

Regime Change Behavior During Asian Monsoon Onset

Ruth Geen*, F. H. Lambert and G. K. Vallis

College of Engineering, Mathematics and Physical Sciences, University of Exeter, Exeter, United Kingdom

*Corresponding author address: Ruth Geen, College of Engineering, Mathematics and Physical Sciences, University of Exeter, North Park Road, Exeter, Devon EX4 4QF, United Kingdom.

E-mail: rg419@exeter.ac.uk

ABSTRACT

8 As the ITCZ moves off the equator the Hadley circulation transitions
9 from an equinoctial regime with two near symmetric, significantly eddy-
10 driven cells, to a monsoon-like regime with a strong, thermally direct cross-
11 equatorial cell, intense low-latitude precipitation, and a weak summer hemi-
12 sphere cell. Dynamical feedbacks appear to accelerate the transition. This
13 study investigates the relevance of this behavior to monsoon onset by using
14 a primitive-equation model in two aquaplanet configurations and a configura-
15 tion with a realistic configuration of Earth's continents and topography.

16 A change in the relationship between ITCZ latitude and overturning strength
17 is identified once the ITCZ moves poleward of about $\sim 10^\circ$. In the monsoon
18 regime this relationship is similar in all simulations, suggesting that similar
19 dynamics are occurring. Monsoon onset is associated with off-equatorial as-
20 cent, in regions of non-negligible planetary vorticity, and this is found to gen-
21 erate a vortex stretching tendency that reduces upper level absolute vorticity.

22 This causes a transition to a cross-equatorial thermally direct regime, inten-
23 sifying the overturning circulation. Analysis of the zonal momentum budget
24 suggests a stationary wave, driven by topography and land-sea contrast, can
25 trigger the transition in the more realistic model configuration, with the wave
26 extending the ascent region of the Southern Hemisphere Hadley cell north-
27 ward, with a thermally direct overturning then developing to the south. These
28 two elements of the circulation resemble the East and South Asian monsoons.

29 1. Introduction

30 The onset of the Asian summer monsoon is characterized by a fast transition from a dry season
31 with easterly flow at lower levels and westerly flow at upper levels, to a wet season with lower
32 level westerlies and upper level easterlies. However, the dynamics involved are still not well
33 understood. In this paper, we try to unify perspectives from previous studies of monsoon dynamics
34 in aquaplanets, and investigate the relevance of these to the real world.

35 The monsoon is forced by the seasonal cycle of solar heating, resulting in migration of the
36 intertropical convergence zone (ITCZ) (Chao and Chen 2001), and warming of land relative to
37 ocean (Li and Yanai 1996). Continents, orography, and sea surface temperature (SST) variations
38 result in regional circulation features and different onset times across different longitudes (e.g. Wu
39 and Wang 2001; Zhang et al. 2002). However, the transition in the circulation during monsoon
40 onset occurs faster than would be expected from the northward progression of the peak insolation
41 (Yin 1949). This suggests that dynamical feedbacks may play an important role.

42 An emerging view in the literature is that monsoon onset marks a sharp transition between two
43 dynamical regimes (e.g., Chao 2000). Considering the circulation on an aquaplanet, where the
44 lower boundary conditions are zonally symmetric, the zonal mean zonal momentum budget may
45 be approximated as:

$$\frac{\partial[u]}{\partial t} = \left(f - \frac{\partial[u]}{\partial y} \right) [v] - \frac{\partial[u^*v^*]}{\partial y}, \quad (1)$$

46 where f is the Coriolis parameter, and u and v are the zonal and meridional wind speeds respec-
47 tively. Square brackets indicate the zonal mean, stars deviations from this. In the deep tropics eddy
48 fluxes are generally assumed to be weak, so that the final term on the right-hand side of Eq. (1)
49 is negligible. In this situation, two steady-state solutions exist: either the zonal mean meridional
50 wind speed, or the zonal mean absolute vorticity, must be equal to zero. Eddies may then allow

51 some deviation from these solutions. In an axisymmetric model with off-equatorial heating this
52 can result in threshold behavior (Plumb and Hou 1992; Privé and Plumb 2007a,b). As the heating
53 becomes stronger there is a change from a very weak meridional circulation (with $[v]$ very small)
54 to a stronger, thermally direct overturning cell (with absolute vorticity near the upper boundary
55 now very small). This latter regime also corresponds to an angular momentum conserving Hadley
56 Cell, as in Schneider (1977), Held and Hou (1980) and Lindzen and Hou (1988).

57 However, as has long-been recognized (e.g., Kim and Lee 2001) eddy effects are almost certainly
58 not negligible in the Hadley Cell, especially near its poleward edge where the dynamics transition
59 into a mid-latitude regime. Relatedly, more recent work suggests that monsoon onset relates to
60 a transition from a largely eddy driven to an axisymmetric, cross-equatorial, Hadley circulation.
61 Bordoni and Schneider (2008) identified abrupt circulation changes, similar to those observed over
62 monsoon onset, in idealized aquaplanet simulations, provided a sufficiently shallow mixed layer
63 depth is used. They proposed that the development of upper level easterlies in the summer hemi-
64 sphere restricts the propagation of eddy activity in this region. The result is a more axisymmetric
65 Hadley cell which, from Eq. (1), will also have low upper level absolute vorticity. Meanwhile, the
66 lower branch of the cell advects cooler, drier air from the winter to summer hemisphere, up the
67 temperature gradient. This pushes the maximum moist static energy, and the ITCZ, poleward. This
68 extends the region of upper level easterlies, preventing the penetration of eddy effects and giving
69 a positive feedback. In their moist model, this regime change was accompanied by the onset of
70 intense precipitation at low latitudes, which they connect to the monsoon. If a similar mechanism
71 can be taken to apply in the real world, then the role of land is essentially to provide a surface
72 with low thermal inertia, in which case temperature gradients associated with land-sea contrast
73 are not necessary to trigger monsoon onset, although they may still be important in localizing the
74 monsoon longitudinally, which is sometimes regarded as one of its defining features.

75 In reality the lower boundary condition on Earth is of course not symmetric, unlike the aqua-
76 planet studies discussed above. To investigate the role played by stationary waves in the seasonal
77 transition of the circulation, Shaw (2014) compared the circulation characteristics observed in
78 ERA-Interim re-analysis data with those induced when a prescribed wave is introduced into the
79 SSTs. She identified a transition to a planetary-scale wave dominated regime, which occurs once
80 the amplitude of the SST wave is increased past a threshold value of 6 K. This regime also ex-
81 hibits low upper level absolute vorticity locally, suggesting a thermally direct circulation. Zhai
82 and Boos (2015) used a dry primitive equation model to investigate this behavior systematically.
83 They compared the effect of steady, off-equatorial zonally symmetric and asymmetric forcings on
84 the meridional circulation. As the zonal mean forcing amplitude was increased, a sudden change
85 in the strength of the meridional circulation was identified and associated with low absolute vor-
86 ticity in the upper branch of the Hadley cell. Asymmetric forcings introduced additional zonal
87 overturning with a structure similar to that seen in Gill (1980). This was found to increase linearly
88 in strength with the forcing amplitude, with no abrupt regime change apparent.

89 Regime change behavior has also been noted in observations. Krishnamurti and Ramanathan
90 (1982) identified a sudden enhancement of the kinetic energy of the nondivergent flow during
91 monsoon onset in observational data from the Global Atmospheric Research Program (GARP)
92 summer monsoon experiment (MONEX). Analyzing the energy conversions during onset in nu-
93 merical experiments, they find these to be sensitive to the structure of the diabatic heating field.
94 The configuration of land, topography, and sea surface temperatures may therefore be important
95 in determining the real-world monsoon dynamics.

96 Here we attempt to reconcile and unify the various mechanisms suggested above, and to link the
97 processes discussed in these idealized studies to the dynamics of a more realistic set-up, with a
98 particular focus on the Asian monsoon systems. We compare results from three experiments per-

99 formed with different configurations of the same model: two aquaplanets with mixed layer depths
100 of 2m and 20m, and an experiment with Earth-like continents and topography. The model used
101 and experimental design are discussed in Section 2. In Section 3 we compare the behavior of the
102 three simulations over onset, and identify behavior consistent with a change to a thermally direct
103 regime. To understand the reduction in upper level absolute vorticity responsible for this change,
104 the vorticity budget is analysed in Section 4. In Section 5 the role of zonal asymmetry and of waves
105 is investigated, giving insight into the limitations of aquaplanet experiments in describing the real
106 world. Section 6 summarizes our results and discusses the implications for our understanding of
107 monsoon onset.

108 **2. Model and experiments**

109 The model used is called ‘MiMA’ (Model of an idealized Moist Atmosphere), which is based
110 around the GFDL idealized model of Frierson et al. (2006). The model physics includes param-
111 eterizations of large-scale condensation, moist convection, vertical diffusion of heat, momentum
112 and moisture, and radiative transfer. For all but the last of these, the parameterizations used are as
113 in Frierson et al. (2006) and Frierson (2007), with the amendments of O’Gorman and Schneider
114 (2008). Rather than using a gray atmosphere with prescribed optical depths, the RRTM radiation
115 scheme (Mlawer et al. 1997) has been coupled to the model. The radiative heating is recalculated
116 every 3600 s of model time, compared with the 720 s time-step used for the dynamics and other
117 physical processes. Clouds are not included in the parameterization of moist processes or of radi-
118 ation. Although this is a limitation in some respects, comparison of the model in its most realistic
119 configuration with reanalysis (e.g. Kållberg et al. 2005) indicates that the key features of Earth’s
120 climatology are successfully captured. Simulations are run at T42 resolution, corresponding to
121 $\sim 2.8^\circ$ at the equator, with 40 unevenly spaced sigma levels. A slab ocean of prescribed mixed

layer depth allows a closed atmospheric energy budget to be achieved. The temperature of the slab responds only to the incident fluxes at the sea surface, with no dynamic heat transport. A sponge layer is applied at the model top to inhibit unstable gravity wave-like behavior.

