

1 **The Dynamics of the Global Monsoon: Connecting**
2 **Theory and Observations**

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8 **Key Points:**

- 9 • Theoretical understanding of the dynamics of Hadley cells, monsoons and ITCZs
10 is developing rapidly
- 11 • Connecting theory to the observed monsoons helps in understanding variability
12 and future projections
- 13 • Key limitations include the lack of firm theoretical frameworks for zonal asym-
14 metries and transients

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Abstract

Earth's monsoons are complex systems, governed by both large-scale constraints on the atmospheric general circulation and regional interactions with continents and orography, and coupled to the ocean. Monsoons have historically been considered as distinct regional systems, and the prevailing view has been, and remains, an intuitive picture of monsoons as a form of large-scale sea breeze, driven by land-sea contrast. However, climate dynamics is seldom intuitive. More recently, a perspective has emerged within the observational and Earth system modeling communities of a global monsoon that is the result of a seasonally migrating tropical convergence zone, intimately connected to the global tropical atmospheric overturning and localized by regional characteristics. Parallel with this, over the past decade, much theoretical progress has been made in understanding the fundamental dynamics of the seasonal Hadley cells and Intertropical Convergence Zones via the use of hierarchical modeling approaches, including highly idealized simulations such as aquaplanets. Here we review the theoretical progress made, and explore the extent to which these theoretical advances can help synthesize theory with observations and understand differing characteristics of regional monsoons. We show that this theoretical work provides strong support for the migrating convergence zone picture, allows constraints on the circulation to be identified via the momentum and energy budgets, and lays out a framework to assess variability and possible future changes to the monsoon. Limitations of current theories are discussed, including the need for a better understanding of the influence of zonal asymmetries and transients on the large-scale tropical circulation.

Plain Language Summary

The monsoons are the wet summer circulations that provide most of the annual rainfall to many countries in the tropics, influencing over one third of the world's population. Monsoons in different regions have historically been viewed as separate continent-scale 'sea breezes', where land heats faster than ocean in the summer, causing warm air to rise over the continent and moist air to be drawn over the land from the ocean. Recent theoretical advances and the latest analyses of monsoon circulations support a rather different picture. We show that both theories and observations support a view of the monsoons as being localized seasonal migrations of the tropical convergence zone: the band of converging air and rainfall in the tropics embedded within the tropical atmospheric overturning circulation. This updated perspective explains commonalities and differences in behaviour between the regional monsoons, and may help to better understand year-to-year variability in the monsoons, and how the global monsoon might change in future. We end by discussing features of the monsoons that are not yet fully included in this new picture, such as the influence of mountains and continent shapes on the circulation, and the relationship of the monsoons with shorter lived weather systems.

1 Introduction

Monsoons are a dominant feature of the tropical and subtropical climate in many regions of the world, characterized by rainy summer and drier winter seasons, and accompanied by a seasonal reversal of the prevailing winds. Fig. 1a shows the difference in precipitation (GPCP; Huffman et al., 2001) and 850-hPa wind velocity (JRA-55; Kobayashi et al., 2015) between June-September and December-March, based on a climatology from 1979-2016. Strong seasonal differences can be seen throughout the tropics and subtropics, localized by the position of the continents. The magenta contour marks regions where local summer minus winter precipitation exceeds 2.5 mm/day and summer accounts for at least 55% of the annual total precipitation and thus identifies the various monsoon regions around the globe.

For practical purposes, such as agriculture, it has generally been of interest to explore the controls on seasonal rainfall at a regional scale. On the interannual to intrasea-

