

The emergence of a summer hemisphere jet in planetary atmospheres

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ABSTRACT

Zonal jets are common in planetary atmospheres, and their character, structure and seasonal variability depend on the planetary parameters. During solstice on Earth and Mars, there is a strong westerly jet in the winter hemisphere and weak, low-level westerlies in the ascending regions of the Hadley cell in the summer hemisphere. This summer jet has been less explored in a broad planetary context, both due to the dominance of the winter jet and since the balances controlling it are more complex and understanding them requires exploring a boarder parameter regime. To better understand the jet characteristics on terrestrial planets, and the transition between winter- and summer-dominated jet regimes, we explore the jet's dependence on rotation rate and obliquity. The broad parameter exploration enables a better understanding of the mechanisms controlling the jets in all regimes. Across a significant portion of the parameter space, the dominant jet is in the winter hemisphere, and the summer jet is weaker and restricted to the boundary layer. However, we show that for slow rotation rates and high obliquities, the strongest jet is in the summer rather than the winter hemisphere. Analysis of the summer jet's momentum balance revealed that the balance is not simply cyclostrophic, and that both boundary layer drag and vertical advection are essential. At high obliquities and slow rotation rates, the cross-equatorial winter cell is wide and strong. The returning poleward flow in the summer hemisphere is balanced by low-level westerlies through an Ekman balance and momentum is advected upwards close to the ascending branch, resulting in a mid-troposphere summer jet.

1. Introduction

Zonal jets are ubiquitous features of planetary atmospheres. A common feature of the solar system's terrestrial planetary bodies with a seasonal cycle is a strong jet in the winter hemisphere. Winter jets are observed in Earth's stratosphere and troposphere (e.g., Schneider 2006; Waugh et al. 2017), Mars' troposphere (e.g., Waugh et al. 2016) and Titan's stratosphere (e.g., Flasar and Achterberg 2009). These jets inhibit transport between low and polar latitudes, leading to unique processes in polar regions. For example, polar ozone depletion in Earth's stratosphere (e.g., Schoeberl and Hartmann 1991), condensation and removal of CO₂ in the Martian polar atmosphere (e.g., Toigo et al. 2017), and formation of ice clouds in Titan's polar stratosphere (e.g., de Kok et al. 2014).

In addition to the winter jet, low-level westerly winds in the summer hemisphere close to the Hadley cell's ascending edge are also common in other terrestrial planets in the solar system. On Mars, these low-level westerlies are strongest during the southern summer, and appear in the southern hemisphere (e.g., Haberle et al. 1993; Hinson et al. 1999), around latitude 30°S in Fig. 1. On Earth, low-level winds dominate locally in the Indian monsoon region during the northern summer (e.g., Joseph and Raman 1966; Findlater 1969), around latitude 15°N in Fig. 1. In both cases, the westerlies are close to the surface with easterlies in the upper levels (Fig. 1). Additionally, observations from the waviness of Titan's lake surfaces suggest that low-level winds are stronger during summer (Hayes et al. 2013; Hofgartner et al. 2016), implying that a similar process might be occurring on Titan. These low-level winds also appear in modeling studies. Specifically, Mars paleoclimate studies have shown that the summer jet's strength and characteristics may have changed during Mars's past, depending on its obliquity (e.g., Haberle et al. 1993, 2003).

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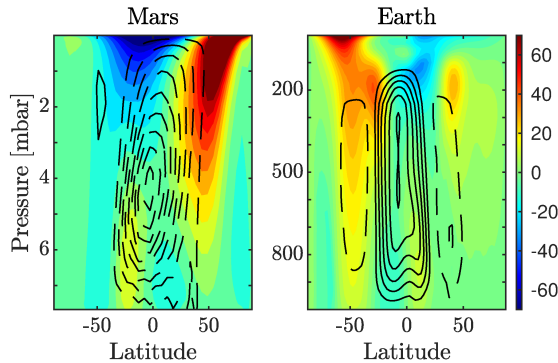


FIG. 1. Zonal mean zonal winds (shading) and mean meridional circulation (contours; solid contours are for counter-clockwise circulation, contour intervals are 5×10^{10} and 2×10^9 kg s^{-1} for Earth and Mars, respectively). Right panel is of the monsoon region on Earth (Era Interim zonally averaged between 40–100E); the circulation is calculated from the divergent component of the meridional velocity (Raiter et al. 2020), using a June–July–August temporal mean. Left panel is of Mars (EMARS (Greybush et al. 2019)), using a 90 solar longitudes (Ls) mean around Ls 270 (southern hemisphere summer solstice, the season where the low-level jet is strong).

Although jets are common, their temporal and spatial characteristics vary among planets. This variability is not well understood, requiring a better understanding of the fundamental processes controlling jet characteristics. Focusing on a specific planet such as Earth, Mars, or Titan can give some insights into these processes; however, these insights will be limited. To better identify the mechanisms responsible for the variability among these planets, we explore, using an idealized three-dimensional general circulation model (GCM) with a seasonal cycle (Guendelman and Kaspi 2019), a wide parameter space spanning both rotation rate (Ω) and obliquity (γ). Previous studies have examined the atmospheric dynamics dependence on different planetary and orbital parameters (Williams and Holloway 1982; Williams 1988; Ferreira et al. 2014; Mitchell et al. 2014; Linsenmeier et al. 2015; Faulk et al. 2017; Wang et al. 2018; Guendelman and Kaspi 2018, 2019; Singh 2019; Lobo and Bordonì 2020). However, these studies have either focused on a single parameter, considered equinox conditions, or have not considered the jet structure.

In section 2, we discuss the model and the simulations used in this study. In section 3, we present the different solstice flow regimes. Surprisingly, for slow rotation rate and high obliquities, we find that the strongest jet is in the summer hemisphere at the ascending branch of a wide, cross-equatorial Hadley cell. To understand the summer jet mechanism, we examine the momentum balances in section 4. We show that in the summer jet, the leading order momentum balance cannot be reduced to geostrophic or cyclostrophic, and that both the boundary layer drag

and the vertical advection play an essential role. Following that, we briefly discuss these regimes in terms of non-dimensional numbers, and we summarize our results in section 5.