Land is incorporated into the model by varying mixed layer depth, albedo, roughness length, and moisture availability over prescribed areas. The latter is controlled by introducing a scaling parameter α into the equation for surface evaporation, E :

$$E = \rho_a C |\mathbf{v}_a| (q_a - \alpha q_s) \quad (2)$$

where ρ_a , $|\mathbf{v}_a|$ and q_a are the density, horizontal wind speed and specific humidity at the lowest model level, C is the drag coefficient, and q_s is the saturation specific humidity at the surface temperature (cf. Eq. 11, Frierson et al. 2006). α acts as a scaling parameter to reduce the effective saturation specific humidity of the surface.

The simulations performed comprise two aquaplanet configurations, and a configuration intended to resemble Earth. The aquaplanets are run with uniform mixed layer depths of 2 m and 20 m, and will be referred to as *ap2* and *ap20* respectively. The Earth-like configuration includes land and topography based on the ERA-Interim masks (Fig. 1), and will be referred to as *full*. Topography is regularized following Lindberg and Broccoli (1996) to reduce the Gibbs ripples that arise from the truncated spherical harmonic expansion. The ocean is given a mixed layer depth of 20 m, while land has a depth of 2 m. For all experiments ocean is given an albedo of 0.25 and land of 0.325, with the high albedo values accounting for the lack of reflection by clouds in the model. Both ocean and land use a roughness length of 2×10^{-4} m. A scale factor, α , of 0.7 is introduced to reduce evaporation over land, as in Eq. (2). The parameter values were chosen to provide a climate comparable to reanalysis. Sensitivity experiments in which the parameters were varied give similar climates and behavior to the experiments presented here.

144 Prescribed, seasonally varying ocean heat-fluxes (Q-fluxes) are additionally included in the *full*
145 configuration. These were derived from the atmosphere model intercomparison project (AMIP)
146 SSTs following the methodology outlined in Russell et al. (1985). The resulting fluxes have a
147 comparable structure to the net surface heat flux seen in reanalysis (e.g. compare Fig. 1 with p.
148 12, Kållberg et al. 2005). The addition of Q-fluxes helps to produce a climatology resembling that
149 of Earth.

150 **3. Regime characteristics**

151 Monsoon onset is associated with sudden arrival of intense rainfall. Fig. 2 shows the evolution
152 of the modelled precipitation as a function of latitude and time for each of our simulations. A
153 zonal average is shown for the aquaplanets. For the *full* simulation, an average between 60°
154 and 150°E is used, which we take to be representative of the South and East Asian monsoon
155 regions. In this simulation, the peak precipitation shifts into the Northern Hemisphere in May
156 and strengthens, before weakening and returning southward in September. The simulated Asian
157 monsoon is consistent with that in observations from the Global Precipitation Climatology Project
158 (e.g. Fig. 1, Bordoni and Schneider 2008), indicating that, in the more realistic configuration, the
159 model successfully captures the timing and behavior of the system.

160 Comparing the *full* experiment with the aquaplanets allows zonally symmetric and asymmetric
161 processes to be distinguished. In *ap2* the peak precipitation shifts quickly between hemispheres.
162 Precipitation is strongest directly after a hemispheric shift. The rainfall is more intense than in
163 the *full* simulation, and is associated with larger shifts of the ITCZ from the equator. The strong
164 precipitation, and sharp transition of the ITCZ from southern to Northern Hemisphere, show that
165 zonal asymmetries are not essential for monsoon-like behavior. In contrast, in *ap20*, peak pre-
166 cipitation undergoes much smaller excursions from the equator. This simulation does not exhibit

167 other features of monsoon onset, with no reversal of the zonal wind direction (not shown). This
168 experiment is presented as an example of a climate that remains in the equinoctial regime.

169 To allow behavior before and after monsoon onset to be compared, 20 day periods, prior to and
170 post onset, have been selected for the *full* and *ap2* experiments. For each latitude and longitude
171 over the monsoon region, the first 5 day period (pentad) at which precipitation exceeds 8 mm/day
172 was identified. This was used as an indicator of the arrival of the monsoon. Maps of this onset date
173 (not shown) were then used to identify pentad ranges before and after the monsoon rains develop
174 over South and East Asia, or over the equivalent latitudes for *ap2*. For both *full* and *ap2*, the
175 monsoon is well established by mid July, and a 20 day period between the 16th of July and 5th of
176 August was selected (note the use of 30 day months in the experiments). As can be seen from Fig.
177 2, monsoon onset is earlier and more gradual in *full* than in *ap2*, suggesting that land-sea contrast
178 and topography advance onset. Pre-onset periods were selected as 1st-20th of June for *ap2*, and
179 1st-20th April for *full*.

180 Before monsoon onset, in both the *ap2* and *full* simulations, two overturning cells are seen,
181 roughly centered around the equator (Fig. 3). Strong, westerly zonal jets are found in both
182 hemispheres, driven by a combination of Coriolis acceleration and convergence of momentum
183 by midlatitude eddies. After onset, the circulation strength of the Southern Hemisphere Hadley
184 cell increases significantly, and the cell extends across the equator. Upper level easterlies develop,
185 and the Northern Hemisphere cell shifts polewards and weakens. In the *full* simulation, warm
186 surface temperatures over the coastline and Indian Ocean, and orographic forcing by the Tibetan
187 Plateau, result in low level ascent at $\sim 10^\circ\text{N}$. A second counter-clockwise overturning cell forms
188 to the north of this, with air descending on the poleward side of the Tibetan Plateau and warm-
189 ing at lower levels. In both simulations, after onset, eddy momentum flux convergence (colors)

190 is enhanced in the tropics at upper levels, indicating that the circulation is not axisymmetric. If
 191 absolute vorticity is low, it may, however, still be near to thermally direct, cf. Eq. (1).

192 Moist static energy (MSE) is also plotted in Fig. 2, which is defined:

$$MSE = c_p T + Lq + gz \quad (3)$$

193 where c_p is the specific heat capacity of dry air at constant pressure, T is temperature, L is the
 194 latent heat of vaporization of water, q is specific humidity, g is the gravitational constant, and
 195 z is height. In all simulations the strongest precipitation lies close to and slightly equatorward
 196 of the peak MSE, which in *ap2* is colocated with the northward boundary of the winter Hadley
 197 cell, consistent with Privé and Plumb (2007a). The peak MSE in the *full* simulation is located at
 198 $\sim 20^\circ\text{N}$, close to the boundary of the lower latitude component of the overturning circulation. The
 199 boundary of this cell will therefore be taken as the boundary of the monsoon circulation for this
 200 simulation.

201 Qualitatively, the general behaviors observed in the *ap2* and *full* experiments are similar. Precip-
 202 itation shifts quickly from one hemisphere to the other, spending little time centered on the equator,
 203 builds to a peak shortly after onset, then weakens and moves equatorward. This is accompanied
 204 by an expansion and strengthening of the Southern Hemisphere Hadley cell, and weakening of
 205 the Northern Hemisphere cell. However, the displacement and magnitude of rainfall observed in
 206 *full* is smaller than in *ap2*, and the structure of the overturning circulations after onset are very
 207 different.

208 Figs. 2 and 3 suggest some relationship between the strength of the meridional overturning cir-
 209 culation, Ψ , (and its associated precipitation), and the displacement of the ITCZ from the equator.
 210 The strongest precipitation occurs when the ITCZ is furthest north. Fig. 4 shows the peak strength
 211 of the meridional overturning circulation associated with the winter cell, Ψ_{max} , versus the latitude,

ϕ_0 , at which Ψ drops below 120×10^9 kg/s, both at 500 hPa. The latter is taken as indicative of the northward extent of the winter cell, and will be referred to as the ITCZ latitude. A non-zero threshold was chosen to isolate the southern component of the double overturning circulation over the Asian monsoon region in *full*, seen in Fig. 3. Each point on the plot corresponds to a multi-year pentad mean, so that the points closest to the equator correspond to the equinoctial season, while those furthest from the equator correspond to Northern Hemisphere summer. We note that the peak strengths in this figure are stronger than those found in reanalysis (e.g. Kållberg et al. 2005). 20 day multi-year means (Fig. 3) appear to be of more comparable magnitude to the reanalysis, suggesting that these large values are due to the shorter 5 day multi-year mean used for Fig. 4.

Fig. 4 confirms that there is a positive relationship between overturning strength and latitude. In addition, the gradient of the scatter increases as the ITCZ latitude increases, consistent with the proposed dynamical regime change during onset. Looking first at the aquaplanet experiments, the data from *ap2* (blue) suggests a change in the relationship between ϕ_0 and Ψ_{max} as the winter cell extends past $\sim 10^\circ\text{N}$. However, due to the large migrations of the ITCZ throughout the year in this experiment, little data is available at low latitudes. In *ap20* (black) the ITCZ does not shift as far north. The scatter is consistent with the relationship indicated by the *ap2* data, adding confidence to the change suggested by the shallower aquaplanet. Data from the *full* (red) simulation, averaged over the monsoon region, follow a qualitatively similar relationship to the simpler aquaplanet experiments.

The dashed lines show least squares best fits between the natural logarithms of both quantities. For the purposes of fitting, the data is divided into ITCZ latitudes north and south of 10°N (this choice of latitude is justified in the following section). The result is two power laws. For the aquaplanets, at lower latitudes a relation $\Psi_{max} = (228 \pm 19)\phi_0^{0.10 \pm 0.07}$ is obtained, while at higher latitudes the data follows $\Psi_{max} = (137 \pm 44)\phi_0^{0.33 \pm 0.11}$. The uncertainties quoted correspond to

2 standard errors, and confirm that the lower latitude and higher latitude fits are significantly different. From *full*, relations $\Psi_{max} = (235 \pm 38)\phi_0^{0.03 \pm 0.12}$ and $\Psi_{max} = (129 \pm 126)\phi_0^{0.32 \pm 0.37}$ are found. Both the lower and higher latitude power laws are similar to those for the aquaplanets, and in particular the exponent obtained for higher latitude ITCZs is strikingly close in value to that for the aquaplanets. However, we note there is large uncertainty in the fits for *full* due to the limited number and large spread of data points. Schneider and Bordoni (2008) investigated similar behavior in dry, steady state aquaplanet simulations in which the latitude of peak thermal forcing was varied. They identified exponents of $\frac{1}{5}$ and $\frac{4}{3}$. These are larger than the values obtained in our experiments, possibly reflecting that the seasonal cycle means that our simulations do not reach an equilibrium state.