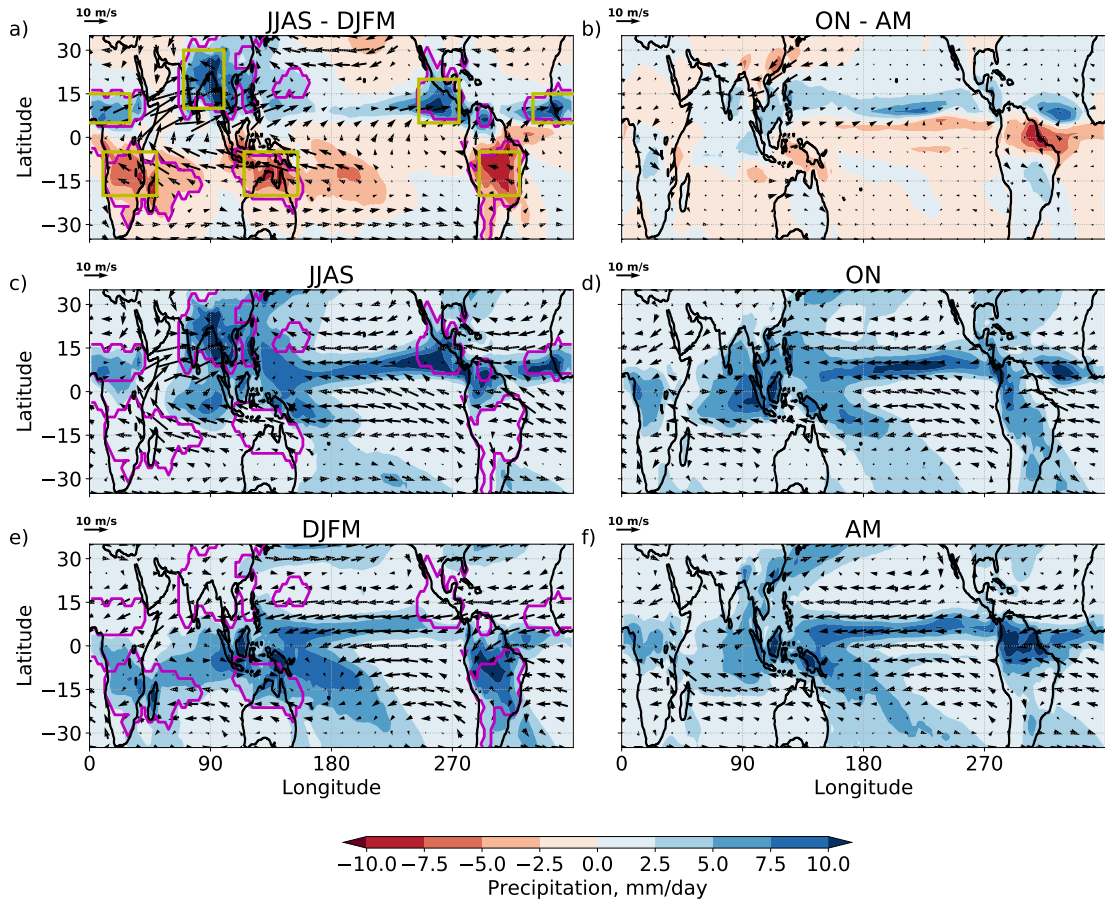


Figure 1. (a) Difference in precipitation (colors) and 850-hPa wind speed (arrows) between Northern Hemisphere summer (defined as June-September) and Southern Hemisphere summer (defined as December-March). (c) and (e) show Northern and Southern Hemisphere summer precipitation and wind respectively. In these panels the magenta contour indicates regions where local summer minus winter precipitation exceeds 2.5mm/day and summer accounts for at least 55% of the annual total precipitation. Yellow boxes in (a) approximate these regions for use in Fig. 2. (b), (d) & (f) are as (a), (c) & (e) but for shoulder seasons defined as October & November and April & May.

65 sonal timeframe, the local monsoons behave as distinct systems, with interannual vari-
 66 ability in precipitation showing limited similarity between regions (Fig. 2). However, as
 67 paleoclimate proxy, observational and reanalysis datasets have become more compre-
 68 hensive and reliable, it has become possible to investigate variability on longer timescales;
 69 a comprehensive review is given in P. X. Wang et al. (2014). These datasets reveal that,
 70 on orbital timescales, the regional monsoons respond coherently to Milankovitch cycles,
 71 on millennial timescales they show similar responses to high latitude forcings, and on in-
 72 terdecadal timescales they also share common trends and modes of variability (e.g. Bi-
 73 asutti et al., 2018; J. C. Chiang & Friedman, 2012; Schneider, Bischoff, & Haug, 2014;
 74 B. Wang & Ding, 2006; P. X. Wang et al., 2014). For example, Fig. 3 shows data from
 75 a range of paleoclimate proxies, which indicate that tropical rainfall in South America,
 76 Indonesia and India covaries with Arctic temperatures on millennial timescales.