2. Model and simulations

In this study, we use an idealized GCM with a seasonal cycle (Guendelman and Kaspi 2019). The surface consists of a uniform slab ocean with a constant 10 m depth that exchanges momentum, moisture, and heat with the boundary layer. Moisture effects are represented by simplified condensation and convection schemes (Frierson et al. 2006), neglecting the effects of clouds and ice. Radiative processes are accounted for through a two-stream gray radiation scheme, where the longwave optical depth varies only with altitude. **Note that keeping the optical depth constant with latitude is a simplification that neglects water vapor feedback in the radiation scheme.**

We run simulations in which we vary both the obliquity (γ), from 10° to 90° , and the rotation rate (Ω), from $1/32$ to 2 times Earth’s rotation rate, at T42 resolution with 25 vertical layers. We use a 360-day orbital period, with zero eccentricity and otherwise an orbital configuration similar to that of Earth. The model is run for 80 years and the climatology is calculated using the latter 50 years. This study focuses on the solstice response and focuses on the northern hemisphere summer. Results presented here are for a time mean between days 150 – 240, equivalent to the June-to-August mean for Earth; the results are insensitive to this period selection.

The model’s results with the closest Earth-like simulation ($\gamma = 30^\circ$, $\Omega = 1$, blue highlighted panel in Figure 2) shows good agreement with the observed flow in Earth’s troposphere, both in terms of the jet strength and latitude (Fig. 1). However, there are some differences between the observed and simulated flow. These differences are due to the model’s simplicity and occur mainly in the stratosphere, where the model fails to reproduce the flow. More specifically, this discrepancy is mainly a result of the simplified radiation scheme (e.g., Tan et al. 2019) and the lack of an ozone layer in the model.

3. Flow regimes

We find that westerly jets occur across the entire parameter space, with variability in their characteristics (Fig. 2). The most dramatic variation is a transition that occurs in high obliquity planets with a slow rotation rate, where the strongest jet is no longer in the winter hemisphere (like in the solar system’s terrestrial planets), but rather in the mid-troposphere of the summer hemisphere (Figs. 2 and 3d). A closer examination of this transition reveals a gradual shift of low-level summer winds, which are observed in Earth’s monsoon region and on Mars, towards

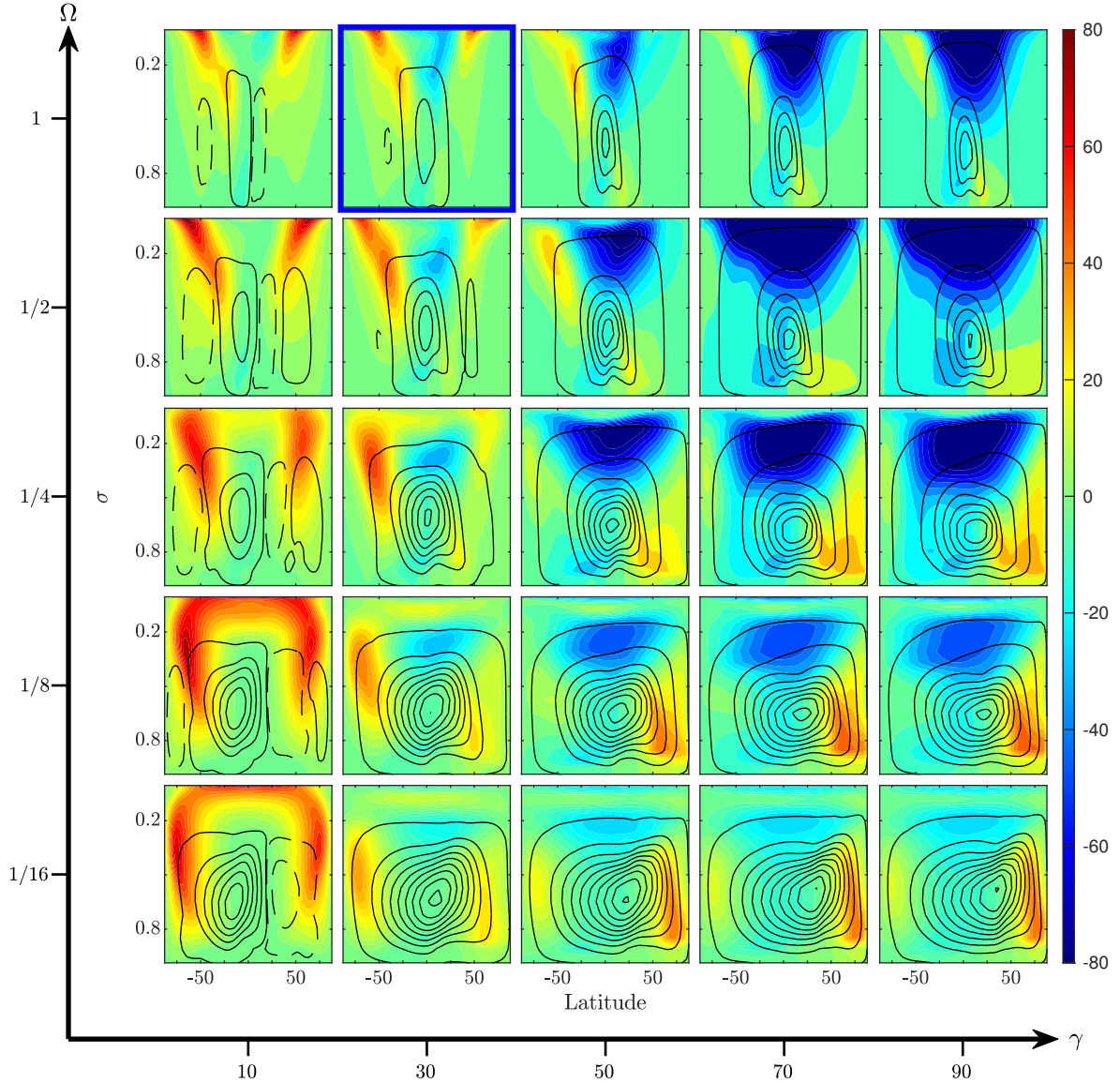


FIG. 2. Zonal mean zonal wind (shading, $\pm 80 \text{ m s}^{-1}$) and mean meridional circulation (contours, solid contours represent counter-clockwise circulation, contour intervals are $2.5 \times 10^{11} \text{ kg s}^{-1}$) for different values of the parameter space. The panel with blue frame represents Earth-like simulations ($\Omega = 1$, $\gamma = 30^\circ$).

the summer pole and their extension upward, towards the mid-troposphere (right column in Fig. 2).

Both the winter jet and the mid-troposphere summer jet are strongly related to the cross-equatorial winter Hadley cell. While the winter jet is located in close proximity to the descending branch of the cell, the summer jet is close to the ascending branch (Fig. 3e-f). Their vertical structure also differs: The winter jet is located near the top of the troposphere, extending downwards, whereas the summer jet is located in the low and mid-troposphere (Fig. 2). Ad-

ditionally, both the winter and summer jets' strength vary non-monotonically in the parameter space (Fig. 3b-c).

