average of PV between latitudes \( \pm y_0 \) is zero.

Consequently, when PV is completely homogenized, we can solve for the resulting zonal wind profile \( u_h(y) \). The solution, with the equatorial \( \beta \)-plane approximation \( f \approx 2\Omega(y/a) \), is

\[u_h(y) = \frac{\Omega}{a} y^2 + u_1, \tag{6}\]

\[h_b(y) = -\frac{\Omega}{ga} \left( \frac{\Omega}{2a} y^4 + u_1 y^2 \right) + h_1, \tag{5}\]

where \( u_1 \) and \( h_1 \) are integration constants. This solution illustrates that complete homogenization of PV forces the wind speed at the equator to be a local minimum. It is interesting to note that similar results have been derived by Hide and James (1983) for flows in thick spherical shells relevant for molecular layer-penetrating deep flows in Jupiter and Saturn.

Similarly, consider a scenario in which a partial homogenization of PV resulted in a wind profile \( u_p(y) = u_2 \) constant. This is motivated by the “truncated” appearance of the equatorial jet profile in Sánchez-Lavega et al. (2003). Geostrophically balancing \( u_p(y) \) requires, again with the equatorial \( \beta \)-plane approximation,

\[h_p(y) = -\frac{\Omega}{ga} u_2 y^2 + h_2, \tag{6}\]

where \( h_2 \) is an integration constant.

We constrain the constants \( u_1, u_2, h_1, \) and \( h_2 \) using conservation of mass and angular momentum about the planetary rotation axis. The results are plotted in Fig. 1 a, where values \( a = 60268 \text{ km}, \ \Omega = 1.638 \times 10^{-4} \text{ s}^{-1}, u_0 = 450 \text{ m s}^{-1}, k = 0.452 \times 10^{-12} \text{ m}^{-1} \text{ s}^{-1}, g = 8.96 \text{ m s}^{-2}, \) and \( h_0 = 10^5 \text{ m} \) are used for the calculation. The value of \( k \) is chosen to make \( u = 0 \) at latitudes \( \pm 30^\circ \). \( h_0 \) represents Saturn’s equatorial atmosphere with the equatorial Rossby deformation radius [for its definition, see, e.g., Gill (1982, Eq. (11.5.4))].

\[L_D = (a(2\Omega)^{-1} \sqrt{g/h_0})^{1/2} \sim 10^4 \text{ km}. \]

PV is homogenized between \( 10^7 \) north and south latitudes in these calculations. The associated PV profiles are shown in Fig. 1 b, illustrating that a wind profile constant in latitude such as \( u_p(y) = u_2 \) can result from a slight homogenization of PV, though the wind speed change in this case is minor. These calculations illustrate that PV homogenization tends to decelerate the wind at the equator and accelerate the surrounding regions.

2.4. Prograde equatorial jet inhibits wave breaking

We now show that a prograde equatorial jet inhibits Rossby waves from breaking. Here, we consider a jet that satisfies

\[\frac{d}{dy} (\zeta + f) > 0. \tag{7}\]

A prograde equatorial jet largely satisfies such a condition, though small-scale violations are not rare. Following McIntyre
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Fig. 1. (a) Zonal wind $\bar{u}$ versus latitude calculated in a simple shallow-water calculation. The profiles demonstrate the effect of PV homogenization on the wind. (b) Corresponding PV profiles. The three profiles in each of the panels are the initial profile (thin solid line), profile with complete PV homogenization (dotted line) and partial homogenization (thick solid line). In the calculation, PV is homogenized between 10° north and south latitudes. PV homogenization decelerates a prograde jet at the equator.

and Palmer (1983)’s criterion, we define breaking Rossby-type waves as overturning in the absolute vorticity gradient (and hence the PV gradient) in latitude. We let $\zeta \equiv \bar{\zeta} + \zeta'$ where $\bar{\zeta}$ is the relative vorticity of the background flow and $\zeta'$ is the wave contribution to the vorticity. For waves to break, we must approximately have

$$\left| \frac{d\zeta'}{dy} \right| > \beta + \frac{d\bar{\zeta}}{dy},$$

where $\beta \equiv 2\Omega/a$.

In absence of the background flow, we easily see that the wave-induced latitudinal relative vorticity gradient $d\zeta'/dy$ needs to overcome only $\beta$ to cause wave breaking. On the other hand, waves in a prograde equatorial jet must overcome the background vorticity gradient $d\bar{\zeta}/dy$ in addition to $\beta$. (Except for the possibility of a narrow strip at the equator, Saturn’s equatorial jet has cyclonic relative vorticity, meaning that $d\bar{\zeta}/dy$ and $\beta$ both have the same sign—positive.) For example, with the equatorial jet we considered earlier, waves can only induce breaking if the amplitude $|d\zeta'/dy|$ is at least $\sim 20\%$ larger than that necessary to induce breaking when $\bar{u} = 0$. A flow becomes highly turbulent when waves break. Thus, inhibition of breaking waves also prevents turbulent homogenization.

3. Model setup

To test whether a GWS-like convective storm can slow down the equatorial jet on Saturn, we use the Explicit Planetary Isentropic Coordinate (EPIC) atmosphere model developed by Dowling et al. (1998) to run our numerical experiments. The model solves the primitive equations in oblate spherical coordinates using $\theta$, the potential temperature, as the vertical coordinate.

For all of our simulations, the initial temperature profiles are initialized such that the levels above 200 mbar are isothermal at 100 K and those below have a constant Brunt–Väisälä frequency of 0.005 s$^{-1}$. This thermal structure crudely mimics that of a jovian planet as done by Showman and Dowling (2000). In some of the simulations, the temperature profiles are relaxed to this initial structure using Newtonian cooling with the radiative time constant profile for Saturn by Conrath et al. (1990). Because of the long radiative-time-constants, however, this relaxation has a negligible effect in our relatively short (<200-day) simulations. The planet’s equatorial radius, polar radius, rotation rate, surface gravity, and specific gas constant adopted in our simulations are 60,330 km, 54,180 km, 1.638 × 10$^{-3}$ s$^{-1}$, 8.96 m s$^{-2}$, and 3900 J K$^{-1}$ kg$^{-1}$, respectively. The simulations use the ideal gas equation of state.

The domain size of the simulations we present here is 120° in longitude, 80° in latitude centered on the storm latitude, and 10 bar–0.18 mbar in the vertical. The longitudinal × latitudinal × vertical resolution for our nominal cases is 128 × 90 × 48. A few simulations with higher horizontal resolutions are also performed to test for model convergence and study the effects of small scale eddies. In the vertical, the model is initialized with layers evenly spaced in log-pressure at the southern domain boundary. For simulations with vertical wind shear, the initial layer spacing becomes weakly dependent on latitude following the thermal wind equation. The layer spacing roughly corresponds to a vertical resolution of 5 km around 50-mbar and 10 km around the 500-mbar levels. The top 16 layers are set up as the “sponge” to reduce wave reflection at the model top. The lowest layer is a non-evolving abyssal layer representing the adiabatic interior. Consequently, the highest active (i.e., non-sponge) layer is at an isentropic layer corresponding to ∼9 mbar (with lowest sponge layer being at ∼7 mbar) and the lowest time-evolving layer is at ∼9 bar. The timestep is 50 s for the nominal resolution simulations.