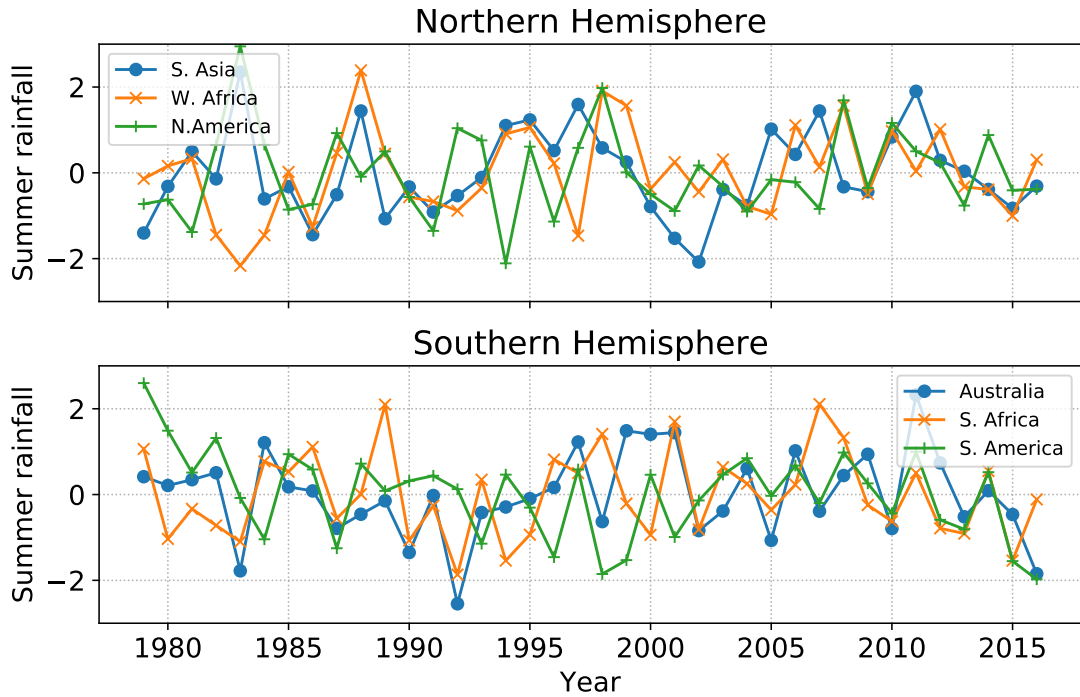


Figure 2. Timeseries of summer-time (July-September mean in the Northern Hemisphere and December-March mean in the Southern Hemisphere) rainfall over the yellow boxes marked in Fig. 1, which are used as approximations to the monsoon regions defined by the magenta contour. For ease of comparison, the timeseries are standardized by subtracting the mean and dividing by the standard deviation. Data are taken from the Global Precipitation Climatology Project (GPCP; Huffman et al., 2001) over 1979-2016.

77 This evidence for coherent global-scale monsoons raises questions about our phys-
 78 ical understanding of the systems. Historically, the localization of summertime tropical
 79 rainfall around land led to the intuitive interpretation of monsoons as a large-scale sea
 80 breeze, with moist air drawn over the continent in the local summer season, when the
 81 land is warm relative to the ocean, resulting in convective rainfall over land (Halley, 1686).
 82 Monsoons were considered distinct phenomena to the Intertropical Convergence Zone
 83 (ITCZ), which is generally defined as the convergence of the trade winds of the North-
 84 ern and Southern Hemispheres and is embedded within the ascending branches of the
 85 Hadley circulations. This perspective of monsoons as a sea breeze has been pervasive,
 86 despite the fact that land-sea temperature contrast has long been known to be greatest
 87 prior to monsoon onset over India (Simpson, 1921), and that drought years are accom-
 88 panied by higher land surface temperatures (Kothawale & Kumar, 2002). More recent
 89 work suggests a perspective of the monsoons as localized and more extreme migrations
 90 of the tropical convergence zone, which may sit near the Equator forming an ITCZ, or
 91 be pulled poleward over the continent as a monsoon (see Gadgil, 2018, and references
 92 therein).