To understand the flow's dependence on obliquity and the rotation rate, we examine the parameter space starting with the low obliquity and a fast rotation rate case ($\gamma = 10^\circ$ and $\Omega = 1$, top left corner in Fig. 2). The flow in this case is close to hemispheric symmetry, with a slightly stronger jet in the winter hemisphere. As the rotation rate decreases (going down the first column in Fig. 2), the flow remains approximately hemispherically symmetric, with

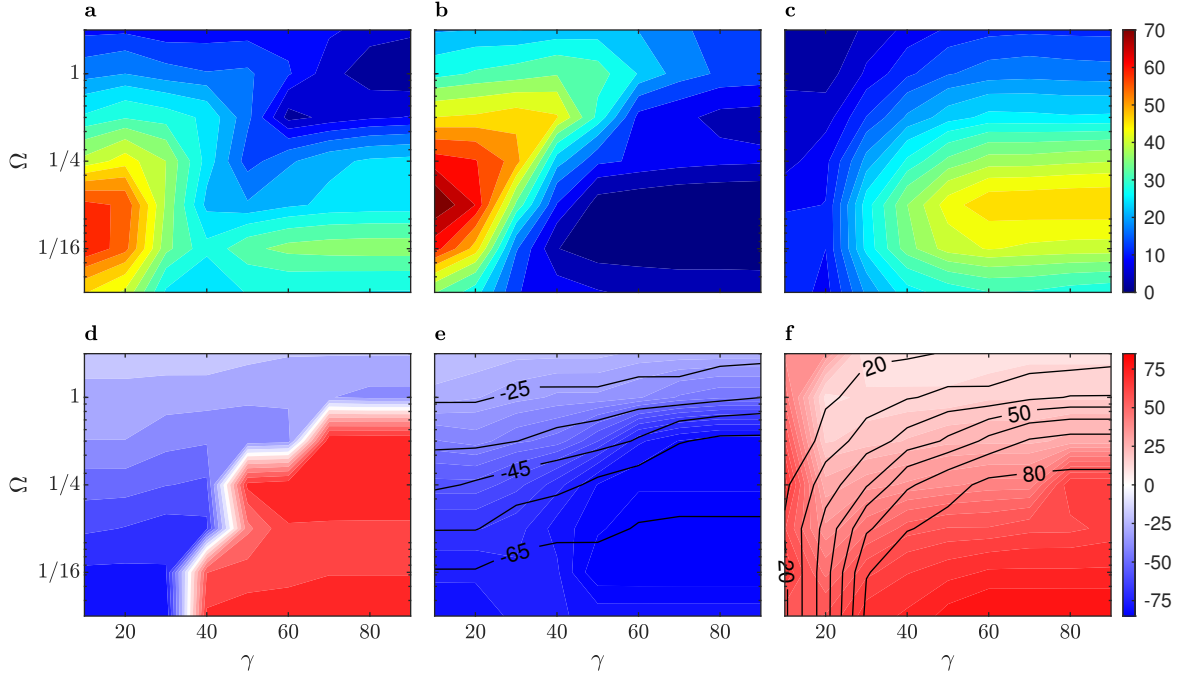


FIG. 3. Global maximum wind (m s^{-1} , a) and its latitude (d) in the mid-troposphere ($\sigma = 0.5$). High-level ($\sigma = 0.25$) winter (southern) hemisphere maximum wind (m s^{-1} , b), its latitude (e, shading) and latitude of the Hadley cell descending branch (e, contours). Low-level ($\sigma = 0.85$) summer (northern) hemisphere maximum wind (m s^{-1} , c), its latitude (f, shading) and latitude of the Hadley cell ascending branch (f, contours).

the circulation becoming wider and stronger and the jet strength varying non-monotonically with decreasing rotation rate (Figs. 2 and 3a-b). This non-monotonic behavior was also noticed in previous studies (e.g., Kaspi and Showman 2015; Wang et al. 2018). Wang et al. (2018) argued that the non-monotonic response of the jet's strength to the reduction in the rotation rate is a result of a competing effect that occurs as the rotation rate decreases. On the one hand, as the rotation rate decreases, the jet's location shifts to a more poleward latitude, and thus obtains more angular momentum. On the other hand, the reduction in the rotation rate reduces the global angular momentum, thereby diminishing the effect of the jet's shifting to a more poleward latitude, especially for slow rotation rates ($\Omega \leq 1/8$), when the jet reaches high latitudes and no longer shifts poleward with reducing rotation rate. In cases with low obliquities (leftmost column in Fig. 2), in addition to the winter jet, there is a jet in the summer hemisphere. Unlike the low-level summer jet discussed earlier, this summer hemisphere jet has similar characteristics to the winter jet, as it is a subtropical jet located at the descending edge of the summer Hadley circulation, unlike the low-level jet that is located at the ascending edge of the cell (Figs. 2 and 3f). In addition to the widening of the circulation with decreasing rotation rate, equatorial super-rotation also starts to develop at slow rotation rates (Fig. 2,

a feature seen on Venus and Titan, e.g., Lebonnois et al. 2014; Sánchez-Lavega et al. 2017).

Increasing the obliquity for fast rotation rates (top row in Fig. 2, from left to right) increases the hemispherical asymmetry and pressure gradients, and the circulation consists of one dominant cross-equatorial winter cell. For fast rotation rates and high obliquities, the cross-equatorial Hadley circulation does not span the entire planet, due to angular momentum constraints (Faulk et al. 2017; Guendelman and Kaspi 2018; Singh 2019; Hill et al. 2019). These constraints are relaxed when the rotation rate is decreased, and, as a result, for intermediate-to-high obliquities, the circulation widens and the winter jet is diverted poleward as the rotation rate decreases (Fig. 3c-e).