Varying the choice of model level, or threshold for the ITCZ latitude, confirms that the qualitative behavior is not dependent on these, but that the precise relation obtained does vary. For ITCZ thresholds between 90 and 150×10^9 kg/s, and model levels between 400 and 850 hPa, values between 0.03 and 0.15 are generally obtained for the exponent in the lower latitude relation, while values between 0.25 and 0.44 are obtained for higher latitudes. The higher latitude exponents are usually more comparable between the aquaplanet and *full* simulations than the exponents for lower latitudes. The similar behavior observed in the aquaplanet and more realistic experiments, particularly in the higher latitude regime, suggests that the same dynamical processes are dominant over monsoon onset in both cases. The following sections will investigate the physics involved in more detail.

4. Vorticity budget

We have shown that once the ITCZ moves sufficiently far from the equator, the overturning circulation strengthens significantly in our model. We now investigate the dynamics responsible in

more detail. There is a general agreement in the literature that the transition to the monsoon regime is associated with a reduction in magnitude, or a reversal, of upper level absolute vorticity, so that the overturning becomes closer to thermally direct (Plumb and Hou 1992; Bordoni and Schneider 2008; Shaw 2014; Zhai and Boos 2015). The difference in absolute vorticity before and after monsoon onset in the *full* experiment (Fig. 5) is roughly consistent with this. Absolute vorticity decreases in a band centered over the Tibetan Plateau, and increases in the Southern Hemisphere. The sign of these changes corresponds to a reduction in magnitude in both hemispheres. However, an increase in absolute vorticity magnitude is found to the south of the Tibetan Plateau, between 0 and 30°N.

The causes of these changes can be investigated using the vorticity budget:

$$\begin{aligned} \frac{\partial \zeta}{\partial t} = & -\mathbf{u} \cdot \nabla (\zeta + f) - \omega \frac{\partial \zeta}{\partial p} \\ & - (\zeta + f) \nabla \cdot \mathbf{u} + \mathbf{k} \cdot \left(\frac{\partial \mathbf{u}}{\partial p} \times \nabla \omega \right) \end{aligned} \quad (4)$$

where ζ is relative vorticity, $\mathbf{u}$ is the horizontal wind vector, and ω is the pressure velocity. From left to right, the terms on the right hand side of the above equation correspond to horizontal advection of vorticity, vertical advection of vorticity, vortex stretching, and vortex tilting. We find that the terms in ω are negligible compared with the other tendencies. In the tropics and subtropics, the time means of the remaining terms are dominant over transient eddies.

To understand the physics of the regime change, we first discuss the behavior of the simpler *ap2* experiment. Fig. 6 shows the seasonal evolution of the multi-year pentad and zonal mean horizontal advection and stretching terms in Eq. (4). The tendencies themselves are shown in the left hand column. The remaining panels show a breakdown of the dominant components of these: $\partial(\bar{\zeta} + f)/\partial y$ and $\bar{v}$ for the horizontal advection term, and $\nabla \cdot \bar{\mathbf{u}}$ and $\bar{\zeta} + f$ for the stretching

term. Note that over these pentad means, relative vorticity is evolving and is not in steady state. Imbalances in the budget may therefore be interpreted as the driver of changes to the upper level absolute vorticity.

When the ITCZ crosses the equator, both tendencies change sign. In Northern Hemisphere summer, the tendency from horizontal advection is positive, and strongest close to the equator. Southward mean flow in the upper branch of the cross equatorial cell advects higher absolute vorticity air down-gradient. Horizontal advection therefore acts to increase the magnitude of absolute vorticity in the Northern Hemisphere, and to reduce the magnitude in the Southern Hemisphere. The tendency from vortex stretching is instead negative during Northern Hemisphere summer, due to divergent flow over regions of ascent in the Hadley and Ferrel cells. Vortex stretching associated with ascent in the ITCZ reduces the Northern Hemisphere absolute vorticity. The combined effect of the two tendencies is to produce and reinforce a broad region of reduced magnitude absolute vorticity air (Fig. 6, Panel f).

Looking more closely at the structure of the advective and stretching vorticity tendencies, it can be seen that monsoon onset and withdrawal are marked by peaks in both terms. We propose that these peaks, at $\sim 10^\circ\text{N}$, connect to the regime change seen in Fig. 4. Breaking down the terms into components reveals a possible explanation for this sudden change. Panels e and f show the divergence and absolute vorticity, which multiply to give the stretching term. Absolute vorticity is near zero close to the equator, where planetary vorticity is small. Divergence, however, is strongest while the ITCZ is close to the equator, where it is narrow (e.g. see Fig. 2). As the ITCZ shifts poleward, it moves into a region with non-negligible absolute vorticity. The stretching tendency increases significantly once the ITCZ reaches $\sim 10^\circ\text{N}$. This tendency acts to reduce the local absolute vorticity, allowing a more thermally direct circulation to develop. This strengthens the overturning, as seen in Fig. 4. The region of ascent expands as the ITCZ moves poleward,

303 so that the divergence is weaker and occupies a larger area. However, the stretching tendency is
304 still sufficient to lower absolute vorticity over a broad area (cf. Fig. 3), further strengthening the
305 circulation.

306 The peak in the stretching term provides a physical justification for the 10°N threshold used for
307 the fitting in Fig. 4. Once ascent moves sufficiently far from the equator that absolute vorticity
308 is not negligible, the stretching tendency increases. This lowers absolute vorticity over a larger
309 region, triggering a change to a thermally driven circulation.

310 The peaks of the horizontal advection tendency also relate to the narrow near-equator ITCZ.
311 The leading order component of this is $\bar{v}\partial(\bar{\zeta} + f)/\partial y$. When the ITCZ is close to the equator,
312 ascent, and the associated upper level divergent winds, occur over a narrower area, resulting in
313 strong meridional flow (Fig. 6, Panel c). At higher latitudes, $\bar{v}$ is negligible. The meridional
314 gradient of absolute vorticity (Panel b) peaks on the equator, decreases poleward, and peaks again
315 in the midlatitudes. The combination of these terms yields strong horizontal advection close to the
316 equator during monsoon onset and withdrawal.

317 The roles of the horizontal wind speed, absolute vorticity, and their gradients in determining
318 the sign and magnitude of the vorticity tendencies in the equinoctial regime and over onset are
319 summarized in Fig. 7. Over most of the year, with the exception of monsoon onset, the model
320 climate evolves slowly (for example the residual of the zonal momentum budget is near zero, see
321 Fig. 9f). Monsoon onset appears to correspond to a fast transition between climates close to the
322 two steady-state solutions to Eq. (1), driven by the feedbacks described above. While the present
323 paper discusses the processes involved over monsoon onset in a time-varying case, the steady state
324 budget will be explored further in future work.

325 The key features of the *ap2* vorticity budget are shared with the *full* experiment (Fig. 8). The
326 reduction in absolute vorticity magnitude during monsoon onset is less marked in *full* than in *ap2*,

327 but a broadening of the low magnitude region is still observed in Northern Hemisphere summer.
 328 Horizontal vorticity advection and vortex stretching are again the dominant terms in the budget,
 329 and act to reduce the magnitude of the southern and Northern Hemisphere absolute vorticity re-
 330 spectively. Differences from *ap2* can, however, be seen, which relate to the structure of the land
 331 and topography. During Northern Hemisphere summer, both terms now have multiple peaks in
 332 latitude. Looking first at the stretching term, peaks are found corresponding to the ascending re-
 333 gions seen in Fig. 3. At $\sim 20^\circ\text{N}$, ascent is forced by the Tibetan plateau, and by warm sea surface
 334 temperatures along the coastline. This is balanced by horizontal advection over the upper branch
 335 of the Hadley cell, resulting in the region of increased absolute vorticity seen in Fig. 5. Air de-
 336 scends to the north of the plateau, and ascends again between $\sim 30 - 40^\circ\text{N}$. The vortex stretching
 337 associated with this ascending region dominates over horizontal advection, so that a broad region
 338 of low absolute vorticity air is still realized. From the breakdown of the stretching term (Fig. 8,
 339 Panels e and f), similar behavior to *ap2* is found. Divergence is strong along the equator, but, as
 340 absolute vorticity is small here, vortex stretching is weak. Once ascent is forced further to the
 341 north, stretching strengthens, and acts to lower absolute vorticity.

342 In both Figs. 6 and 8, we have identified a sharp increase in the vorticity tendency due to
 343 stretching during monsoon onset. This occurs once the ITCZ reaches latitudes where absolute
 344 vorticity is no longer negligible. Vortex stretching, and cross-equatorial advection of vorticity,
 345 then create a broad region of low, although non-zero, absolute vorticity (see Fig. 7). In *ap2* it is
 346 the warming of the surface by the seasonally varying insolation that results in off-equatorial ascent,
 347 while in *full* the Tibetan plateau also plays a role. Regardless of the driving mechanism, once the
 348 ITCZ shifts sufficiently far from the equator, vortex stretching and advection by the meridional
 349 overturning flow result in a stronger, more thermally direct cell.

5. Role of zonal asymmetry

While we have found that vortex stretching in off-equatorial regions of ascent is key in the transition to a thermally direct regime in both *ap2* and *full*, differences in behavior between the experiments are evident from the vorticity budgets. The processes responsible for triggering the transition in the more realistic experiment will now be investigated in further detail. In particular, we would like to distinguish the roles in *full* of zonally symmetric warming of a low thermal inertia surface, the key process in *ap2*, and land-sea contrast and orographic forcing. In the derivation of the vorticity budget, pressure gradient terms cancel. We therefore evaluate the zonal momentum budget in order to assess the effect of these gradients, which we expect to be forced by zonal asymmetries in the *full* simulation.