The nominal-case simulations use the sponge layers that relax $u$ to the initial profile and $v$ to zero. In order to assess the effect of the sponge layers to the top active layers in our model, we performed the following sensitivity tests. First, we placed the sponge higher in the atmosphere so that the lowest sponge layer is at 1 mbar. Second, we used modified sponge layers which relax zonal wind $u$ to its zonal mean so that the sponge does not act as a momentum source/sink. Although placing the sponge higher allows the wind to change at greater magnitudes at higher altitudes, wind responses below the 20-mbar level are very similar in analogous simulations using the alternative sponge settings and the nominal sponge. Also, as will be discussed in Section 4.2, the sponge settings have much less influence on the wind responses when a simulation has an initial wind with a realistic vertical shear.

Hyperviscosity (horizontal hyperdiffusion) coefficient $\nu_4 = 4.131 \times 10^{18}$ m$^4$ s$^{-1}$ is used for our nominal resolution simulations. The value represents a minimal value necessary to sta-
ibilize a simulation with our strong forcing. Fourth-order hyper-
diffusion’s characteristic smoothing timescale $\Delta t$ for a length
scale $\Delta x$ can be written $\Delta t \approx \frac{\nu}{(\Delta x)^4}$. With the aforementioned $\nu$ value, a wave with wavelength $\Delta x \approx 6000 \text{ km}$ (6 grid
points) is damped in timescale $\Delta t \approx 3 \times 10^8 \text{ s} \approx 3600 \text{ days}$,
which is an order of magnitude longer than our simulations.
Thus, waves with wavelengths exceeding $\sim 6000 \text{ km}$ are rela-
tively unaffected by the hyperviscosity. This is also sufficient
to ensure that the overall strength and behavior of the equato-
rial jet is not affected. Note that, in EPIC, hyperviscosity acts
only on horizontal wind shear; the model contains no explicit
diffusion in the vertical. Numerical inaccuracies in the conserva-
tion laws, including the so-called “numerical diffusion,” can
potentially become pronounced when material advects across
grid surfaces. Because EPIC uses potential temperature as the
vertical coordinate, adiabatic motions are completely confined
in each layer. Diabatic motion, i.e., mass exchange between isen-
tropic layers, occurs only due to radiation, which is irre-

evant to our project, thus numerical artifacts on the vertical
transport in our simulations are minimal. This is a key advan-
cage of EPIC over other general circulation models (GCMs)
that use pressure as the vertical coordinate. We have also com-
pared simulation results with varying vertical resolutions, and
are confident that 48 layers is sufficient to ensure that storms’
large-scale effects on zonal winds are independent of the verti-
cal resolution.

To test the convergence of our results, simulations with
higher horizontal resolution and lower hyperviscosity were per-
formed. Although these simulations result in wind changes with
spatial patterns slightly different from our nominal cases, the
magnitudes of the equatorial jet speed changes are all similar.
This indicates that our simulations adopted sufficiently high
resolution and low hyperviscosity to produce qualitative re-
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cal resolution.
\( r_n \) is the mass injection peak time for the \( n \)th storm cell. \( \tau_s \) is the characteristic life time of a cell, which is the same for all cells in a simulation. In our simulations, the mass injection is initiated (terminated) \( 3\tau_s \) before (after) \( \tau_s \) for each cell. \( r_n(\lambda, \phi) \) is the arc-distance from \((\lambda_n(t), \phi_n)\), the center of the \( n \)th storm, in degrees:

\[
\begin{align*}
 r_n(\lambda, \phi) &= \frac{180^\circ}{\pi} \arccos(\cos \phi \cos \lambda \cos \phi_n \cos \lambda_n \\
 &\quad + \cos \phi \sin \lambda \cos \phi_n \sin \lambda_n + \sin \phi \sin \phi_n). \quad (12)
\end{align*}
\]

Only the longitude \( \lambda_n(t) \), and not the latitude \( \phi_n \), is time dependent because we assume that storm cells advect with the abyssal flow, which has no latitudinal component. \( r_s \) is the characteristic arc-radius of a storm cell in degrees. All storm cells have the same \( r_s \) in a simulation. Equation (12) does not correctly calculate the arc-distance when the longitudinal boundary periodicity is not 360\(^\circ\); however, an appropriate treatment is done in the simulations so that the calculation is performed correctly for all longitudinal domain sizes.

\( Z(\theta) \) represents the result of vertical convective mass flux. Physically, it depends on how a convective plume mixes with the ambient air as it ascends and detrains mass at different isentropic levels. Its unit, \( m^{-1} s^2 Pa K^{-1} \), is equivalent to \( kg m^{-2} K^{-1} \), thus this function times \( \alpha \) represents the total amount of mass detrained by a cumulus cell per area per potential temperature level. Also, note that \( Z(\theta) \) has the same dimension as the layer thickness \( h \), thus another way to paraphrase its nature is that the function represents the change in the local layer thickness a storm cell would cause in absence of dynamics (in reality, diffusion and advection will transport some of the injected mass away from the mass injection site and the resulting layer thickness change will be smaller). In our simulations, we treat this \( Z(\theta) \) as a free parameter since it has never been measured on Saturn, and choose parameter values that maximize the storm’s dynamical effects on the equatorial jet. Fig. 2 shows the \( Z(\theta) \) profiles used in simulations we present in this paper. The nominal profile, in the solid line, places the majority of mass injection close to the expected condensation altitude near ~10 bar. We chose to present simulations using this particular \( Z(\theta) \) function primarily because our study’s goal is to test the extreme upper limit of the effects a cumulus event may have on the equatorial jet, and detraining most of the mass lower in the atmosphere allows the largest amplitude mass forcing (i.e., amount of total mass injected) in our simulations without drastically altering the background thermal structure.

This formulation makes the total mass injected by a storm proportional to \( N_o \alpha r_s^2 \) (this proportionality does not hold when comparing storms with different \( Z(\theta) \) profiles). Note that it is independent of \( \tau_s \) because \( T(t) \) is a normalized gaussian in time. A storm with \( N_o \alpha r_s^2 = 1.0 \) degree\(^2\) injects \( \sim 4.6 \times 10^{18} \) kg of mass when the nominal \( Z(\theta) \) profile is applied, and \( \sim 2.3 \times 10^{18} \) kg when the alternative profile is used. As a result, with the amplitude factor \( \alpha = 1.0 \), a storm cell with the nominal (alternative) \( Z(\theta) \) profile raises the pressure on the lowest layer at its center by almost 120 bar (60 bar) in absence of dynamics, i.e., without diffusion or advection. This is an extreme upper limit, and in Section 4, we present simulations with \( \alpha \) values ranging from 1/64 to 0.5, and \( N_o \alpha r_s^2 \) values ranging from 4.0 to 32.0 degree\(^2\). Based on the initial horizontal expansion rate of the GWS spot, Sánchez-Lavega (1994) estimates an average vertical wind velocity \( w \approx 0.1 \) m s\(^{-1}\) in the spot over a horizontal length scale of 20,000 km. Assuming this vertical velocity at the 100-mbar level with temperature of 100 K, and using ideal gas law with the gas constant \( R = 3900 \) J kg\(^{-1}\) K\(^{-1}\), we estimate the average vertical mass flux in the spot to be \( \sim 2.6 \times 10^{-3} \) kg m\(^{-2}\) s\(^{-1}\). In a circular area with 20,000-km diameter over a time period of a month, this gives an estimate for the total vertical mass transport of \( \sim 2 \times 10^{18} \) kg by the GWS spot, and all storms in our simulations inject at least an order of magnitude more mass than this estimate.