As during winter, the temperature differences increase with increasing obliquity (e.g., Guendelman and Kaspi 2019; Lobo and Bordonni 2020) one would expect a stronger jet with increasing obliquity. However, similar to the variations with rotation rate for low obliquity case, the jet strength varies non-monotonically with the obliquity, reaching a maximum strength at low or moderate obliquities. A similar non-monotonic behaviour with obliquity has been shown in Mars paleoclimate simulations (e.g., Toigo et al. 2020). Angular momentum considerations again provide insights into the variations of jet strength in the parameter space. Consider the angular momentum

conserving wind of a parcel at the jet latitude (ϕ_j), that starts at rest at the ascending branch of the Hadley circulation (ϕ_1)

$$U_M = \Omega a \frac{\cos^2 \phi_1 - \cos^2 \phi_j}{\cos \phi_j}, \quad (1)$$

where a is the planetary radius (e.g., Vallis 2017). U_M depends on whether $|\phi_j|$ is larger or smaller than $|\phi_1|$ and on how poleward ϕ_j is (Eq. 1). For low obliquities, ϕ_1 is close to the equator and $|\phi_j| - |\phi_1|$ is significantly greater than zero, so the jet is strong. As the obliquity increases, there are again competing effects, where on the one hand, $|\phi_j| - |\phi_1|$ becomes smaller and thus acting to reduce the jet strength, and on the other hand, ϕ_j moves more poleward, acting to increase the jet strength. For high obliquity, $|\phi_j| \sim |\phi_1|$ (Fig. 3e,f) and the first effect dominates so the jet weakens. It is important to note that this is a very idealized representation, and in reality there are other processes (e.g., eddies) that play a role in determining the jet's strength, however, it gives insight into the simulation results.

The latitude with the maximum winds in the mid-troposphere ($\sigma = 0.5$) changes sharply between hemispheres at high obliquities ($\gamma \geq 40$) and slow rotation rates ($\Omega < 0.5$, Fig. 3d), marking the transition from a winter jet to a summer jet regime. This regime transition occurs as the low-level winds in the summer hemisphere shift poleward and extend deeper in the atmosphere with increasing obliquity and decreasing rotation rate (Fig. 2). At moderate-to-high obliquities and fast rotation rates, the summer jet appears as low-level westerly winds close to the ascending edge of the Hadley cell (Fig. 2,3f). These low-level winds resemble the low-level westerlies that are observed on both Earth (monsoon regions, e.g., Joseph and Raman 1966), and Mars (e.g., Hinson et al. 1999) and are seen in Mars's paleoclimate studies (Haberle et al. 2003; Toigo et al. 2020). In these cases, during the solstice season, when the warmest latitude is off the equator, in higher latitudes, the circulation consists of a cross-equatorial circulation with air ascending in the summer hemisphere. The low-level returning poleward flow in the summer hemisphere is balanced by the boundary layer drag that results in the lower level westerlies (e.g., Schneider and Bordonni 2008).

Exploring the parameter space, as the rotation rate decreases, the low-level winds strengthen (non-monotonically with Ω) and extend deeper into the mid-troposphere (Figs. 2-3). The boundary layer process used to explain the low-level winds in the fast rotating planets cannot explain the deepening of the jet at slow rotation rates. In the following section, we study the momentum balance to understand the jet changes across the parameter space.

4. Balance and mechanism

a. Meridional momentum balance

In order to better understand the summer jet, it is insightful to consider the momentum balance. Assuming a steady-state, the zonal mean meridional momentum equation, in pressure coordinates, is given by (Vallis 2017)

$$f\bar{u} + \frac{\overline{u^2 \tan \phi}}{a} + \omega \frac{\overline{\partial v}}{\partial p} = -\frac{1}{a} \frac{\partial \bar{\Phi}}{\partial \phi} + D_v, \quad (2)$$

where v is the meridional velocity, u the zonal velocity, ω the vertical (pressure) velocity, $f = 2\Omega \sin \phi$ the Coriolis parameter, ϕ latitude, Φ the geopotential and D_v the zonal mean meridional boundary layer drag. We neglect here meridional advection terms (terms of the form $\partial_\phi(v^2)$) and other forms of dissipation that are small. The over-bar notation denotes a zonal mean.

The planetary rotation rate influences the transition of the leading order meridional momentum balance. For fast-rotating planets (Fig. 4A,C), such as Earth and Mars, the leading order balance is geostrophic (Vallis 2017)

$$f\bar{u} \approx -\frac{1}{a} \frac{\partial \bar{\Phi}}{\partial \phi}. \quad (3)$$

At moderate rotation rates, the leading order balance shifts to a thermal gradient balance (Sanchez-Lavega 2010)

$$f\bar{u} + \frac{\overline{u^2 \tan \phi}}{a} \approx -\frac{1}{a} \frac{\partial \bar{\Phi}}{\partial \phi}. \quad (4)$$

For slow-rotating planets with low-to-moderate obliquity (Fig. 4B), such as Venus and Titan, the leading order balance becomes cyclostrophic (Read and Lebonnois 2018)

$$\frac{\overline{u^2 \tan \phi}}{a} \approx -\frac{1}{a} \frac{\partial \bar{\Phi}}{\partial \phi}. \quad (5)$$

The cyclostrophic balance does not hold for the summer jet, at slow rotation rates and high obliquities (Fig. 4D). Unlike all the above cases, where the vertical advection and friction are negligible, for slow rotation rates and high obliquities the leading order balance is more complex and all the terms in Eq. 2 contribute to the balance (Fig. 4D).

We can divide the summer jet balance into a boundary layer balance

$$0 \approx -\frac{1}{a} \frac{\partial \bar{\Phi}}{\partial \phi} + D_v, \quad (6)$$

and a free atmosphere balance

$$f\bar{u} + \frac{\overline{u^2 \tan \phi}}{a} + \omega \frac{\overline{\partial v}}{\partial p} \approx -\frac{1}{a} \frac{\partial \bar{\Phi}}{\partial \phi}. \quad (7)$$

The meridional momentum balance in both the boundary and free atmosphere differ between the summer jet