Separating nonlinear terms into a zonal-temporal mean state, and transient and stationary eddies (e.g. Yang et al. 2012), the primitive equation for u in pressure coordinates can be expressed:

$$\begin{aligned} \frac{\partial [\bar{u}]}{\partial t} = & f[\bar{v}] - \frac{\partial [\bar{\Phi}]}{\partial x} + [\overline{\mathcal{F}^{(x)}}] \\ & - \left([\bar{u}] \frac{\partial [\bar{u}]}{\partial x} + [\bar{v}] \frac{\partial [\bar{u}]}{\partial y} + [\bar{\omega}] \frac{\partial [\bar{u}]}{\partial p} \right) \\ & - \left[\bar{u}^* \frac{\partial \bar{u}^*}{\partial x} + \bar{v}^* \frac{\partial \bar{u}^*}{\partial y} + \bar{\omega}^* \frac{\partial \bar{u}^*}{\partial p} \right] \\ & - \left[\frac{\partial \overline{u'u'}}{\partial x} + \frac{\partial \overline{u'v'}}{\partial y} + \frac{\partial \overline{u'\omega'}}{\partial p} \right] \end{aligned} \quad (5)$$

where u , v and ω are the zonal, meridional and vertical winds in pressure coordinates, $\mathcal{F}^{(x)}$ describes frictional damping, and Φ represents geopotential. As in Eq. (1), square brackets indicate a zonal mean, and stars denote a deviation from this. Overbars and dashes similarly represent a temporal mean and transient eddies. The temporal mean state is again defined as the multi-year pentad mean. The terms on the top row of the right hand side correspond to acceleration of the zonal and temporal mean zonal wind via Coriolis force, geopotential gradients, and friction. The

368 bracketed terms represent advection of zonal momentum by the mean flow, stationary eddies, and
369 transient eddies.

370 To identify the response when boundary conditions are zonally symmetric, we again look first
371 at the *ap2* experiment. The terms in the zonal momentum budget are plotted as a function of
372 latitude and time in Fig. 9. Zonal means have been taken over all longitudes. The advective terms
373 are shown as sums of the contributions associated with the mean state, transient and stationary
374 eddies. Geopotential gradient and the total zonal wind tendency are included for completeness.
375 The former is zero for a zonal average around a latitude circle, as it is a zonal gradient. The
376 contribution from stationary eddies is also negligible, as this experiment has zonally symmetric
377 boundary conditions.

378 In late June, a fast change in the sign of $f[\bar{v}]$ (Panel a) can be seen, reflecting the change in the
379 direction of the upper level meridional wind from northward to southward flow. This occurs as
380 the ITCZ shifts into the Northern Hemisphere, bringing intense precipitation. The transition is
381 associated with a change in sign, and an increase in magnitude, of the advection of momentum by
382 the zonal and temporal mean state (Panel b). This term is dominated by the meridional component,
383 $[\bar{v}] \frac{\partial [\bar{u}]}{\partial y}$. The balance of Coriolis acceleration and advection is consistent with low zonal mean
384 absolute vorticity throughout the tropics and subtropics, as was discussed above.

385 While the boundary conditions in *ap2* are symmetric, transient eddies (Panel d) still play a
386 role in the momentum budget. As the monsoon arrives, the transient eddy activity associated
387 with the midlatitude storm track shifts poleward in the winter hemisphere, and weakens in the
388 summer hemisphere. A weak increase in low latitude eddy momentum flux convergence is also
389 seen in the summer hemisphere after the shift in the precipitation band. This low latitude eddy
390 momentum flux convergence lags monsoon onset and damps the upper level easterlies (see Fig.
391 3), suggesting it relates to instability associated with the easterly jet. Regions of stronger eddy

activity are associated with regions of higher magnitude absolute vorticity, e.g. Eq. (1) and Fig. 6f.

The equivalent budget for the *full* simulation is shown in Fig. 10. Eddy quantities are here defined relative to a zonal mean state spanning 60-150°E. As in *ap2*, the arrival of monsoonal precipitation is associated with a reversal and increase in magnitude of the upper level meridional flow (Panel a). The geopotential gradient (Panel c) is no longer zero when averaged over this confined longitude region. The dominant balance is now between Coriolis and geopotential gradient north of $\sim 20^\circ\text{N}$, and Coriolis and mean state advection (Panel b) to the south of this. At these lower latitudes the budget resembles that of *ap2*, with axisymmetric balance dominant, but transient and stationary eddies (Panels d and e) also contributing weakly. The implication is that the southern cell seen in Fig. 3, which is responsible for the majority of precipitation, behaves similarly to the overturning cell in *ap2*, e.g. Fig. 4. However, consistent with the multiple peaks in stretching tendency seen in the vorticity budget, geopotential gradients also play a key role in extending the overturning circulation northward, and so lowering the absolute vorticity in the Northern Hemisphere.

Subtropical planetary scale waves have been demonstrated to trigger a monsoon-like regime transition in an aquaplanet simulation with SST anomalies (Shaw 2014). The momentum budget analysis allows us to connect this to zonally symmetric theories for the monsoon regime change. Fig. 11 shows the geopotential and horizontal streamfunction at 150 hPa and 850 hPa after monsoon onset. The Asian and American continents are warm relative to the ocean, and produce a wavenumber 2 signal in the lower level geopotential, centered on 30°N , with troughs over the land. This is mirrored by increased geopotential at upper levels. The streamfunction shows a cyclone over the Asian continent, in geostrophic balance with the geopotential. The South and East Asian monsoon regions lie in the area of this wave where the sign of the geopotential gradient is

such as to extend the overturning circulation further north, as seen in Fig. 3. South of 20°N , where geopotential gradient is weak, and streamlines are roughly oriented east-west, the wave forces a localized thermally direct circulation, with dynamics similar to that of the aquaplanets.

The Asian monsoon is often considered to consist of two interacting systems, the South and East Asian monsoons. The former is considered an essentially tropical system, whereas in the latter midlatitude dynamics are thought to play more of a role. The monsoon arrives first over the Indochina peninsula, then progresses westward over India (Wu and Wang 2000). The results presented here suggest that the East Asian component of the monsoon relates to the stationary wave that emerges as the Northern Hemisphere continents heat up. Northward flow and ascent are forced over the continent, and vortex stretching acts to lower upper level absolute vorticity. The broad region of low absolute vorticity then triggers a transition of the lower latitude overturning circulation into an intense, thermally direct regime over South Asia.

6. Summary and Discussion

Various studies using aquaplanet models have proposed that monsoon onset relates to a change of the tropical circulation from a transient eddy-driven to a thermally direct regime. To investigate the relevance of this perspective to the Earth's monsoons, we have performed a range of experiments with an idealized atmosphere model coupled to a slab ocean, with and without a simple description of land.

In both our aquaplanet simulations and in a semi-realistic set-up that has an Earth-like distribution of continents and topography, we find behaviour consistent with a transition to a thermally direct regime is observed. When the ITCZ shifts sufficiently off the equator, the strength of the cross equatorial Hadley cell increases, accompanied by a sudden increase in precipitation. In the aquaplanet experiments, for small excursions of the ITCZ we find that the change in overturning

439 strength per degree latitude can be described by a power law, with exponent 0.10 ± 0.07 (Fig. 4).
440 After a threshold latitude of $\sim 10^\circ\text{N}$ is passed, a larger increase in strength with latitude is ob-
441 served, and the best-fit exponent increases to 0.33 ± 0.11 . The uncertainty ranges confirm that the
442 two power laws are significantly different, and that some change in behaviour occurs as the ITCZ
443 moves to higher latitudes. Similar relations are obtained for the *full* simulation, with exponents
444 0.03 ± 0.12 and 0.32 ± 0.37 . These are similar to those for the aquaplanets, particularly in the off-
445 equatorial ITCZ regime, suggesting that similar dynamical processes are at work despite the large
446 differences in the surface boundary conditions between the aquaplanet and *full* experiments.

447 A consistent feature of the various theories for monsoon onset in idealized studies is a reduction
448 in the magnitude of the upper-level absolute vorticity, suggesting a transition to a thermally direct
449 regime. However, the cause of this reduction and the role of zonal asymmetries, and hence the rel-
450 evance of this result to the real world, has been disputed. Figure 7 shows a schematic summarizing
451 the findings of the present study on the vorticity tendencies in both the equinoctial and monsoon
452 regimes. Our analysis of the upper level vorticity budget for the *ap2* experiment shows that as
453 the ITCZ migrates into the Northern Hemisphere, the magnitude of the vortex stretching term in
454 the budget increases rapidly. This term is the product of divergence and absolute vorticity, and it
455 appears that a regime change is triggered when the region of divergence moves into a region of
456 non-negligible absolute vorticity. In our simulations this occurs at $\sim 10^\circ\text{N}$. The circulation shifts
457 towards a thermally direct regime and so strengthens, and the region of ascent broadens. Vortex
458 stretching associated with the ascending air, and horizontal advection of vorticity southward across
459 the equator, act as a positive feedback, maintaining and extending a broad region of low magnitude
460 upper level absolute vorticity air.

461 Similar processes are at work in the *full* experiment, which shows a strong decrease in Northern
462 Hemisphere absolute vorticity centered over the Tibetan plateau. However, we find that, in this

simulation, warming of the continent and Tibetan plateau forces a stationary wave, and ascent over the plateau. Off-equatorial ascent, so that divergence is colocated with non-negligible absolute vorticity, again triggers a strong, negative upper level vortex stretching tendency. The reduced upper level vorticity helps to move the lower latitude circulation into a more thermally direct regime, which resembles that in *ap2*. Vortex stretching and horizontal advection of vorticity again act to maintain low absolute vorticity air in the upper troposphere.

The upper level zonal momentum budget highlights the different natures of the lower and higher latitude circulations in *full*. North of $\sim 20^\circ\text{N}$, the flow is predominantly geostrophic. To the south the balance instead resembles a localized thermally direct circulation, similar to the aquaplanet. We propose that these differently forced components of the monsoon circulation connect to the East and South Asian monsoon regimes. This would suggest that the East Asian monsoon onset can be expected to be sensitive to changes to midlatitude stationary waves, with changes to the South Asian monsoon then occurring as a consequence of this.