Note that our storm parameterization scheme injects mass moving at the velocity of the local wind field where it is added; thus, we do not take account of convective momentum transport that may occur in a GWS-like large storm. However, we have no reason to believe this effect is significant as we described in Section 2.1. Also, 3D anelastic simulations by Hueso and Sánchez-Lavega (2004) showed that, at low latitudes on Saturn, westward Coriolis acceleration acting on ascending vertical motions in a cumulus cell may be significant. However, they used rigid, free-slip conditions for their lateral boundaries, and their simulation domain was not significantly larger than the area covered by the convective plume. It is therefore difficult to apply their result to the large-scale environment. We do not include this effect primarily because the main focus of our study is the effect of atmospheric waves, but we also believe that the westward Coriolis acceleration of an ascending plume cannot significantly affect the zonally averaged zonal wind. Plumes ascending at a vertical wind speed \( w \) experience westward Coriolis force per volume \( -\rho w \Omega \) where \( \rho \) is the mass density. Assuming that the plumes cover a fractional area \( \sigma \ll 1 \), the average westward force per volume over the planet is \( -\sigma \rho w \Omega \). Through mass continuity, the net vertical mass flux must be zero and an upwelling must be accompanied by subsidence elsewhere, with a mean descending speed.
\(-\sigma w/(1 - \sigma) \approx -\sigma w\), and the average Coriolis force per volume it experiences becomes \(+\sigma pw\Omega\). We thus suspect that the westward acceleration caused by ascending air and eastward acceleration caused by descending air largely cancel out, leading to no net acceleration of the layer by this mechanism. Nevertheless, this remains an open issue that must be resolved by 3D convection simulations that properly treat the interactions between the storm plumes and the large-scale environment.

García-Melendo et al. (2005) also suggest that the fluid dynamic stability of an atmospheric zonal jet against perturbations may depend on the vertical thermal structure, and we do not explore this possibility here. García-Melendo et al. tested a scenario in which a disturbance, observed in 1990 in the fastest prograde jet of Jupiter, triggered instabilities that resulted in a slowdown of the jet. As noted by García-Melendo et al., this jovian disturbance resulted in multiple vortices, which is a characteristic behavior of fluid-dynamic flow instabilities. However, there is no reason to believe that the 1990 saturnian disturbance triggered by the GWS was the same type of phenomenon. For example, no vortices were observed in the mature stage of the disturbance as noted by Sánchez-Lavega (1994). HST observations by Westphal et al. (1992) show cloud features that can be interpreted as breaking wave modes such as a Kelvin–Helmholtz instability; however, their spatial scales were much smaller than the extent of the jet, and there is no evidence that the equatorial jet became unstable. Also, if Saturn’s equatorial jet is unstable, the extremely strong forcing in our simulations would have revealed so.

The EPIC model by Dowling et al. (1998) uses third-order Adams–Bashforth timestepping scheme; however, its original implementation contained an erroneous multiplication of the mass source by a factor of 23/12 upon updating the values of \(h\) in the continuity equation (Eq. (18) in Dowling et al., 1998). This error has been fixed as done by Showman (2006).

4. Numerical experiments

In this section, we present the results of our numerical experiments. We performed simulations with two different initial wind profiles, and compare their results to illustrate a GWS-like intense storm’s effects on the zonal wind. First, in Section 4.1, we show results of simulations with no initial wind. Second, in Section 4.2, we present simulations initialized with a realistic equatorial jet decaying with altitude. In the realistic initial wind cases, nonlinear interactions between the storm-triggered disturbances and the background flow have important effects on the outcome, and we find it useful to compare their results against the zero initial wind counterparts in analyzing the effects of the storm-induced disturbance on the zonal winds.

In our simulations, we vary the following storm parameters to explore the storm’s effects. Changing the storm’s latitude alters the wind response. We vary the storm center between the equator and \(10^\circ\) N. \(r_s\) and \(\tau_s\) control the characteristic scale of a disturbance, e.g., smaller \(r_s\) and \(\tau_s\) tend to excite short-wavelength, high-frequency waves such as gravity waves, while greater values more effectively generate planetary-scale waves including Rossby and Kelvin modes. We choose the total duration and horizontal size for the simulated storms motivated by the observed GWS 1990 disturbance, which had an initial characteristic horizontal size of \(\sim 20,000\) km followed by a rapid east–west expansion over a \(\sim 3\)-week period (Sánchez-Lavega, 1994). Larger \(\alpha\) values generate higher-amplitude waves. We tune \(\alpha\) to maximize the amplitude of the storm-excited waves, thus maximizing the momentum flux. \(N_s\) is varied to test the effects of a multi-celled storm on the winds. A \(N_s = 1\) storm represents a smooth convective mass flux in both time and space, while the large \(N_s\) cases represent more realistic storms with multiple convective cells. Because each storm cell acts as an independent wave source, a large \(N_s\) storm results in a substantially more wave radiation than a smaller-\(N_s\) storm. The increased eddy activities triggered by a large-\(N_s\) storm also help homogenize the PV. Reports of multiple bright nuclei in the GWS 1990 (Beebe et al., 1992; Sánchez-Lavega et al., 1991) hint that the GWS “spot” was a large anvil cloud under which multiple convective cells were active as is the case for thunderstorms on Earth. As noted in Section 3, we choose the values of \(\alpha\) and \(N_s\) to test the extreme upper limit of storm forcing. When comparing different simulations, we keep the total mass added by the storms, which is proportional to \(N_s\alpha r_s^2\), to be approximately constant. See Table 1 for the list of runs presented in this paper and their parameters.

4.1. Zero initial wind cases

4.1.1. Effects on the zonal wind

Figs. 3a–3d show typical zonal wind patterns produced by storms in zero initial wind cases. The figures display the zonally averaged \(u\) calculated on isobaric surfaces and projected on the latitude–pressure (YP) plane. Figs. 3a–3c display the wind produced by a \(N_s = 1\) storm centered on the equator (Z0N1 in Table 1), \(5^\circ\) N (Z5N1) and \(10^\circ\) N (Z10N1), respectively. Fig. 3d shows the wind generated by a \(N_s = 80\) storm centered on \(10^\circ\) N (Z10N80). All results shown in Fig. 3 are on the day 170 of the respective simulations. We compare the day 170 of each simulation because storm-triggered transient disturbances have mostly decayed by then. As the figures show, the magnitude of the wind changes vary depending on the storm parameters while their spatial patterns are very similar. Above the 100-mbar level, the prevailing wind response in the zero-initial wind simulations is the substantial westward acceleration of the equatorial stratosphere. This wind behavior indicates that the momentum transport mechanism at work here is not a simple diffusion, which cannot produce local extrema of momentum (furthermore, EPIC has no vertical diffusion). A non-local momentum transport mechanism such as atmospheric waves must be responsible for transporting the westward momentum to the stratosphere to cause this wind behavior.

Fig. 4a compares, at the 10-mbar level, wind responses of the cases Z0N1 (dotted line), Z5N1 (dashed), Z10N1 (dot-dashed), and Z10N80 (solid). We find that an \(N_s = 1\) storm causes a stronger westward acceleration of the equatorial stratosphere when the storm is further away from the equator. This westward acceleration can be viewed as a “slowdown” in the sense that
we define eastward as positive. Storms with larger \( N_s \) (and with correspondingly smaller \( r_s \) and \( \tau_s \) values) than the \( N_s = 1 \) cases result in stronger stratospheric wind responses. However, the resulting wind did not vary significantly for \( N_s \geq 80 \) cases, and the strongest equatorial “slowdown” we obtained is \( \sim 80 \text{ m s}^{-1} \) at the 10-mbar level as shown in Fig. 4a. In our zero initial wind simulations, the storms cause significant westward acceleration at the highest active (i.e., non-sponge) layer. Our nominal case simulations place the highest active layer at \( \sim 9 \) mbar, and at this altitude, the Z10N1 storm causes a \( \sim 30 \text{ m s}^{-1} \) westward acceleration. When the sponge is placed at a higher altitude such that the highest active layer is at the 1-mbar level, a storm identical to Z10N1 causes a \( \sim 70 \text{ m s}^{-1} \) westward acceleration at the top active layer; nevertheless, the wind responses at the 20-mbar level \( \sim 10 \text{ m s}^{-1} \) westward in both cases and below are similar in the nominal and alternative sponge settings. We emphasize that this dependence on the sponge settings occurs only in the zero initial wind simulations at the highest model levels, and the wind behaviors are much less dependent on the sponge layers in the realistic wind simulations presented in Section 4.2. Our zero initial wind results conclusively show that the storm-triggered disturbances induce upward flux of westward momentum, and multi-cell storms trigger greater momentum flux than the single-cell counterparts for a given storm total mass.