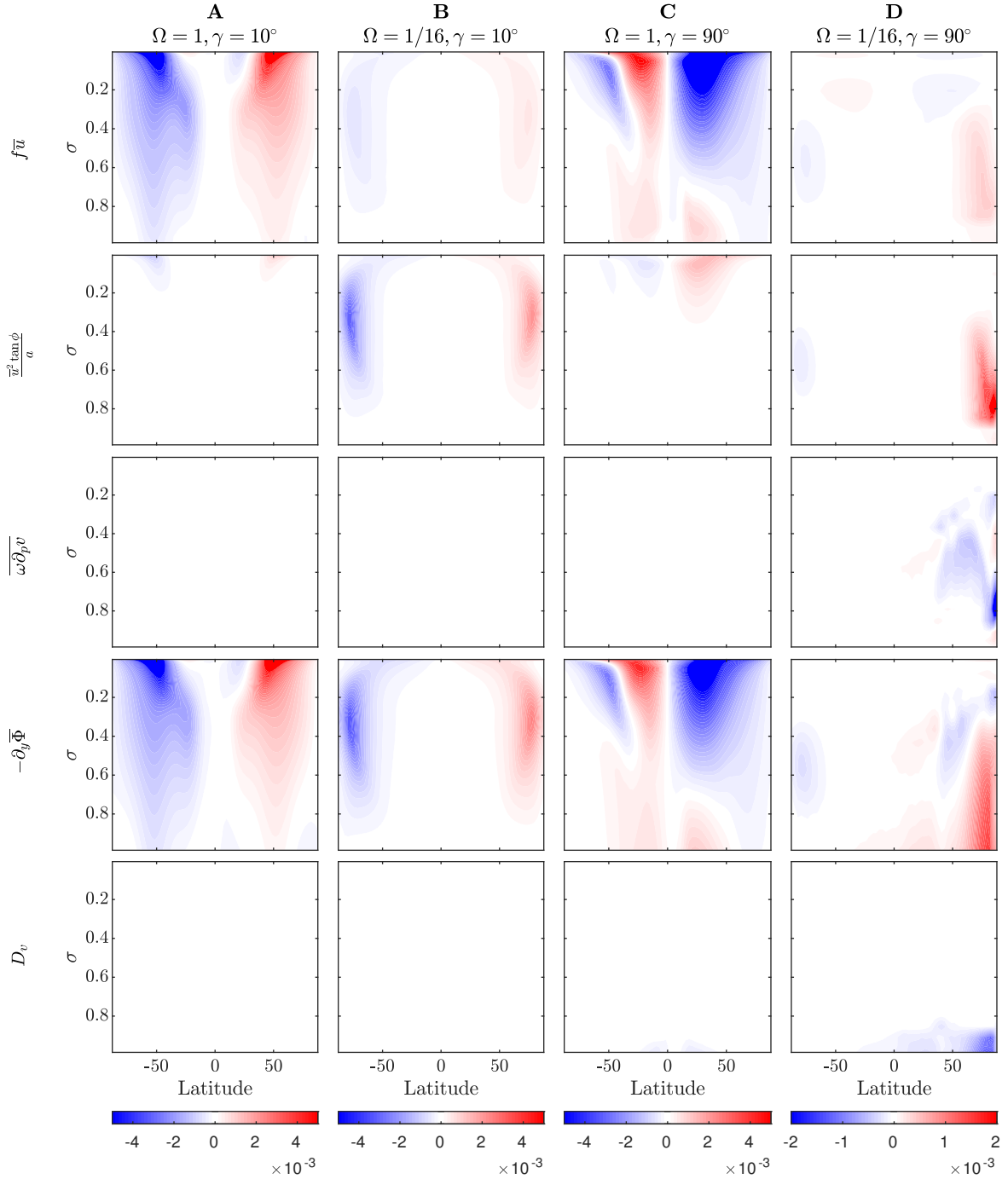


FIG. 4. The different terms in the meridional momentum equation (rows, as written in Eq. 2, ms^{-2}) for four different cases (columns).

and other regimes. First, for slow rotation rates and high obliquities, the boundary layer balance is comparable to the free atmosphere, but in all other cases, it is much weaker (Fig. 4). A second difference is in the boundary layer balance. In previous studies that explored either

rapid rotation rate and moderate-high obliquity or moderate obliquity with slower rotation rates, the boundary layer balance was found to consist of a three- or four-way balance between the Coriolis acceleration, geopotential meridional gradient, boundary layer drag, and, for slow

rotation, the cyclostrophic term (e.g., Schneider and Bordoni 2008; Faulk et al. 2017; Lobo and Bordoni 2020, see also Fig. 4C). However, for slow rotation rates and high obliquities, the boundary layer leading order balance is mainly between the drag and geopotential meridional gradient (Fig. 4D). Third, for slow rotation rates and high obliquities, the vertical advection term is essential, as it transports momentum between the boundary layer and the free atmosphere.

While the geostrophic and cyclostrophic balances relate the zonal wind to geopotential gradients, making the balance more intuitive, in the summer jet, the balance involves terms like vertical advection of meridional momentum and drag on meridional momentum, which make the balance more complex and less straightforward to understand. To obtain a more complete picture of the summer jet's maintenance, it is important to understand the coupling between the meridional and zonal momentum balances. To do so, it is essential to understand the zonal mean zonal momentum balance.

b. Zonal momentum balance

The steady state, zonal mean zonal momentum equation is given by (Vallis 2017)

$$-(f + \bar{\zeta})\bar{v} + \overline{\omega \frac{\partial u}{\partial p}} = \overline{\zeta'v'} + D_u. \quad (8)$$

Eq. 8 is written using the vorticity notation, where $\bar{\zeta} = -(a \cos \phi)^{-1} \partial_\phi (\bar{u} \cos \phi)$, and D_u is the zonal mean zonal boundary layer drag.

This balance gives insight into the processes responsible for the jet acceleration, as the terms of this balance represent zonal winds' tendencies. In general, when discussing the zonal mean circulation, one can distinguish between a flow mediated by eddies and one that is more axisymmetric. When the flow is axisymmetric, an air parcel at the top of the circulation moving poleward conserves its angular momentum, but, this is not the case when eddies influence the circulation (e.g., Schneider and Bordoni 2008).

For fast rotation rates and low obliquities, the leading order balance at the top of the Hadley cell is $-(f + \bar{\zeta})\bar{v} \approx \overline{\zeta'v'}$, meaning that the circulation is influenced by eddies ($\overline{\zeta'v'}$, Fig. 5A). Therefore, the angular momentum conserving wind is not a good approximation for the jet at the descending edge of the Hadley circulation (e.g., Schneider 2006). When the rotation rate is decreased while preserving a low-to-moderate obliquity or the obliquity is increased in a fast rotation rate scenario, there is a transition from an eddy-influenced circulation to a more axisymmetric one. For example, in slow rotation rates and a low obliquities, the leading order balance at the top of the circulation, except at the ascending and descending branches, is $-(f + \bar{\zeta})\bar{v} \approx 0$ (Fig. 5B). This balance means

that the circulation in these regions is close to axisymmetric, and the jet is a result of angular momentum conservation at the top of the circulation (Held and Hou 1980; Lindzen and Hou 1988; Schneider and Bordoni 2008). A similar balance appears for fast rotation rates with high obliquities, with the distinction in the boundary, where the boundary layer drag is stronger (Fig. 5C). This transition from an eddy-mediated circulation to a more axisymmetric one was previously shown for both increasing seasonality (e.g., Schneider and Bordoni 2008; Bordoni and Schneider 2008; Faulk et al. 2017; Guendelman and Kaspi 2018, 2019; Lobo and Bordoni 2020; Guendelman and Kaspi 2020) and for decreasing rotation rate (e.g., Del Genio and Suozzo 1987; Kaspi and Showman 2015; Guendelman and Kaspi 2018, 2019; Colyer and Vallis 2019; Komacek and Abbot 2019).