These results reconcile some of the apparent disagreement between the regime changes previously discussed. The key process for monsoon onset is demonstrated to be a reduction of upper level absolute vorticity, but this may be caused by a variety of mechanisms. Shallow aquaplanets respond strongly to the seasonal cycle, shifting the ITCZ off of the equator and instigating a feedback which lowers absolute vorticity (cf. Bordoni and Schneider 2008). In this case, we find that the resulting circulation is more thermally direct, but not necessarily axisymmetric, with low latitude eddy activity not appearing to hamper the transition (e.g. Fig. 3). Stationary planetary waves have previously been shown to be able to produce an absolute vorticity reversal, which results in a thermally direct circulation over areas with a broad region of low absolute vorticity (Shaw 2014). This case is more relevant to our more realistic experiment, where the cyclone that forms over the

low pressure region associated with the Tibetan plateau seems responsible for the onset of a low latitude, thermally direct circulation, localized over Asia.

This study raises several avenues for future work. Although we have established that ascent over the Asian continent is important in triggering monsoon onset, the individual roles of land-sea contrast and topography, and the effect of continental geometry, remain to be investigated. While we believe we have identified the mechanism causing the sharp transition during monsoon onset, we do not yet have an analytical theory to describe this process. Additional idealized simulations, both time-varying and steady state, may help in the formulation of this. Finally, we also hope to investigate the relevance of the regime change processes discussed here to climate change predictions over Asia, by connecting the results of these behavior simulations to data from Earth System models, and from reanalysis.

Acknowledgments. The work was supported by the UK-China Research & Innovation Partnership Fund, through the Met Office Climate Science for Service Partnership (CSSP) China, as part of the Newton Fund. GKV also acknowledges support from the Royal Society (Wolfson Foundation), the Leverhulme Trust, and NERC. We thank Martin Jucker for coupling RRTM to the GFDL idealized model, and Stephen Thomson for many modelling innovations.

References

- Bordoni, S., and T. Schneider, 2008: Monsoons as eddy-mediated regime transitions of the tropical overturning circulation. *Nature Geoscience*, **1** (8), 515–519, doi:10.1038/ngeo248.
- Chao, W. C., 2000: Multiple quasi equilibria of the ITCZ and the origin of monsoon onset. *J. Atmos. Sci.*, **57** (5), 641–652.

Chao, W. C., and B. Chen, 2001: The Origin of Monsoons. *Journal of the Atmospheric Sciences*, **58** (22), 3497–3507, doi:10.1175/1520-0469(2001)058<3497:TOOM>2.0.CO;2.

Frierson, D. M. W., 2007: The Dynamics of Idealized Convection Schemes and Their Effect on the Zonally Averaged Tropical Circulation. *Journal of the Atmospheric Sciences*, **64** (6), 1959–1976, doi:10.1175/JAS3935.1.

Frierson, D. M. W., I. M. Held, and P. Zurita-Gotor, 2006: A Gray-Radiation Aquaplanet Moist GCM. Part I: Static Stability and Eddy Scale. *Journal of the Atmospheric Sciences*, **63** (10), 2548–2566, doi:10.1175/JAS3753.1.

Gill, A. E., 1980: Some simple solutions for heat induced tropical circulation. *Quarterly Journal of the Royal Meteorological Society*, **106** (449), 447–462, doi:10.1002/qj.49710644905.

Held, I. M., and A. Y. Hou, 1980: Nonlinear Axially Symmetric Circulations in a Nearly Inviscid Atmosphere. 515–533 pp., doi:10.1175/1520-0469(1980)037<0515:NASCIA>2.0.CO;2.

Kållberg, P., P. Berrisford, B. Hoskins, A. Simmons, S. Uppala, S. Lamy-Thépaut, and R. Hine, 2005: ERA-40 Atlas. *ECMWF Re-Analysis Project Report Series 19*.

Kim, H.-K., and S. Lee, 2001: Hadley cell dynamics in a primitive equation model. Part II: Non-axisymmetric flow. *J. Atmos. Sci.*, **58** (19), 2859–2871.

Krishnamurti, T. N., and Y. Ramanathan, 1982: Sensitivity of the Monsoon Onset to Differential Heating. 1290–1306 pp., doi:10.1175/1520-0469(1982)039<1290:SOTMOT>2.0.CO;2.

Li, C., and M. Yanai, 1996: The onset and interannual variability of the asian summer monsoon in relation to land-sea thermal contrast. 358–375 pp., doi:10.1175/1520-0442(1996)009<0358:TOAIVO>2.0.CO;2.

528 Lindberg, C., and A. J. Broccoli, 1996: Representation of topography in spectral climate mod-
 529 els and its effect on simulated precipitation. 2641–2659 pp., doi:10.1175/1520-0442(1996)
 530 009<2641:ROTISC>2.0.CO;2.

531 Lindzen, R. S., and A. Y. Hou, 1988: Hadley circulation for zonally averaged heating centered off
 532 the equator. *J. Atmos. Sci.*, **45**, 2416–2427.

533 Mlawer, E. J., S. J. Taubman, P. D. Brown, M. J. Iacono, and S. A. Clough, 1997: Radiative trans-
 534 fer for inhomogeneous atmospheres: RRTM, a validated correlated-k model for the longwave.
 535 *Journal of Geophysical Research*, **102 (D14)**, 16 663–16 682, doi:10.1029/97JD00237, URL
 536 <http://doi.wiley.com/10.1029/97JD00237>.

537 O’Gorman, P. A., and T. Schneider, 2008: The Hydrological Cycle over a Wide Range of Climates
 538 Simulated with an. *Journal of Climate*, **21**, 3815–3832, doi:10.1175/2007JCLI2065.1.

539 Plumb, R. a., and A. Y. Hou, 1992: The response of a zonally symmetric atmosphere to subtropical
 540 thermal forcing - Threshold behavior. 1790–1799 pp., doi:10.1175/1520-0469(1992)049<1790:
 541 TROAZS>2.0.CO;2.

542 Privé, N. C., and R. A. Plumb, 2007a: Monsoon Dynamics with Interactive Forcing. Part I: Ax-
 543 isymmetric Studies. *Journal of the Atmospheric Sciences*, **64 (5)**, 1417–1430, doi:10.1175/
 544 JAS3916.1, URL <http://journals.ametsoc.org/doi/abs/10.1175/JAS3917.1>.

545 Privé, N. C., and R. A. Plumb, 2007b: Monsoon Dynamics with Interactive Forcing. Part II:
 546 Impact of Eddies and Asymmetric Geometries. *Journal of the Atmospheric Sciences*, **64 (5)**,
 547 1431–1442, doi:10.1175/JAS3917.1.

548 Russell, G. L., J. R. Miller, and L.-c. Tsang, 1985: Seasonal oceanic heat transports computed
 549 from an atmospheric model. *Dynamics of Atmospheres and Oceans*, **9**, 253–271.

- 550 Schneider, E. K., 1977: Axially symmetric steady-state models of the basic state for instability
551 and climate studies. Part II: Nonlinear calculations. *J. Atmos. Sci.*, **34**, 280–297.
- 552 Schneider, T., and S. Bordoni, 2008: Eddy-Mediated Regime Transitions in the Seasonal Cycle
553 of a Hadley Circulation and Implications for Monsoon Dynamics. *Journal of the Atmospheric*
554 *Sciences*, **65** (1), 915–934, doi:10.1175/2007JAS2415.1.
- 555 Shaw, T. A., 2014: On the Role of Planetary-Scale Waves in the Abrupt Seasonal Transition of
556 the Northern Hemisphere General Circulation. *Journal of the Atmospheric Sciences*, **71** (5),
557 1724–1746, doi:10.1175/JAS-D-13-0137.1, URL [http://journals.ametsoc.org/doi/abs/10.1175/](http://journals.ametsoc.org/doi/abs/10.1175/JAS-D-13-0137.1)
558 [JAS-D-13-0137.1](http://journals.ametsoc.org/doi/abs/10.1175/JAS-D-13-0137.1).
- 559 Wu, R., and B. Wang, 2000: Interannual variability of summer monsoon onset over the western
560 North Pacific and the underlying processes. *Journal of Climate*, **13** (14), 2483–2501, doi:10.
561 1175/1520-0442(2000)013<2483:IVOSMO>2.0.CO;2.
- 562 Wu, R., and B. Wang, 2001: Multi-stage onset of the summer monsoon over the western North
563 Pacific. *Climate Dynamics*, **17** (4), 277–289, doi:10.1007/s003820000118.
- 564 Yang, W., R. Seager, and M. Cane, 2012: Zonal Momentum Balance in the Tropical Atmospheric
565 Circulation during the Global Monsoon Mature Months. *J. Atmos. Sci.*, **70** (2), 583–599, doi:10.
566 1175/JAS-D-12-0140.1, URL <http://journals.ametsoc.org/doi/abs/10.1175/JAS-D-12-0140.1>.
- 567 Yin, M. T., 1949: Synoptic-Aerologic Study of the Onset of the Summer Monsoon Over India and
568 Burma. 393–400 pp., doi:10.1175/1520-0469(1949)006<0393:SASOTO>2.0.CO;2.
- 569 Zhai, J., and W. Boos, 2015: Regime Transitions of Cross-Equatorial Hadley Circulations
570 with Zonally Asymmetric Thermal Forcings. *Journal of the Atmospheric Sciences*, **72**,

571 3800–3818, doi:10.1175/JAS-D-15-0025.1, URL <http://journals.ametsoc.org/doi/abs/10.1175/>
572 JAS-D-15-0025.1.

573 Zhang, Y., T. Li, B. Wang, and G. Wu, 2002: Onset of the summer monsoon over the Indochina
574 Peninsula: Climatology and interannual variations. *Journal of Climate*, **15** (**22**), 3206–3221,
575 doi:10.1175/1520-0442(2002)015<3206:OOTSMO>2.0.CO;2.