The wind responses at low altitudes are illustrated in Fig. 4b, which compares the 2-bar level winds generated by the same storms as shown in Fig. 4a. To the north of the storm center, the wind accelerates eastward. Between the storm center and the equator, the wind accelerates westward and the acceleration turns eastward to the south of the equator. When the storm is centered on the equator, no westward acceleration of the wind occurs below 100 mbar. These responses are consistent with the conditions necessary to balance a pressure bulge created by the storm. These wind changes become stronger when more mass is injected by the storm. The low-level wind responses do not have a strong dependence on parameters other than the storm center latitude and the total mass injected by the storm.

The initial development of the storm also illustrates the wind’s response to the storm. Figs. 5a–5c show the pressure in grayscale and wind vectors in arrows on an isentropic layer corresponding to the \( \sim 400\)-mbar level, on days 10, 30, and 50, respectively, of the Z10N1 storm. High pressure on an isentrope is equivalent to high temperature on the corresponding isobar. The figures illustrate that the geostrophic adjustment to the storm’s mass injection initially creates an anticyclonic circulation around the storm center (60° longitude). The strong \( \beta \) effect prevents formation of a vortex, however, and wave radiation spreads the injected energy over the full range of longitudes. By day 50, perturbations in zonal wind exceed those in meridional wind. The net eastward flow at latitudes \( \sim 18^\circ \)–\( 30^\circ \) N and 8°–20° S, and the net westward flow at 0°–10° N (which can be seen on Fig. 3c) are evident in Fig. 5c.

### 4.1.2. Waves excited by the storms

The storms in our simulations excite atmospheric waves. Fig. 6 shows the contours of \( u \) on equatorial longitude–pressure (XP) cross-section. The wave shown here is a snapshot of the case Z10N1 on day 50 of the simulation; similar waves are present in all zero initial wind cases. The \( N_s = 1 \) storms predominantly couple to long-wavelength waves as shown in Fig. 6. At the 1-bar level where \( \bar{u} \approx 0 \text{ m s}^{-1} \), this wave-like

---

### Table 1

<table>
<thead>
<tr>
<th>Identifier</th>
<th>Storm center lat. (degree)</th>
<th>Number of storm cells ( N_s )</th>
<th>Storm duration (day)</th>
<th>Storm cell decay time ( \tau_s ) (day)</th>
<th>Storm cell radius ( r_s ) (degree)</th>
<th>Storm cell amplitude ( \alpha )</th>
<th>Total storm mass ( \text{Kg} )</th>
<th>Resolution</th>
<th>Horizontal hyperviscosity ( \nu_4 ) (m^4 s^{-1})</th>
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</thead>
<tbody>
<tr>
<td>Z0N1</td>
<td>0</td>
<td>1</td>
<td>21.0</td>
<td>10.5</td>
<td>8.0</td>
<td>1/16</td>
<td>18.4</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
</tr>
<tr>
<td>Z5N1</td>
<td>+5</td>
<td>1</td>
<td>21.0</td>
<td>10.5</td>
<td>8.0</td>
<td>1/16</td>
<td>18.4</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
</tr>
<tr>
<td>Z10N1</td>
<td>+10</td>
<td>1</td>
<td>21.0</td>
<td>2.1</td>
<td>2.0</td>
<td>1/64</td>
<td>23.0</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
</tr>
<tr>
<td>Z10N80</td>
<td>+10</td>
<td>80</td>
<td>21.0</td>
<td>2.1</td>
<td>2.0</td>
<td>1/64</td>
<td>23.0</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
</tr>
<tr>
<td>V0N1</td>
<td>0</td>
<td>1</td>
<td>21.0</td>
<td>10.5</td>
<td>8.0</td>
<td>1/2</td>
<td>147.2</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
</tr>
<tr>
<td>V5N1</td>
<td>+5</td>
<td>1</td>
<td>21.0</td>
<td>10.5</td>
<td>8.0</td>
<td>1/2</td>
<td>147.2</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
</tr>
<tr>
<td>V10N1</td>
<td>+10</td>
<td>1</td>
<td>21.0</td>
<td>10.5</td>
<td>8.0</td>
<td>1/2</td>
<td>147.2</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
</tr>
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<td>4.0</td>
<td>1/64</td>
<td>92.0</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
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<tr>
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<td>320</td>
<td>63.0</td>
<td>0.525</td>
<td>2.0</td>
<td>1/40</td>
<td>147.2</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
</tr>
<tr>
<td>V10N1280</td>
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<td>42.0</td>
<td>0.525</td>
<td>1.0</td>
<td>1/64</td>
<td>92.0</td>
<td>256 × 180 × 48</td>
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</tr>
<tr>
<td>V10N80c</td>
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<td>21.0</td>
<td>0.525</td>
<td>4.0</td>
<td>1/64</td>
<td>92.0</td>
<td>128 × 90 × 48</td>
<td>4.131 × 10^18</td>
</tr>
</tbody>
</table>

\( a \) \( N_s, \tau_s, r_s, \) and \( \alpha \) are defined in Section 3 of the text.

\( b \) Here, we define storm duration as the interval between times \( t_1 - \tau_s \) and \( t_{N_s} + \tau_s \), where \( t_1 \) and \( t_{N_s} \) are the peak times for the first and last storm cells, respectively. The peak time for the \( n \)th storm cell \( t_n \) is defined by Eq. (10).

\( c \) Total mass injected by a storm is proportional to \( N_s \alpha r_s^2 \). One \( N_s \alpha r_s^2 = 1.0 \text{ degree}^2 \) storm with the nominal \( Z(\theta) \) profile injects approximately \( 4.6 \times 10^{18} \text{ Kg} \) of mass in the simulation. All simulations listed in this table use the nominal \( Z(\theta) \) profile, the solid curve in Fig. 2.

\( d \) The grid resolutions are denoted as longitude × latitude × pressure. All simulations have the same simulation domain size.

\( e \) The storm cells in this simulation are advected collectively as a cluster unlike in the other Voyager initial wind simulations.
structure propagates westward at 46 m s\(^{-1}\) with respect to the model coordinates rotating at the System III rate, \(\Omega = 1.638 \times 10^{-4}\) s\(^{-1}\). The wave’s phase speed far exceeds the wind speed everywhere in the simulation, implying that the westward propagation must result from wave dynamics rather than passive advection of a static structure. The structure and behavior of this wave is characteristic of Rossby-like modes (e.g., see Fig. 4.21 of Andrews et al. (1987) for the westward-propagating Rossby-gravity wave structure). The fact that Rossby-type waves cause upward flux of westward momentum (e.g., see Section 8.3.2 of Andrews et al., 1987) is consistent with the westward acceleration of the equatorial stratosphere in our simulations.