93 Simultaneously, a significant body of work investigating the fundamental dynam-
 94 ics of the monsoon has been undertaken via hierarchical modeling approaches, ranging
 95 from dry axisymmetric models (e.g. Bordoni & Schneider, 2010; Hill, Bordoni, & Mitchell,
 96 2019; Schneider & Bordoni, 2008), to cloudless moist models (e.g. Bordoni & Schneider,
 97 2008; Faulk, Mitchell, & Bordoni, 2017; Geen, Lambert, & Vallis, 2018, 2019; Privé &
 98 Plumb, 2007a), to more comprehensive models including full physics and realistic orog-

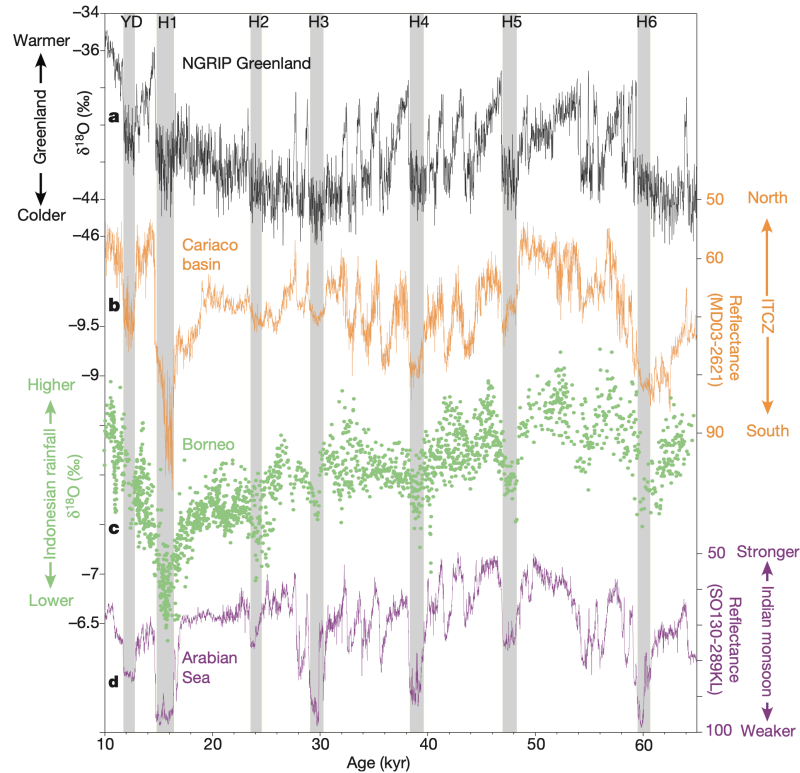


Figure 3. Paleoclimate proxy records from over the past 64,000 years. (a) Record of $\delta^{18}\text{O}$ from the North Greenland Ice Core Project, a proxy for Arctic temperatures (Bond et al., 1993; Wolff et al., 2010). (b) Reflectance of Cariaco basin sediments, which is high when rainfall and runoff are low, and is used here as a proxy for the latitude of the ITCZ in the western tropical Atlantic (Deplazes et al., 2013). (c) $\delta^{18}\text{O}$ in cave stalagmites from Borneo (Carolin et al., 2013), a proxy for rainfall in the equatorial western Pacific. (d) Reflectance of Arabian Sea sediments (Deplazes et al., 2013); low reflectance indicates high runoff from Indian monsoon rainfall. Figure taken from Schneider et al. (2014).