In cases of slow rotation rates and high obliquities, away from the region of the summer jet, the balance is $-(f + \bar{\zeta})\bar{v} \approx 0$, meaning that the flow is close to axisymmetric. However, this is not the case in the summer jet, where the balance can be again divided into a boundary layer and a free atmosphere balance. Additionally, the balance differs between the regions poleward and equatorward of the jet core. Equatorward of the jet core, in the boundary layer the balance is $-(f + \bar{\zeta})\bar{v} \approx D_u$, and at the top of the atmosphere, $(f + \bar{\zeta})\bar{v} \approx \overline{\omega \partial_p u}$. Poleward of the jet core, the balance is somewhat different. In the boundary layer, the balance is $-(f + \bar{\zeta})\bar{v} + \overline{\omega \partial_p u} \approx D_u$, meaning that the returning flow of the Hadley circulation results in surface westerlies through an Ekman balance. Above the boundary layer, the balance is $(f + \bar{\zeta})\bar{v} \approx \overline{\omega \partial_p u}$, meaning that the mean meridional circulation is balanced by vertical advection of zonal momentum. Alternatively, momentum is transported vertically from the boundary layer to the free atmosphere (Fig. 5D).

For the summer jet to extend deeper into the mid-troposphere, the combination of high obliquity and a slow rotation rate is essential. Although past studies have connected between the latitude of maximum moist static energy (and temperature) with the ascending motion at the edge of the Hadley cell (e.g., Privé and Plumb 2007), recent studies have shown that this is not necessarily the case (e.g., Faulk et al. 2017; Guendelman and Kaspi 2018, 2019; Lobo and Bordoni 2020). This separation is a result of angular momentum constraints, and as the rotation rate decreases, the ascending motion shifts more towards the summer pole, i.e., closer to the latitude of maximum temperature (e.g., Guendelman and Kaspi 2018).

As the obliquity is increased, the latitude of maximum temperature moves poleward (e.g., Guendelman and Kaspi 2019), resulting in strong cross-equatorial circulation. Low-level westerlies balance the returning flow in the boundary layer. For fast rotation rates, the circulation width is confined, resulting in a separation between

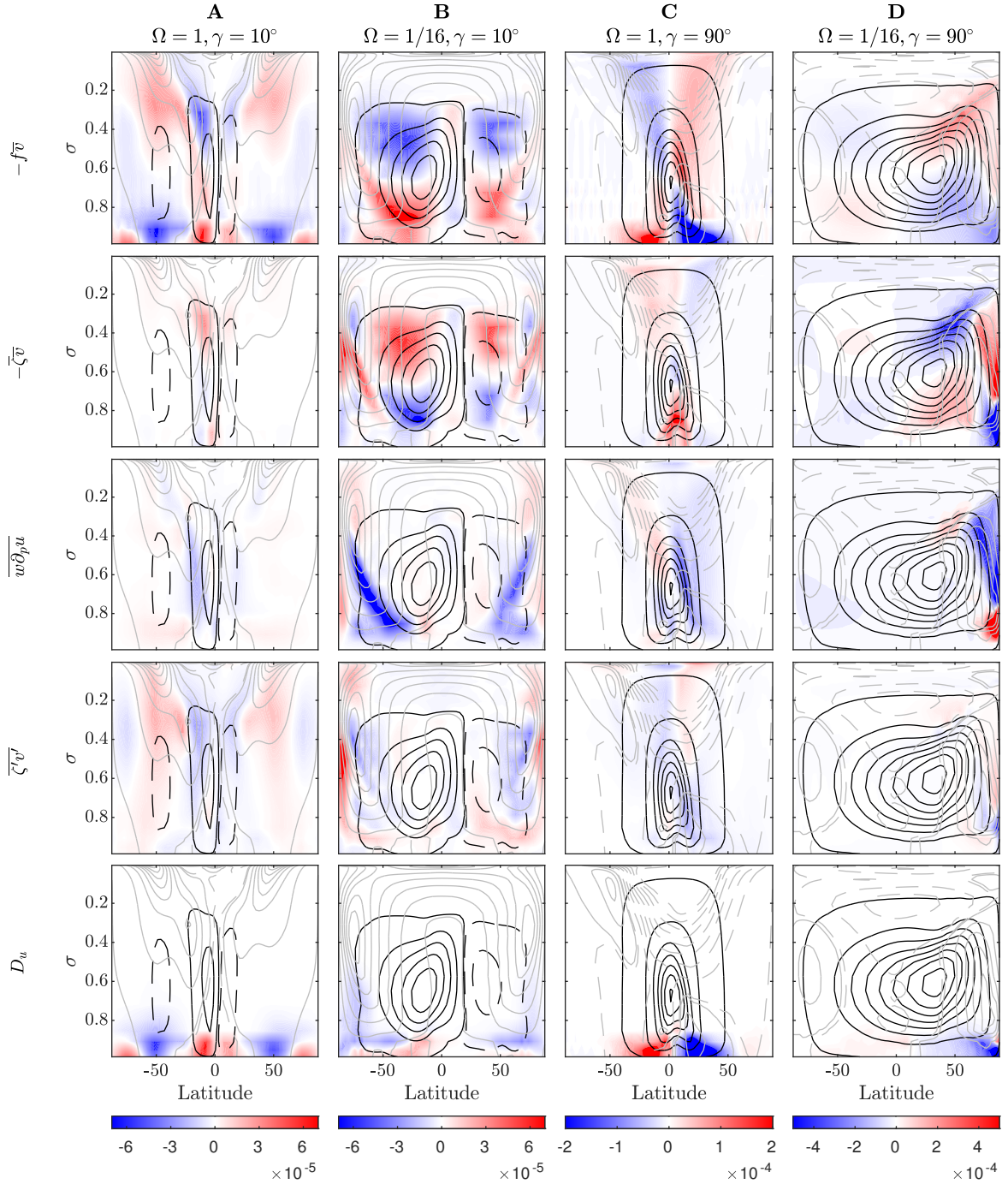


FIG. 5. The different terms in the zonal momentum equation (rows, as written in Eq. 8, ms^{-2}) for four different cases (columns). Black contours are for the mean meridional streamfunction, solid contours represent counterclockwise circulation. Contour intervals for fast rotation rate (A,C) as $1.5 \times 10^{11} \text{ kg s}^{-1}$ and for slow rotation rate (B,D) $3 \times 10^{11} \text{ kg s}^{-1}$. Gray contours are for the zonal mean zonal wind, contour interval is 10 ms^{-1} .