Storms with fast \(r_s\), small \(\tau_s\) and large \(N_s\) excite a much wider spectrum of waves. In such simulations, the wave amplitudes become large enough at high altitudes that the waves break. Fig. 7 shows contour plots of PV on an isentropic layer corresponding to the \(\sim\)10-mbar level on days 10, 60, and 80 of the simulation Z10N80. Here, we define “wave breaking” following McIntyre and Palmer (1983)’s criterion. Breaking waves manifest as overturning of PV contours in latitude at several longitudes. As a result of wave breaking, PV becomes a non-monotonic function of latitude and then homogenizes as shown on day 80, where the typical contour spacing in the region of wave breaking becomes much wider than on day 10. The latitudes where the waves break roughly correspond to the location of the westward acceleration, and this most likely contributed to the much faster westward wind here than in the \(N_s = 1\) cases in addition to the high-frequency wave modes absent in the \(N_s = 1\) cases. Such breaking waves play very little role in the \(N_s = 1\) cases and result in minimal PV homogenization. Although not shown, a small \(r_s\), fast \(\tau_s\) storm cell also generated substantial gravity waves.

4.2. Voyager initial wind cases

Now we present simulations with a more realistic initial wind profile. Our initial wind follows the Voyager wind profile (Sánchez-Lavega et al., 2000) below the 1-bar level and gradually becomes slower with altitude in a manner crudely mimicking the vertical shear revealed by Flasar et al. (2005). Unlike in the zero initial wind cases, the storm cells are advected eastward at \(\sim\)400–460 m s\(^{-1}\) by the abyssal flow at the cells’ respective latitudes in these simulations. In our Voyager initial wind simulations, minor but non-negligible hyperviscous
smoothing occurs. Thus, we measure our storm’s effects by comparing the storm-affected winds against that of a simulation without a storm, which contains only the hyperviscous effects. The zonal average \( \bar{u} \) unaffected by a storm but smoothed by hyperviscosity for 170 days is shown in Fig. 8, projected on the YP-plane. Hereafter, we will denote the zonal wind unaffected by a storm as \( \bar{u}_0 \). It should be emphasized that, in a simulation without a storm, hyperviscosity leads to no change larger than 10 m s\(^{-1}\) in the equatorial jet’s peak speed at any altitude and the wind speed changes mostly occur at the flanks of the jet. Although the particular simulation shown in Fig. 8 has a domain identical to the 10° N storm simulations, all jet speed change measurements are made against a stormless simulation with the same simulation domain. For the large \( N_s \)-storm simulations with Voyager initial winds, we focus on the 10° N-centered storms as we found that they had greater effects on the stratospheric winds than the lower latitude-centered ones. This choice is also motivated by the observation of the 1990 GWS event, which initially had a bright nucleus at \( \sim 12° \) N on September 25, and the spot subsequently expanded and shifted its center to \( \sim 5° \) N by October 5 (Sánchez-Lavega et al., 1991). In each multi-cell-storm simulation, the \( N_s \) storm cells are triggered randomly in an 8° radius circle centered on 10° N as in the zero initial wind case. However, the cells do not move together as a cluster in the \( N_s > 1 \) simulations presented in this section because each cell advects away from the initial location.

Fig. 4. The \( \bar{u} \) profiles for the zero initial wind simulations at (a) 10-mbar level and (b) 2-bar level. Both panels show the \( \bar{u} \) profiles on the day 170 of simulations Z0N1 (dotted), Z5N1 (dashed), Z10N1 (dot-dashed), and Z10N80 (solid).

Fig. 5. Panels (a)–(c) show pressure (grayscale) and wind fields (arrows) on an isentrope near the 400-mbar surface for the simulation Z10N1 on days 10, 30, and 50, respectively. A lighter color in the grayscale denotes higher pressure. The vector arrows’ lengths are proportional to the wind speed. On day 10, pressure variation on the isentrope is 0.06 mbar, and the fastest wind is 0.03 m s\(^{-1}\). On day 30, the pressure variation is \( \sim 7 \) mbar and the fastest wind is 2.73 m s\(^{-1}\). On day 50, the pressure variation is \( \sim 20 \) mbar and the fastest wind is 9.0 m s\(^{-1}\). The storm’s mass injection creates pressure gradients on isentropic surfaces, and subsequent geostrophic adjustment generates winds and waves.
Fig. 6. Equatorial cross-section of \( u \) on day 50 of the simulation Z10N1. The figure shows waves propagating westward at 46 m s\(^{-1} \) measured at the 1-bar level. The waves’ vertical structure and propagation direction are highly characteristic of Rossby-type waves. \( u \) varies from \(-4.5\) to \(5.7\) m s\(^{-1} \) on this panel.

after it is activated. The cells are activated one at a time evenly spaced in time over the storm durations listed in Table 1.

4.2.1. Effects on the zonal wind

Fig. 9 shows the responses of the zonal winds to the \( N_s = 1 \) storms centered on the equator (V0N1 in Table 1), 5° N (V5N1) and 10° N (V10N1) on day 170 of the respective simulations. Figs. 9a–9c show \( \bar{u} \) affected by the storm, and Figs. 9d–9f display the zonal wind speed change triggered by the storm (i.e., \( \bar{u} - \bar{u}_0 \)), respectively, projected on the YP plane. Figs. 9d–9f reveal that, unlike in the zero initial wind cases, none of the \( N_s = 1 \) storms here cause a significant slowdown of the equatorial jet at any pressure level even though these simulations’ storms injected more mass by a factor of eight than the zero initial wind \( N_s = 1 \) cases shown in Fig. 3. Below the 100-mbar level, the spatial patterns of wind speed changes in response to the storms here are very similar to those in the zero initial wind counterparts. The storm slightly widens the equatorial jet in all three cases, compared to the stormless result shown in Fig. 8. Note also that the storms V5N1 and V10N1 shifted the peak of the jet southward, away from the storm center. Similar shifts of the peak latitude occurs in the 10° N-centered large \( N_s \) simulations presented later. In these simulations, the storm accelerates the wind westward (eastward) to the north (south) of the equator; consequently, the wind change shifts the equatorial jet’s peak southward, but this jet shift does not represent a shift of the air mass. Hints of a similar equatorial jet shift are found in Barnet et al. (1992)’s HST wind speed measurements on November 17–18, 1990, approximately 50 days after the onset of the GWS, in which the equatorial jet’s peak latitude is shifted southward of the Voyager–observed profile. Fig. 8 of Sánchez-Lavega et al. (2000) compares the Barnet et al. profile against the Voyager wind. The HST measurement at 546 nm has its jet peaked at \(~-3^\circ\) S whereas the Voyager profile peaks at \(~5^\circ\) N. If the observed difference in the jet peak latitude was caused by the GWS 1990, it may signify a substantial mass transport to the cloud level by the cumulus convection as modeled here.