LIST OF FIGURES

- Fig. 1.** Configuration of land and topography used in *full* experiment. Colors over land show surface height, contour interval 500 m. Colors over ocean show the average ocean heat flux in JJA, contour interval 25 Wm^{-2} 30
- Fig. 2.** Seasonal cycle of zonal mean precipitation, mm/day, (colors) for (a) *full*, (b) *ap2*, (c) *ap20*. Averages are over all longitudes for the aquaplanets, and between 60 and 150°E for *full*. The 8 mm/day contour is marked in black. Gray contours indicate moist static energy at 850 hPa , the contour interval is 20 kJ/kg 31
- Fig. 3.** Mass streamfunction, Ψ (black contours), zonal wind (gray contours), and combined transient and stationary eddy momentum flux convergence (colors) for (a, b) *ap2* and (c, d) *full*. Quantities are zonal and time averages over (a, c) 20 days prior to monsoon onset and (b, d) 20 days post monsoon onset. Onset is defined as arrival of precipitation above 8 mm/day (cf. Fig. 2). contour interval is $60 \times 10^9 \text{ kg/s}$ for Ψ , and 10 m/s for zonal wind. Eddy momentum flux convergence is in units of $\text{ms}^{-1}\text{day}^{-1}$ 32
- Fig. 4.** Peak strength of the meridional overturning associated with the cross-equatorial Hadley cell versus latitude of the ascending branch, both at 500 hPa and in Northern Hemisphere summer. Black crosses denote *ap2*, blue *ap20* and red *full*. For *full*, points are a zonal average between 60 and 150°E . The black and red lines indicate best fit power laws for the aquaplanets and the *full* experiment respectively. Latitudes south and north of 10°N are fitted separately. 33
- Fig. 5.** Difference between 150 hPa absolute vorticity averaged over 4 pentads before and after monsoon onset, day^{-1} (colors). Black contours show the average 150 hPa absolute vorticity after onset, with contour interval 2 day^{-1} 34
- Fig. 6.** Seasonal cycle and breakdown of terms in the vorticity budget Eq. (4) for *ap2* at 150 hPa , as a multi-year pentad mean, averaged over all longitudes: (a) $-\bar{\mathbf{u}} \cdot \nabla(\bar{\zeta} + f)$ (day^{-2}), (b) $\partial(\bar{\zeta} + f)/\partial y$ ($\text{m}^{-1}\text{s}^{-1}$), (c) $\bar{v}$ (m/s), (d) $-(\bar{\zeta} + f)\nabla \cdot \bar{\mathbf{u}}$ (day^{-2}), (e) $\nabla \cdot \bar{\mathbf{u}}$ (day^{-1}), (f) $\bar{\zeta} + f$ (day^{-1}). 35
- Fig. 7.** Schematic summarising the behaviour in the equinoctial regime and over monsoon onset. Shading indicates absolute vorticity, arrows indicate the direction and strength of the overturning circulation. a) **Equinoctial regime:** Ascent occurs close to the equator where absolute vorticity is near zero. Off the equator, absolute vorticity and its gradient are non-negligible, but the weak zonal mean tendencies are balanced by eddies. The total vorticity tendency is small, and the circulation is near to the ‘weak overturning’ steady-state solution of Eq. (1). b) **At monsoon onset:** Ascent occurs off the equator where absolute vorticity is non-negligible. This results in a negative upper level vorticity tendency due to vortex stretching in the Northern Hemisphere, reducing the magnitude of the absolute vorticity. The circulation begins to transition to the thermally direct steady-state solution of Eq. (1), and the overturning strengthens. The cross-equatorial meridional flow is associated with a positive vorticity tendency due to horizontal advection, which reduces absolute vorticity magnitude in the Southern Hemisphere. The vorticity tendencies due to vortex stretching and horizontal advection therefore both act to further reduce the magnitude of absolute vorticity. This positive feedback allows overturning strength, and the associated precipitation, to increase rapidly over onset. 36
- Fig. 8.** As Fig. 6 but for *full*, with zonal means here defined between 60 and 150°E 37

- 620 **Fig. 9.** Seasonal cycle of the terms in the zonal momentum budget Eq. (5) for *ap2* at 150 hPa:
621 (a) $f[\bar{v}]$, (b) $-\left([\bar{u}]\frac{\partial[\bar{u}]}{\partial x} + [\bar{v}]\frac{\partial[\bar{u}]}{\partial y} + [\bar{\omega}]\frac{\partial[\bar{u}]}{\partial p}\right)$, (c) $-\frac{\partial[\Phi]}{\partial x}$, (d) $-\left[\bar{u}^*\frac{\partial\bar{u}^*}{\partial x} + \bar{v}^*\frac{\partial\bar{u}^*}{\partial y} + \bar{\omega}^*\frac{\partial\bar{u}^*}{\partial p}\right]$, (e)
622 $-\left[\frac{\partial\bar{u}'\bar{u}'}{\partial x} + \frac{\partial\bar{u}'\bar{v}'}{\partial y} + \frac{\partial\bar{u}'\bar{\omega}'}{\partial p}\right]$, (f) $\partial u/\partial t$. All quantities are multi-year pentad and zonal averages.
623 Units are $\text{ms}^{-1}\text{day}^{-1}$. The 8 mm/day precipitation contour is marked on all plots in gray
624 as an indicator of monsoon onset. Absolute vorticity is overplotted in black on (b), with a
625 contour interval of 2 day^{-1} 38
- 626 **Fig. 10.** As Fig. 9 but for *full*, with the zonal average taken between 60 and 150°E 39
- 627 **Fig. 11.** (a) 850 hPa geopotential, contour interval $250 \text{ m}^2\text{s}^{-2}$, (colors) and 150 hPa geopotential,
628 contour interval $2000 \text{ m}^2\text{s}^{-2}$, (contours) for the *full* experiment, averaged over 20 days post
629 monsoon onset. (b) Equivalent plots for the horizontal streamfunction, contour intervals are
630 $5 \times 10^6 \text{ m}^2\text{s}^{-1}$ at 850 hPa, and $2 \times 10^7 \text{ m}^2\text{s}^{-1}$ at 150 hPa. Dashed lines indicate negative val-
631 ues. The land mask used is shown in gray. Missing values at 850 hPa are due to intersection
632 with topography. 40

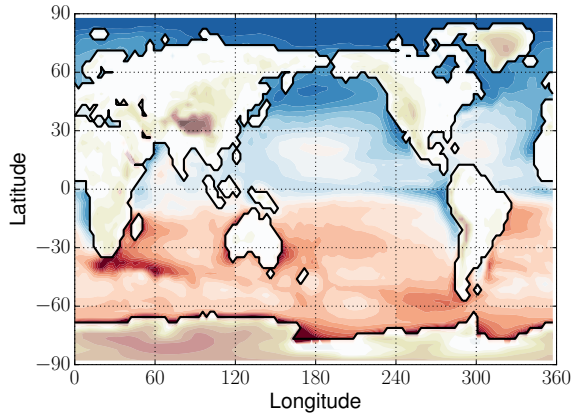


FIG. 1. Configuration of land and topography used in *full* experiment. Colors over land show surface height, contour interval 500 m. Colors over ocean show the average ocean heat flux in JJA, contour interval 25 Wm^{-2} .

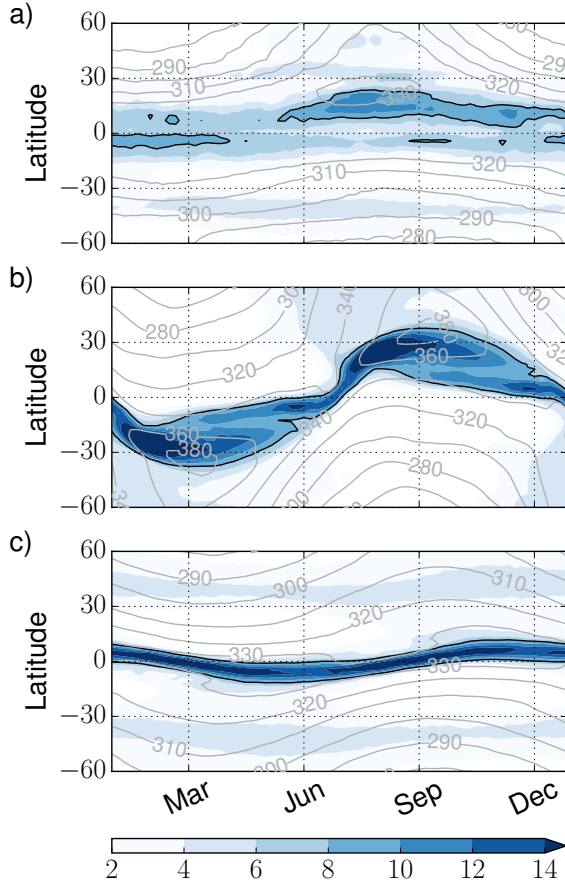


FIG. 2. Seasonal cycle of zonal mean precipitation, mm/day, (colors) for (a) *full*, (b) *ap2*, (c) *ap20*. Averages
are over all longitudes for the aquaplanets, and between 60 and 150°E for *full*. The 8 mm/day contour is marked
in black. Gray contours indicate moist static energy at 850 hPa, the contour interval is 20 kJ/kg.