the warmest latitude and the Hadley cell ascending branch (Guendelman and Kaspi 2018; Hill et al. 2019; Singh 2019). Decreasing the rotation rate, the ascending branch of the Hadley cell aligns with the warmest latitude (e.g.,

Guendelman and Kaspi 2018, 2019). The alignment between the ascending motion of the Hadley circulation and the warmest latitude, together with a general strengthening of the circulation with decreasing rotation rate and in-

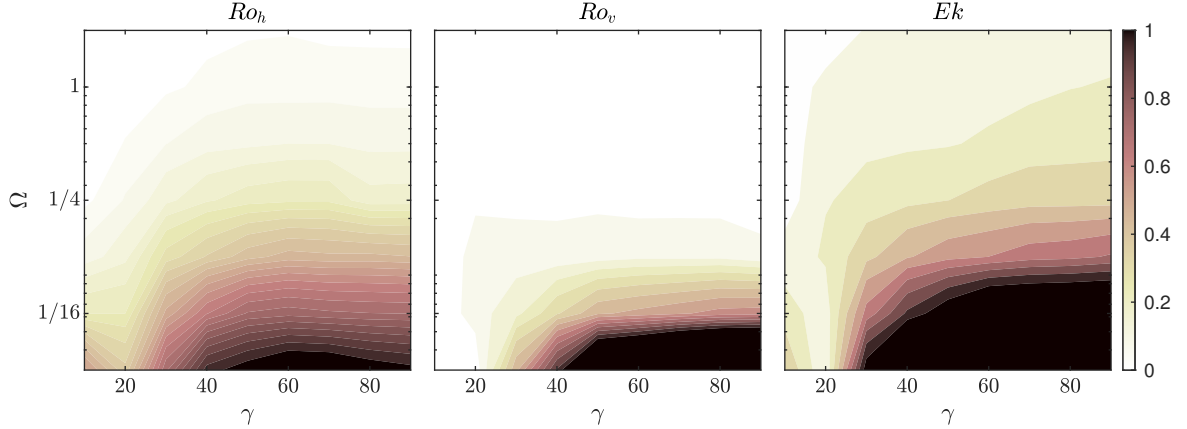


FIG. 6. Values of non-dimensional numbers across the parameter space in the boundary layer in the summer jet core latitude. (a) Ro_h , horizontal Rossby number, (b) Ro_v , vertical Rossby number, and (c) Ek Ekman number.

creasing obliquity (Guendelman and Kaspi 2019), intensifies the ascending motion of the ascending branch of the cross-equatorial cell. This intensification is strongest close to the pole, the warmest latitude. As a result, the vertical motion at the polar latitude advects momentum upwards (more significantly poleward of the jet core), resulting in the low-level jet expanding into the mid-troposphere.

The separation between the Hadley cell ascending branch and the warmest latitude also explains the summer jet's split in latitude for intermediate rotation rates. At intermediate rotation rates, although the ascending branch is in the midlatitudes, there is a weak poleward flow poleward of the Hadley cell ascending branch, which also results in a westerly flow through an Ekman balance. Additionally, in the warmest latitude, there is an ascending motion of air due to local convection (Lobo and Bordoni 2020). The combination of a weak poleward flow with an ascending motion in polar latitudes in these cases results in a secondary peak of zonal winds close to the pole at intermediate rotation rates, with the two peaks merging at slow rotation rates (two rightmost columns in Fig. 2).

c. Non-dimensional numbers

An alternative way to understand the transition between the winter- and summer-jet regimes, is by considering non-dimensional numbers. Non-dimensionalizing Eq. 2 gives

$$\hat{u} + Ro_h \hat{u}^2 \tan \phi + Ro_v \hat{\omega} \frac{\partial \hat{\phi}}{\partial \hat{p}} = -\frac{\partial \hat{\phi}}{\partial \phi} + Ek, \quad (9)$$

where we scale the horizontal wind as $u, v \sim U$, the vertical wind as $\omega \sim W$, and where

$$Ro_h = U / f a \quad (10)$$

is the horizontal Rossby number,

$$Ro_v = W / f \Delta p \quad (11)$$

is the vertical Rossby number,

$$Ek = D_v / f u \quad (12)$$

is the Ekman number, and

$$\hat{\phi} = \frac{\Phi}{f U a}. \quad (13)$$

is the non-dimensional geopotential. The planetary radius, a , is taken as the typical horizontal length scale, and Δp is taken as the typical height of the circulation (in pressure coordinates). For the winter jet, where the jet is in the free atmosphere, surface drag and vertical advection are negligible ($Ro_v, Ek \ll 1$) and the transition from a geostrophic balance (Eq. 3) to a thermal gradient balance (Eq. 4) and a cyclostrophic balance (Eq. 5) is represented by the increase in Ro_h (Fig. 6).

The variations of the non-dimensional numbers at the boundary layer indicate that the transition to the summer jet regime occurs when all three non-dimensional parameters are of order one or larger (Fig. 6). Although all the non-dimensional parameters increase with decreasing rotation rate and increasing obliquity, the increase in Ro_v at low rotation rates and moderate-high obliquities is the most abrupt one and represents the transition from a low-level summer jet to a mid-tropospheric one, due to the intensification of the vertical advection that occurs as the Hadley cell's ascending branch and warmest latitude align in this part of the parameter space.

In addition, the non-dimensional numbers vary spatially and can represent the different balance between the boundary layer and free atmosphere. For example, in the summer jet regime, in the boundary layer, $Ro_h, Ro_v > 1$, and

the balance is given by Eq. 6 (Fig. 6), whereas in the free troposphere, $Ro_h, Ro_v \sim 1$ and $Ek \ll 1$ and the balance is given by Eq. 7.

5. Conclusions

The solar system's terrestrial planetary bodies with a seasonal cycle have a dominant jet in the winter hemisphere, close to the Hadley cell descending branch. In the monsoon region on Earth, where the ascending motion of the circulation is most dominant (Raiter et al. 2020), and on Mars, this winter jet is accompanied by weaker low-level westerlies close to the Hadley cell ascending branch in the summer hemisphere. To better understand this phenomenon and general jet dynamics, we examine the jet characteristics' dependence on the rotation rate and obliquity, which allows a better separation of scales needed to identify the physical mechanisms controlling the dynamics. Consistent with observations of the solar system, at a fast rotation rate and strong seasonality, the dominant jet is in the winter hemisphere. However, we show that for planets with high obliquity and a slow rotation rate, the low-level summer westerlies extend into the middle troposphere, and the dominant jet is in the summer hemisphere and occurs within the ascending, rather than descending, branch of the Hadley cell (Figs. 2,3).