Fig. 10 shows results of the 10° N-centered large \( N_s \) storm simulations, V10N80, V10N1280, and V10N320, in the same format as in Fig. 9. Fig. 10 shows that the spatial patterns of the wind changes are qualitatively similar to those of the zero initial wind cases shown in Fig. 3, characterized by the westward acceleration of the equatorial stratosphere and the eastward accelerations of the low-level winds in the latitudes surrounding the storm. However, the stratospheric wind slowdowns occur over a wider range of altitudes and they are much weaker in the

Fig. 7. PV distribution on an isentropic surface corresponding to the 10-mbar level on days 10, 60, and 80 of the simulation Z10N80. The unit of the contour values is \(10^{-4} \) K Pa\(^{-1} \) m\(^{-1} \) s. The contour spacing is the same for all four panels. The figures show that waves cause overturning of the PV gradient and break to homogenize PV.
Voyager initial wind simulations than the stratospheric westward accelerations in the zero-wind counterparts. Figs. 10a and 10d show $\bar{u}$ and $u - \bar{u}_0$, respectively, of the simulation V10N80. Fig. 10d shows that the storm decelerated the equatorial jet by $\sim 30$ m s$^{-1}$ in the stratosphere centered around the 15-mbar level, and the slowdown is much weaker at the lower altitudes. Below the 100-mbar level, the spatial pattern of the wind change is very similar to the $N_z = 1$ cases shown in Fig. 9. This qualitative wind behavior does not change when a storm has substantially more cells with smaller size $r_s$. Figs. 10b and 10c show the result of V10N1280, a $N_z = 1280$ storm simulation, which injects an equal amount of mass as V10N80. The 1280 storm cells in V10N1280 is triggered over 42 days, longer than in the V10N80 storm by a factor of two, to reduce the chance of overlapping multiple storm cells. Overlapping cells tend to act as one larger cell and thus effectively weaken the dynamical forcing. The simulation V10N1280 is performed at an increased resolution to resolve the smaller storm cells and the fine-scale disturbances they trigger. While its spatial pattern of the wind change in the stratosphere differ slightly, most likely due to the reduced hyperviscosity in the simulation, the magnitudes of the wind speed changes at most altitudes agree with those in V10N80. A longer-duration storm that injects more mass does not significantly change the outcome. Figs. 10c and 10f show the result of a storm with 320 cells (V10N320) triggered over 63 days. The storm injects 60% more mass than the aforementioned large $N_z$ storms. The storm results in a slightly stronger slowdown around the 10-mbar level, and the increased mass injection results in larger speed changes at the flanks of the jet at lower altitudes; however, no significant change in the peak speed of the equatorial jet occurs below the 50-mbar level. Although Figs. 10d–10f show substantial westward wind accelerations of up to 50 m s$^{-1}$ at lower altitudes ($\rho > 1$ bar), they do not cause a 50 m s$^{-1}$ change in the peak speed of the actual jet. Instead, the pattern of westward accelerations at the storm latitude with eastward acceleration on either flank, coupled with the latitudinal offset of these accelerations from the jet peak, has the net effect of shifting the latitude of the jet’s peak zonal wind while causing only a minimal change in the peak speed. These ideas will be further illustrated in Fig. 11.

None of our simulations produced a strong equatorial jet slowdown at any altitude like that described by Sánchez-Laveaga et al. (2003, 2004). Rather than a large-scale slowdown of the equatorial jet, the storm’s effect is characterized by the eastward accelerations at the mass injection altitudes in the latitudes surrounding the storm, and the weak equatorial jet slowdown in the stratosphere. Fig. 11 illustrates the storm’s effect on the zonal wind profile, using the V10N320 result, which exhibited the largest wind speed changes among the Voyager initial wind simulations. The figure shows V10N320’s results in the dark lines, and those of the no-storm simulation in gray, at 2 bar (solid line), 50 mbar (dashed), and 10 mbar (dot-dashed) altitudes, all on day 170 of the respective simulations. At the 2-bar level, the equatorial jet speed change is negligible; however, the peak latitude shifts southward as discussed earlier. The slowdown is minor at the 50-mbar level, and a $\sim 40$ m s$^{-1}$ slowdown occurs at the 10-mbar level. A careful inspection of the solid curves in Fig. 11 shows that, although the peak jet speed at the 2-bar level remains almost exactly constant at 430 m s$^{-1}$, the actual zonal wind at a particular latitude changes by up to 50 m s$^{-1}$ as the jet shifts southward. This explains how the difference $\bar{u} - \bar{u}_0$ can reach 50 m s$^{-1}$ (Figs. 10d–10f) while not affecting the peak jet speeds in the deep troposphere at $\rho > 1$ bar. We also tested cases with an alternative Z(θ) profile, the dashed curve in Fig. 2, which places most of its mass around the 500-mbar level. We confirmed that such a storm results in a same qualitative wind behavior as the nominal Z(θ) profile cases; an alternative Z(θ) storm shifts the equatorial jet’s peak latitude in a similar manner at the mass-injection altitudes, and decelerates the equatorial jet in the stratosphere. However, the alternative Z(θ) profile storms did not produce any equatorial jet slowdown stronger than the nominal Z(θ) cases.

4.2.2. Waves excited by the storms

The weaker stratospheric wind speed changes in the Voyager initial wind simulations than in the zero initial wind cases are the effect of the vertical and horizontal wind shear in the initial condition. Those shears significantly affect the propagation of waves generated by the storm. An episodic disturbance such as the GWS of 1990 most likely generated atmospheric waves, and the wind speed measurements following the GWS 1990 by Barnet et al. (1992) show wave-like wind speed variations in longitude, although it is possible that some of the wind-speed measurements are affected by waves as will be discussed in Section 5. Below, we illustrate the effects of the background wind shears on the storm-excited wave propagations.

Figs. 12a–12c show the equatorial XP cross-sections of $u$ on day 70 of the simulations V0N1, V5N1 and V10N1, respectively. Figs. 12d–12f show the same cross-sections of the deviation of $u$ from its zonal mean calculated on isobaric surfaces, $u - \bar{u}$, of the same simulations, respectively. These figures show that the phase fronts of the waves are sloped positively (i.e., the constant-phase surfaces rise with increasing longitude) for the V5N1 case and negatively for V0N1 and V10N1. The negatively sloped phase fronts in V0N1 and V10N1 translate
Fig. 9. Panels (a)–(c) show the zonal wind affected by the storms V0N1, V5N1, and V10N1, respectively; the grayscales represent $\bar{u}$, and scale from $-20$ m s$^{-1}$ (black) to 450 m s$^{-1}$ (white). Panels (d)–(f) show changes in the zonal wind caused by the storms, $\bar{u} - \bar{u}_0$ for the same simulations as in panels (a)–(c). The grayscales indicate $-60$ m s$^{-1}$ (black) to +60 m s$^{-1}$, with the neutral gray representing 0 m s$^{-1}$. All three cases slightly widen the equatorial jet at the low altitudes. The cases V5N1 and V10N1 also shift the equatorial jet’s peak latitude southward.

Eastward at $\sim 370$ m s$^{-1}$ (measured at the 1-bar level) with respect to the model coordinates (System III). The zonal mean wind is $\bar{u} \approx 410$ m s$^{-1}$ at that altitude; thus, the phase front propagates westward at $\sim 40$ m s$^{-1}$ with respect to the wind at the 1-bar level. As in the zero initial wind cases, this behavior is highly characteristic of Rossby-type wave modes. The positively sloped phase fronts in V5N1 propagate eastward at $\sim 530$ m s$^{-1}$ with respect to System III, which is $\sim 100$ m s$^{-1}$ eastward with respect to the $\bar{u} \approx 430$ m s$^{-1}$ wind, all measured at 1 bar. This behavior is highly characteristic of the Kelvin wave as described in Andrews et al. (1987), Section 4.7.1. The Rossby-type modes in our $0^\circ$ N and $10^\circ$ N storm simulations are inhibited from propagating above $\sim 100$-mbar level near where the wind speed reaches 370 m s$^{-1}$, the phase speed of the waves. In a flow with a vertical shear, vertically propagating waves are damped at the critical level where the background
Fig. 10. Same as in Fig. 9, but for the large $N_s$ simulations V10N80, V10N1280, and V10N320. All large $N_s$ simulations caused significant equatorial jet slowdowns in the stratosphere, but the slowdown magnitudes are not enough to explain the wind speed difference between the Voyager data (Sánchez-Lavega et al., 2000) and the recent HST observation (Sánchez-Lavega et al., 2003, 2004).