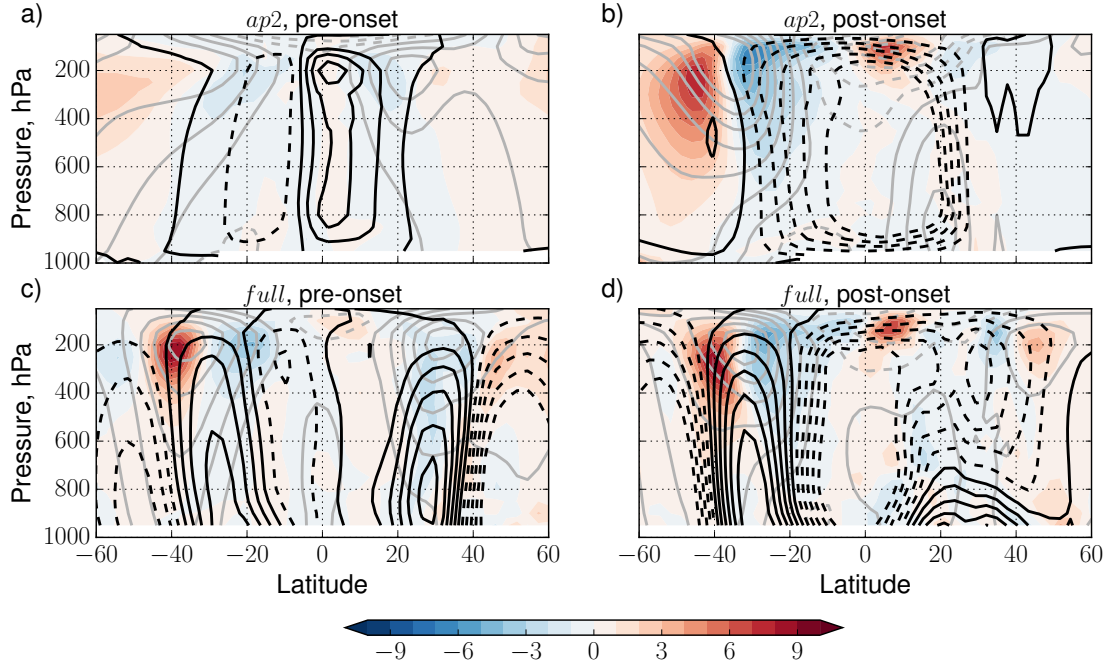
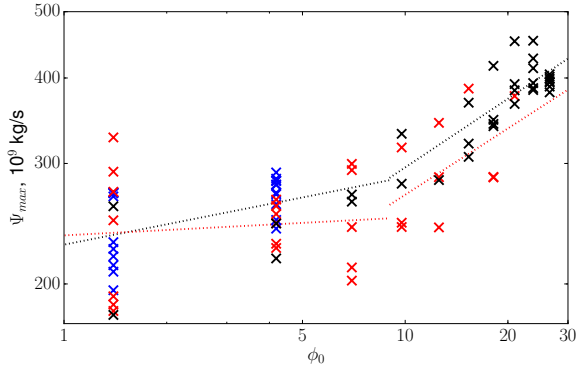


FIG. 3. Mass streamfunction, Ψ (black contours), zonal wind (gray contours), and combined transient and stationary eddy momentum flux convergence (colors) for (a, b) *ap2* and (c, d) *full*. Quantities are zonal and time averages over (a, c) 20 days prior to monsoon onset and (b, d) 20 days post monsoon onset. Onset is defined as arrival of precipitation above 8 mm/day (cf. Fig. 2). contour interval is 60×10^9 kg/s for Ψ , and 10 m/s for zonal wind. Eddy momentum flux convergence is in units of $\text{ms}^{-1}\text{day}^{-1}$.



643 FIG. 4. Peak strength of the meridional overturning associated with the cross-equatorial Hadley cell versus
 644 latitude of the ascending branch, both at 500 hPa and in Northern Hemisphere summer. Black crosses denote
 645 *ap2*, blue *ap20* and red *full*. For *full*, points are a zonal average between 60 and 150°E. The black and red lines
 646 indicate best fit power laws for the aquaplanets and the *full* experiment respectively. Latitudes south and north
 647 of 10°N are fitted separately.

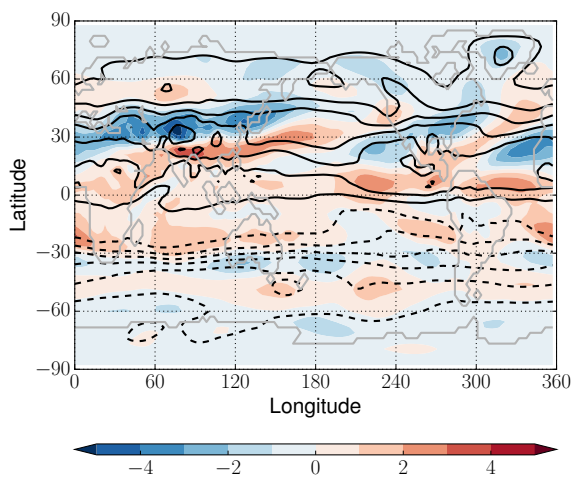


FIG. 5. Difference between 150 hPa absolute vorticity averaged over 4 pentads before and after monsoon onset, day^{-1} (colors). Black contours show the average 150 hPa absolute vorticity after onset, with contour interval 2 day^{-1} .

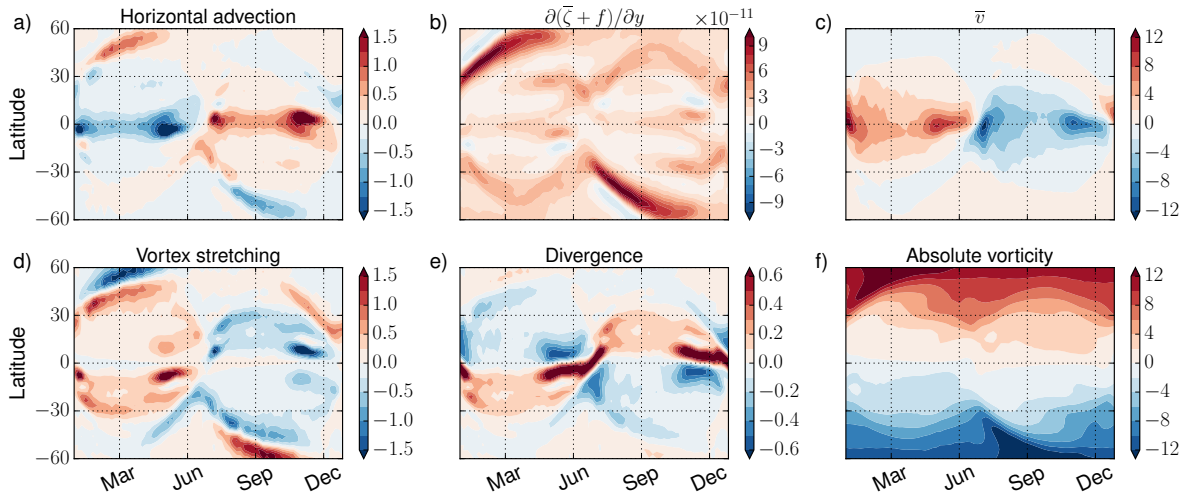


FIG. 6. Seasonal cycle and breakdown of terms in the vorticity budget Eq. (4) for *ap2* at 150 hPa, as a multi-year pentad mean, averaged over all longitudes: (a) $-\bar{\mathbf{u}} \cdot \nabla(\bar{\zeta} + f)$ (day^{-2}), (b) $\partial(\bar{\zeta} + f)/\partial y$ ($\text{m}^{-1}\text{s}^{-1}$), (c) $\bar{v}$ (m/s), (d) $-(\bar{\zeta} + f)\nabla \cdot \bar{\mathbf{u}}$ (day^{-2}), (e) $\nabla \cdot \bar{\mathbf{u}}$ (day^{-1}), (f) $\bar{\zeta} + f$ (day^{-1}).

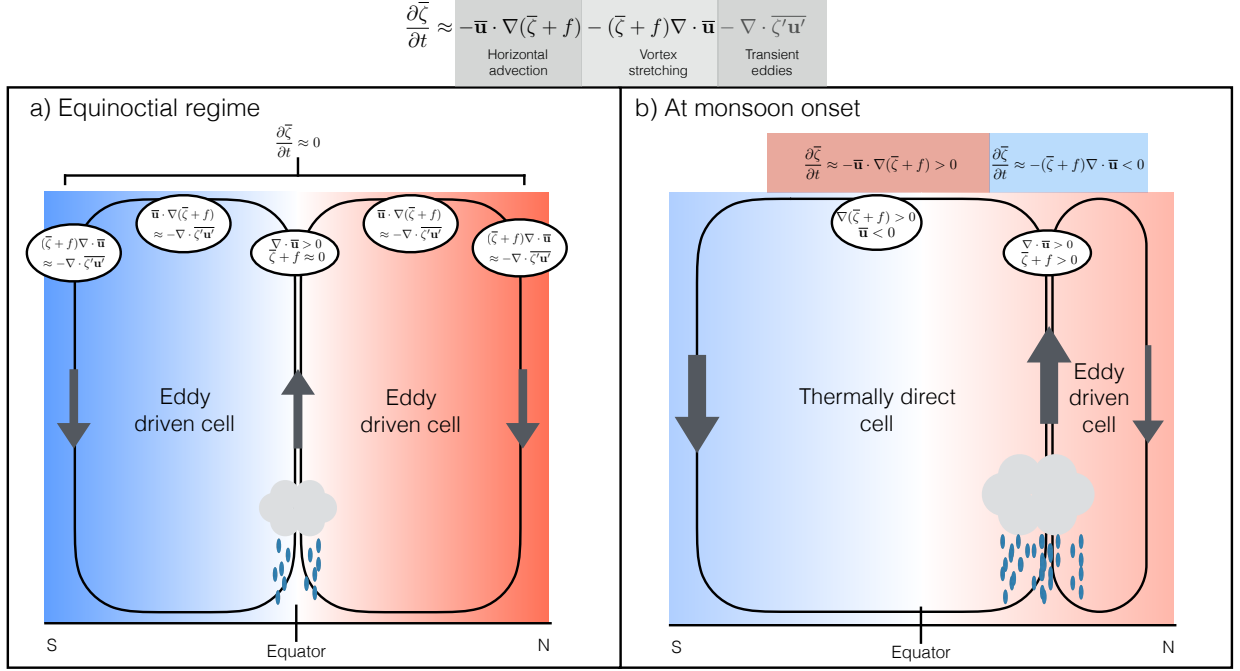


FIG. 7. Schematic summarising the behaviour in the equinoctial regime and over monsoon onset. Shading indicates absolute vorticity, arrows indicate the direction and strength of the overturning circulation. a) **Equinoctial regime**: Ascent occurs close to the equator where absolute vorticity is near zero. Off the equator, absolute vorticity and its gradient are non-negligible, but the weak zonal mean tendencies are balanced by eddies. The total vorticity tendency is small, and the circulation is near to the ‘weak overturning’ steady-state solution of Eq. (1). b) **At monsoon onset**: Ascent occurs off the equator where absolute vorticity is non-negligible. This results in a negative upper level vorticity tendency due to vortex stretching in the Northern Hemisphere, reducing the magnitude of the absolute vorticity. The circulation begins to transition to the thermally direct steady-state solution of Eq. (1), and the overturning strengthens. The cross-equatorial meridional flow is associated with a positive vorticity tendency due to horizontal advection, which reduces absolute vorticity magnitude in the Southern Hemisphere. The vorticity tendencies due to vortex stretching and horizontal advection therefore both act to further reduce the magnitude of absolute vorticity. This positive feedback allows overturning strength, and the associated precipitation, to increase rapidly over onset.