Figure 7 summarises schematically the different circulation regimes that exist in this parameter space. At fast rotation rates and low obliquities, the mean meridional circulation is strongly influenced by eddies and consists of two cells (Fig. 5A), with the cross-equatorial winter cell becoming wider and stronger. At the edge of each cell (winter and summer), there is a westerly jet that is in a geostrophic balance (Fig. 4A), with the winter hemisphere jet being stronger. For slower rotation rates and low obliquities, the circulation still consists of two cells that are wider and more axisymmetric (Fig. 5B). The jets at the descending branch of each cell are now in a cyclostrophic balance (Fig. 4B), with the winter jet the more dominant one.

Once the obliquity is high enough, the circulation consists of one strong cross-equatorial cell that becomes more axisymmetric. The returning boundary layer poleward flow in the summer hemisphere is balanced by a low-level westerly flow through an Ekman balance (Fig. 5C). For fast rotation rates, the circulation's width is limited and does not align with the warmest latitude due to angular momentum constraints (Guendelman and Kaspi 2018; Hill et al. 2019; Singh 2019). In this case, the summer hemisphere westerlies remain close to the boundary layer, and the winter jet is the dominant jet. As the rotation rate is decreased, the Hadley cell ascending branch and the warmest latitude align. The alignment, together with the strengthening of the circulation with increasing obliquity and decreasing rotation rate (Guendelman and Kaspi 2019), re-

sults in a more efficient vertical advection of momentum, extending the winds further upwards in the atmosphere, which results in a mid-troposphere summer jet, which also becomes the more dominant jet (Fig. 5D).

The importance of the vertical advection in the momentum balance can be represented by the increase in the vertical Rossby number, Ro_v . For moderate-fast rotation rates or moderate-low obliquities, $Ro_v \ll 1$ and the vertical advection is negligible, but for slow rotation rates and high obliquities, $Ro_v \sim 1$ and the vertical advection cannot be neglected (shading in Fig. 7).

The results presented here not only increase our understanding of the wind patterns on terrestrial planets in the solar system, but also provide insights into possible flows on exoplanets. While a strong winter westerly jet is a common feature of terrestrial planets in the solar system, the simulations presented here indicate this may not always be the case for all exoplanets. In particular, if a terrestrial exoplanet has a slow rotation rate like Venus and Titan but a higher obliquity, it may have summer, rather than winter, westerly jets. The impact of a summer jet needs to be further examined. For example, one possible impact of a summer jet is an effect on the seasonality of trace constituents, transport to polar regions. On Earth, Mars, and Titan, the winter jets act to reduce transport between low and high latitudes, and unique chemical-microphysical processes occur in these planets' winter polar atmospheres. The situation may be different for planets with dominant summer jets.

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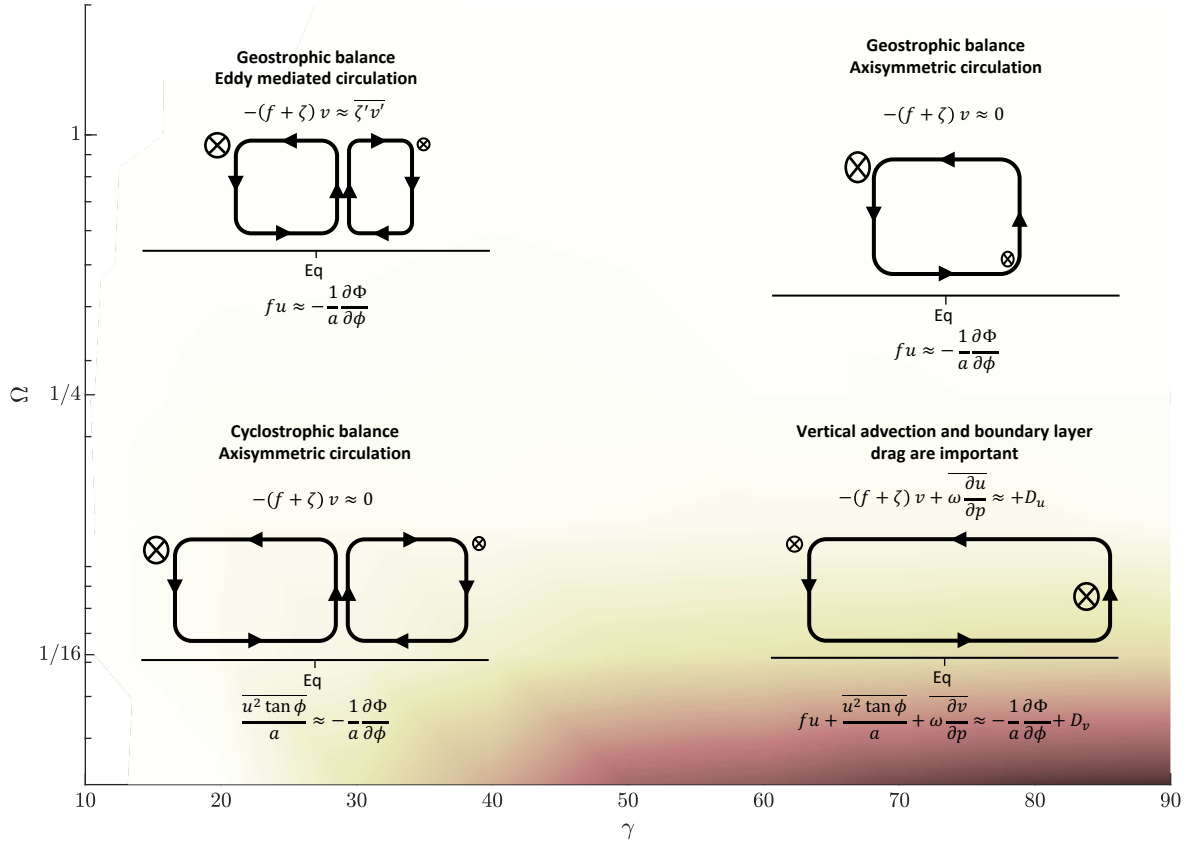


FIG. 7. Schematic picture summarizing the different circulation regimes presented in this study. The background shading shows the value of $Ro_v = \omega / f \Delta p$, with the scale between 0 (white) and 2 (dark).

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