Wind speed approaches the wave propagation speed (Andrews et al., 1987, Section 5.7), and this is precisely what we see in our simulations for the Rossby-type modes (see Figs. 12d and 12f). The Kelvin waves in the $5^\circ$ N case are free to propagate upward because the waves never encounter a critical level.

The westward-propagating waves transport momentum to the critical layers where they are absorbed. We tested alternative sponge settings as discussed in Section 3, and confirmed that, below the 10-mbar level, the Voyager initial wind simulation results are not significantly affected by the sponge; this suggests that most of the waves relevant to the results are absorbed before reaching the sponge layers. The atmosphere is denser at lower altitudes, thus it requires a greater impulse to cause a wind speed change of the same magnitude than at a higher altitude. This interpretation is consistent with our results, in which weaker equatorial wind speed changes occur over a wider range.
of altitude in the Voyager initial wind simulations (shown in Fig. 10) than in the zero initial wind cases (Fig. 3). We also tested a case (not shown) with the equatorial jet decaying at a lower altitude to place the critical level at a lower altitude. The simulation had a storm identical to the one in V10N80. In this alternative vertical shear simulation, the storm-triggered wind-speed changes occurred at a lower altitude with a smaller magnitude, as expected. This suggests that, although the magnitude of wind speed changes are dependent on the background vertical wind shear, a storm in an equatorial jet decaying at a lower altitude than the nominal simulations causes a smaller wind speed change at a lower altitude, and such a scenario does not explain the large equatorial jet speed change revealed by Sánchez-Lavega et al. (2003, 2004). A jet that decays at a higher altitude will most likely result in a greater slowdown at a higher altitude; however, we did not test such scenarios since trackable cloud features are unlikely to exist at such high altitudes.

As in the zero initial wind $N_z = 80$ case, waves break to homogenize PV at high altitudes in the $N_z = 80$ Voyager initial wind simulation. Figs. 13a–13c show PV contours on isentropic layers corresponding to the 50-mbar level on days 10, 20, and 30. In this simulation, waves break at levels above ~60 mbar, homogenize PV, and cause an acceleration which contributes to the equatorial jet slowdown at those altitudes. Note that the extent of wave breaking and the degree of PV homogenization are both much smaller compared to the zero initial wind case in Fig. 7, even though the V10N80 storm injected 4 times more mass than Z10N80. Also, no wave breaking occurs in the $N_z = 1$ Voyager initial wind cases, while hints of breaking waves are found in the $N_z = 1$ zero initial wind simulations (though they play negligible role in PV homogenization). This apparent suppression of Rossby wave breaking in the Voyager initial wind cases most likely occurs because the wave breaking, if any, must happen below the critical level of ~50 mbar, where as in the zero-initial wind cases it can occur as high as the top of the model near 7 mbar. The greater density at deeper levels means that breaking can only occur for much greater wave energy in the Voyager wind cases than in the zero-wind cases. This is consistent with the fact that, in Z10N80, no waves break below the 40-mbar level even though the breaking at the 10-mbar level cause significant overturning of PV contours. The inhibition of wave breaking in the presence of a prograde equatorial jet, shown in Section 2.4, may also play some role in allowing waves to break more easily in absence of the strong prograde equatorial jet.

Our simulation’s PV distribution at the 50-mbar level exhibit spatial patterns qualitatively similar to the wave-like cloud features found in the November 1990 HST observation by Westphal et al. (1992) (Fig. 16d). The observed clouds exhibit a pattern characteristic of breaking waves, although it is clear that the observed waves are not the same type as those in our simulations (Westphal et al.’s features are observed around ~10°–20° N latitudes while the waves in our simulations are equatorially trapped waves). We discuss the effects of wave dynamics on the cloud morphology in the next section.

5. Wave effects on apparent cloud motions

In this section, we discuss the effects of waves on the apparent motion of cloud patterns. Contamination of cloud-tracking wind measurements by waves has been speculated observationally (e.g., Beebe et al., 1996; García-Melendo and Sánchez-Lavega, 2001). Although a few works have been published on the effect of waves on cloud motions (e.g., Stratman et al., 2001; Showman and Dowling, 2000), no numerical simulations have been done in the past to show their effects on cloud-tracking zonal wind measurements. We find that the waves such as ones shown in Fig. 12 have a significant impact on the time evolution of cloud morphologies. Here, we have no intention of drawing any connection between our findings in this section and the apparent equatorial jet slowdown revealed by Sánchez-Lavega et al. (2003).

For this study, we marked the mass injected by the storm with passive tracers and analyzed their motions. The simulation we present below is almost identical to V10N80, except that the storm cells all advect at the 10° N abyssal wind speed and stay as a cluster so that the storm mass and tracers are added to the system from a small coherent source to mimic the release of cloud particles from a single cumulus storm system. The tracers are purely passive, and no effects analogous to evaporation or condensation are included in our simulations.

Figs. 14a–14c show the tracer concentration in grayscale and deviation of wind velocity from its zonal mean (i.e., $\mathbf{u} - \bar{\mathbf{u}}$) in arrows on an isentropic layer corresponding to the 50-mbar level on days 30, 31, and 32 of the simulation. Light colors indicate higher concentrations of tracers and thus the storm-injected mass. Newtonian cooling is turned off for this simulation and the mass injections by the storm cells end on day ~23, so there is no mass source/sink during the period shown in these figures. By day 30, the wind shears have spread the storm tracers to all longitudes around 10° N latitude. The wave-like features in the tracer concentration is produced by wave dynamics as the tracer patterns match that of the wind deviations.
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The wave-like patterns visible in the tracer concentration (Fig. 14) coherently move eastward at \( \sim 20^\circ \text{ day}^{-1} \) in longitude, or \( \sim 240 \text{ m s}^{-1} \), as shown in Figs. 14a–14c. However, the zonal average \( u \) at this altitude and latitude, \( \sim 300 \text{ m s}^{-1} \), is substantially faster than the cloud pattern’s translation rate, which does not even fall within the root-mean-square scatter of winds as shown in Fig. 14d. This result, a cloud feature can move at a speed substantially different from the wind speed, has observational implications. For example, it is possible that apparent motions of the wave-like features noted by Westphal et al. (1992) were poor representations of the local wind speeds. Barnet et al. (1992) employed the cross correlation method on the HST image mosaics to retrieve the zonal wind profiles. The correlation method does not discriminate wave-like fea-
Fig. 13. Panels (a)–(c) show PV distribution on an isentropic surface corresponding to the 50-mbar level in the simulation V10N80 on days 10, 20, and 30, respectively. The unit of the contour values is $10^{-5}$ KPa$^{-1}$ m$^{-1}$ s. The contour lines are drawn for the same PV values for all three panels. The waves excited by the storm on day 10 are seen breaking on day 20 and homogenize PV (i.e., widen the contour line spacings) in the equatorial region on day 30 between around $\pm 5^\circ$ latitudes. Panel (d) shows a map-projected HST image of Saturn at 439 nm on November 17, 1990, $\sim 50$ day after the onset of the GWS 1990 from Westphal et al. (1992). The image shows planetographic latitudes between $7^\circ$ S and $30^\circ$ N and a longitudinal domain $\sim 120^\circ$ wide. The image shows breaking wave-like cloud patterns between $\sim 10^\circ$–$20^\circ$ N latitudes.

Figures from other discreet features, and it is possible that their zonal wind measurements are contaminated by waves.