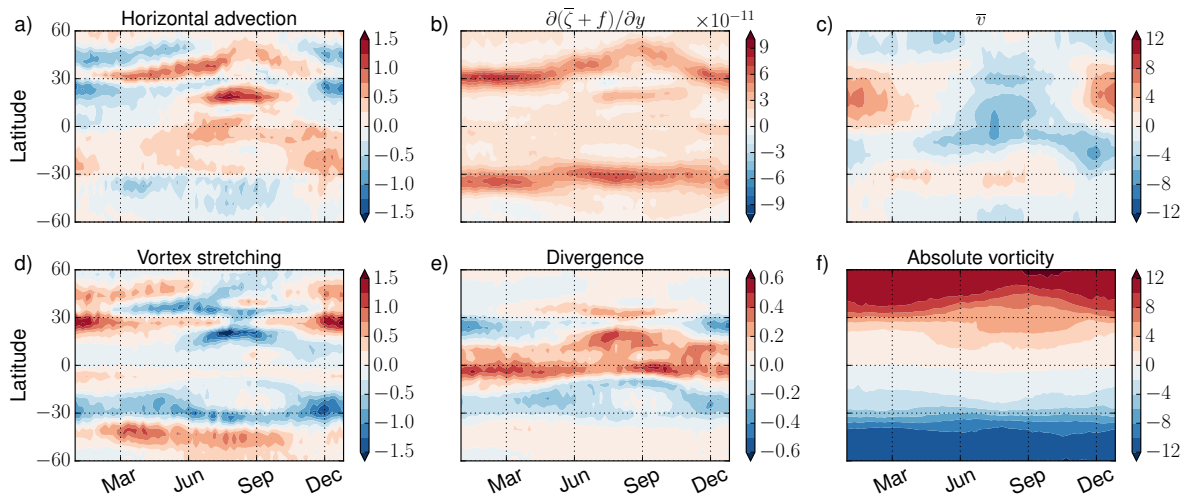


FIG. 8. As Fig. 6 but for *full*, with zonal means here defined between 60 and 150°E.

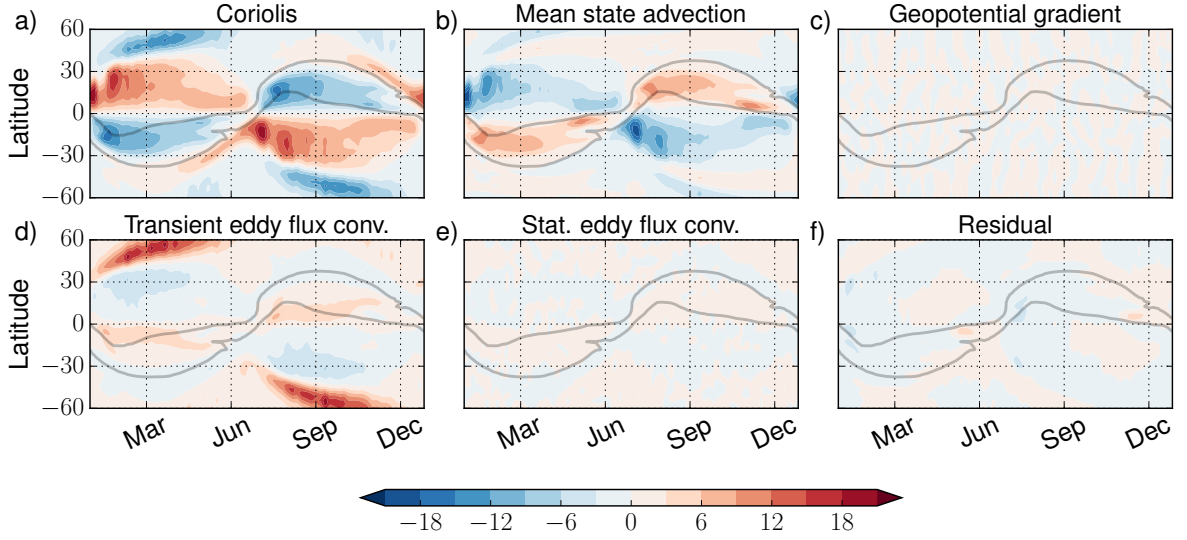


FIG. 9. Seasonal cycle of the terms in the zonal momentum budget Eq. (5) for *ap2* at 150 hPa: (a) $f[\bar{v}]$,
 (b) $-\left(\bar{u}\frac{\partial\bar{u}}{\partial x} + \bar{v}\frac{\partial\bar{u}}{\partial y} + \bar{\omega}\frac{\partial\bar{u}}{\partial p}\right)$, (c) $-\frac{\partial\bar{\Phi}}{\partial x}$, (d) $-\left[\bar{u}^*\frac{\partial\bar{u}^*}{\partial x} + \bar{v}^*\frac{\partial\bar{u}^*}{\partial y} + \bar{\omega}^*\frac{\partial\bar{u}^*}{\partial p}\right]$, (e) $-\left[\frac{\partial\bar{u}'\bar{u}'}{\partial x} + \frac{\partial\bar{u}'\bar{v}'}{\partial y} + \frac{\partial\bar{u}'\bar{\omega}'}{\partial p}\right]$, (f)
 $\partial\bar{u}/\partial t$. All quantities are multi-year pentad and zonal averages. Units are $\text{ms}^{-1}\text{day}^{-1}$. The 8 mm/day precipita-
 tion contour is marked on all plots in gray as an indicator of monsoon onset. Absolute vorticity is overplotted in
 black on (b), with a contour interval of 2 day^{-1} .

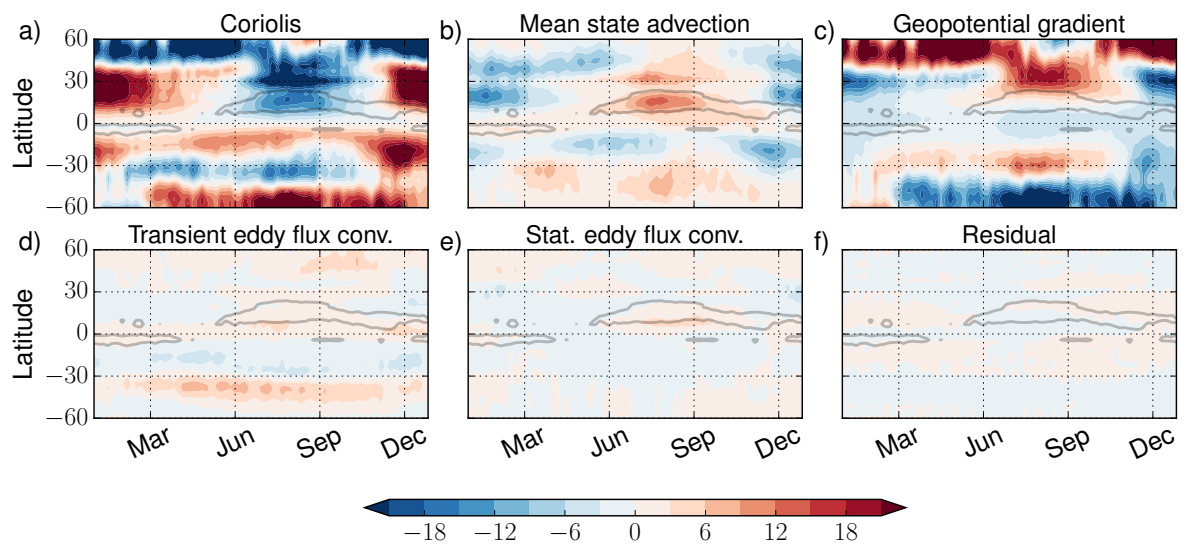


FIG. 10. As Fig. 9 but for *full*, with the zonal average taken between 60 and 150°E.

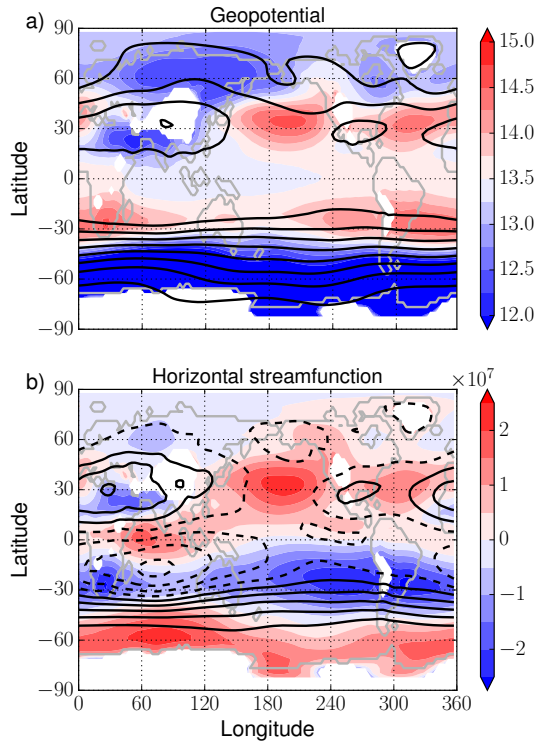


FIG. 11. (a) 850 hPa geopotential, contour interval $250 \text{ m}^2\text{s}^{-2}$, (colors) and 150 hPa geopotential, contour interval $2000 \text{ m}^2\text{s}^{-2}$, (contours) for the *full* experiment, averaged over 20 days post monsoon onset. (b) Equivalent plots for the horizontal streamfunction, contour intervals are $5 \times 10^6 \text{ m}^2\text{s}^{-1}$ at 850 hPa, and $2 \times 10^7 \text{ m}^2\text{s}^{-1}$ at 150 hPa. Dashed lines indicate negative values. The land mask used is shown in gray. Missing values at 850 hPa are due to intersection with topography.