From the above, we expect that the motion of wave-like cloud features poorly reflects the behaviors of the local winds, at least when the cloud features manifest as the oscillation of the boundary between two cloud bands (as in Fig. 14) rather than as discreet features. This also applies when low image resolution affords tracking of only large-scale cloud features as previously discussed by Beebe et al. (1996). Vertical motions induced by wave dynamics can further affect cloud morphologies through condensation and evaporation of cloud particles (e.g., Showman and Ingersoll, 1998; Friedson, 2005), which are not included in our present analysis.

6. Discussion

Our simulations tested the hypothesis that the GWS of 1990 slowed the equatorial jet on Saturn. Our order of magnitude analyses in Section 2 predicted the following. First, vertical momentum redistribution can cause a large change in the stratospheric zonal mean wind with a minimum impact to the deep wind. Second, PV conservation requires a prograde equatorial jet to decelerate as a result of storm-triggered turbulent homogenization. Third, wave breaking, which induces turbulent homogenization, is inhibited in the presence of a prograde equatorial jet. We ran full-3D numerical simulations to test whether the storm triggered these highly nonlinear effects.

The storms in all our simulations generated atmospheric waves. The zero initial wind simulations demonstrated that Rossby-type waves caused an upward flux of westward momentum. The westward acceleration of the stratosphere is balanced by a moderate eastward acceleration below 100 mbar. Breaking waves homogenized PV, albeit weakly, in all simulations that experienced significant equatorial jet speed changes. The altitudes of the westward acceleration in the equatorial stratosphere ($\sim 10$–$50$ mbar) roughly corresponded to the altitudes of wave breaking. This shows that the breaking waves transferred momentum and affected the wind speed. The comparison between zero and Voyager initial wind $N_s = 1$ cases demonstrated that a prograde equatorial jet decaying with height inhibits PV homogenization. In those Voyager initial wind $N_s = 1$ simulations, no waves broke, PV did not homogenize and the equatorial jet speed did not change at any altitude even though the storms injected substantially more mass than in the storms in
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Fig. 14. Effects of wave dynamics on the apparent cloud motion. Panels (a)–(c) show the concentrations of the storm-injected mass in grayscale and the deviation of the wind vectors from its zonal mean (i.e., $u - \bar{u}$) on an isentrope at the 50-mbar level on days 30–32 of the simulation V10N80t. Lighter colors indicate higher concentrations of the storm mass, and longer arrows represent greater deviation of the wind vectors from the zonal mean. Panel (d) shows the zonal mean $u$ profile (solid) and the zonal root-mean-square scatter of $u$ (dotted). The wave-like patterns in panels (a)–(c), which crests are marked as A, B, and C, translate eastward at $\sim 20^\circ$ day$^{-1}$ in longitude, or $\sim 240$ m s$^{-1}$, a speed much slower than the zonal mean $u$ at the latitude, shown in panel (d).

The zero initial wind $N_s = 1$ cases. We also showed that the large $N_s$ storms generated larger amplitude waves than the $N_s = 1$ counterparts, for a given total storm mass. Breaking waves in these simulations were more prevalent, helped homogenize PV and triggered much greater slowdowns (i.e., westward acceleration) of the equatorial wind in the stratosphere in both zero and Voyager initial wind simulations. We believe that a multi-cellled storm is a more realistic representation of the storm because multiple bright nuclei were observed in the GWS 1990 (Sánchez-Lavega, 1994).

Our simulations reproduced some features found by the HST observations in November 1990, approximately 50 days after the onset of the 1990 GWS. The breaking waves in the large $N_s$ Voyager wind simulations exhibit many qualitative similarities to the wave-like features found by Westphal et al. (1992) (Fig. 13d). Also, our simulations shifted the equatorial jet’s peak latitude southward, away from the storm. Barnet et al. (1992)'s wind profiles hint at a similar southward shift of the jet. Our simulation outputs also suggest that the large longitudinal wind variations observed by Barnet et al. were caused by the storm-triggered disturbances. We also showed that wave dynamics influence the apparent cloud motion even when evaporation and condensation effects are not included.

However, the wind speed changes in our simulations are nowhere near enough to explain the difference between the $\sim 470$ m s$^{-1}$ Voyager equatorial wind in 1980 and the recent measurements of $\sim 275$ m s$^{-1}$. The large $N_s$ Voyager initial wind simulations caused $\sim 15$ m s$^{-1}$ slowdowns at the 50-mbar level; this is not enough to explain the $50–100$ m s$^{-1}$ difference between the wind measurements by Barnet et al. (1992) at 890 nm and the measurements by Sánchez-Lavega et al. (2003, 2004) (multi-wavelengths including 890 nm), both estimated to be at $\sim 50$ mbar. Our result does not explain the $\sim 50$ m s$^{-1}$ difference between the Flasar et al. (2005)'s wind extrapolation at $\sim 50$ mbar and the recent HST measurements either. Our storm mass injection rates are probably an extreme upper limit, and we expect that less massive storms would cause even smaller decelerations of the jet than described here.

Nevertheless, our results show that, if a cumulus storm lifts a substantial amount of mass from the condensation level to an upper level, the storm should shift the equatorial jet peak and also widen the jet at the mass-detrainment level. Cassini VIMS instrument may be able to observe a range of altitudes that may have been influenced by the GWS; however, no analogous observational data exists from the Voyager fly-bys, and we suspect that any change at depth will be difficult to characterize only with Cassini data. Perhaps we should aim at comparing the deep winds in this decade and after the next GWS outburst expected to occur around 2020 based on the event’s quasi-periodicity (Sánchez-Lavega and Battaner, 1987).

Our results do not rule out the possibility that the short-wavelength waves not included in our model caused a further slowdown. On Earth, gravity waves are believed to play a significant role in shaping the stratospheric zonal wind (Fritts and Alexander, 2003). Gravity waves can have very short vertical wavelengths, and if small-scale waves transport enough momentum, we cannot fully rule out the possibility that their
additional contributions are enough to cause a ~50 m s\(^{-1}\) slow-
down at the 50-mbar level. However, as we have mentioned, we tuned our parameters at the extreme upper limit to test the storm’s effects on the dynamics, and we investigated GWSs ranging from a single large storm cell with a diameter of 16° and lifetime of ~21 days to a storm composed of 1280 small cells each 2° in diameter and one day long. Despite the fact that the multi-cell storms produce more small-scale waves, none of our simulations over this wide range of parameters came close to producing the required 50–100 m s\(^{-1}\) slowdown. Therefore, we suspect that it will be difficult to cause a stronger equatorial jet slowdown with wave momentum flux alone. It is doubtful that taking account of a wider spectrum of waves will halve the ~470 m s\(^{-1}\) Voyager wind at any altitude.

Our results strongly suggest that a true temporal equatorial jet slowdown caused by the GWS 1990 alone cannot explain the speed difference between the ~470 m s\(^{-1}\) Voyager wind and the ~275 m s\(^{-1}\) wind observed by Sánchez-Lavega et al. (2003, 2004). Our results, together with the cloud tracer altitude estimates by Pérez-Hoyos and Sánchez-Lavega (2006), indicate that a combination of higher clouds and a true slowdown resulting from producing the required 50–100 m s\(^{-1}\) Voyager wind would slow the speed difference between the GWS 1990 alone and other factors may contribute to the slowdown. Our results, together with the cloud tracer altitude estimates by Sánchez-Lavega et al. (2003, 2004), indicate that a combination of higher clouds and a true slowdown resulting from producing the required 50–100 m s\(^{-1}\) Voyager wind would slow the speed difference between the GWS 1990 alone and other factors may contribute to the slowdown.