99 raphy (e.g. Boos & Kuang, 2010; Chen & Bordoni, 2014). This hierarchy has allowed a
 100 wide range of factors controlling the structure of tropical precipitation to be explored.
 101 Findings from these studies strongly support the view of monsoons as local expressions
 102 of the global tropical convergence zone, and provide valuable, theoretically grounded in-
 103 sights into the controls on the tropical circulation.

104 In this review, we attempt to synthesize the results of studies on the observed char-
 105 acteristics of Earth’s monsoon systems with recent theoretical analyses that provide con-
 106 straints on the large-scale dynamics of monsoons, with the aim of taking stock of the progress
 107 achieved and identifying avenues for future work. Note that throughout the review, ‘mon-
 108 soon’ refers to the local summer, as opposed to winter, monsoon. Specifically, as we will
 109 motivate through discussion of theoretical work, we reserve the term ‘monsoon’ to de-
 110 scribe precipitation associated with cross-equatorial overturning circulations with ascend-
 111 ing branches located poleward of $\sim 10^\circ$ latitude. We will show that, unlike the ITCZs,
 112 these circulations are not strongly influenced by momentum transport by large-scale ex-
 113 tratropical eddies. The term ‘ITCZ’ is instead reserved to describe the zonally oriented
 114 precipitation bands that remain within $\sim 10^\circ$ of the Equator and whose dynamics are
 115 much more strongly influenced by large-scale eddies. The term *convergence zone* will be

116 used to refer to both monsoonal and ITCZ precipitation because, regardless of their gov-
 117 erning dynamics, both are associated with ascending branches of overturning cells.

118 The goals of this article are:

- 119 1. To assess the relevance of idealized modeling results to the real-world monsoons;
- 120 2. To help to motivate relevant simulations from the modeling community to answer
 121 open questions on monsoon dynamics;
- 122 3. To provide an introduction to both of these aspects for readers new to the field.

123 With these aims in mind, Section 2 discusses theoretical results derived from ide-
 124 alized models, particularly aquaplanets with symmetric boundary conditions and heat-
 125 ing perturbations. Section 3 discusses the features of the observed regional monsoons,
 126 their combined role in the global monsoon, and the applicability of the dynamical pro-
 127 cesses identified in idealized models to the various systems. Section 4 explores the roles
 128 of asymmetries in the boundary conditions and transient activity in the monsoons and
 129 ITCZs. These factors are sometimes overlooked in formulating theories in idealized mod-
 130 els. In Section 5 we summarize the successes and limitations of this synthesis of theory
 131 and observations, and propose some areas on which to focus future research.

132 **2 Idealized modeling of monsoons**

133 Reanalyses, observations, and state of the art global circulation models (GCMs)
 134 give our best estimates of Earth’s climate. However, when viewed as a whole, the Earth
 135 system is dizzyingly complex, and identifying the processes controlling the various el-
 136 ements of climate is hugely challenging. Idealized models provide a valuable tool for break-
 137 ing down some of this complexity, and for proposing mechanisms whose relevance can
 138 then be investigated in more realistic contexts.² In this section, we review the use of ide-
 139 alized models in understanding the dynamics of the monsoons.

140 Some key insights into the controls on tropical rainfall and monsoons have come
 141 from a perhaps unexpected source: aquaplanets. Despite lacking zonal asymmetries such
 142 as land-sea contrast, which localize regional monsoons, these models have been shown
 143 to capture the basic elements of a monsoon. For example, in aquaplanets with moist physics
 144 and a low thermal inertia slab ocean, the ITCZ migrates rapidly away from the Equa-
 145 tor into the summer hemisphere during the warm season (Bordoni & Schneider, 2008).
 146 This migration of the ITCZ is associated with a rapid reversal of the upper- and lower-
 147 level wind in the summer hemisphere, and the onset of intense off-equatorial precipita-
 148 tion, similar to the behaviors seen in Earth’s monsoons, (e.g. Fig. 4). Thus, in so far as
 149 the rapid development of an off-equatorial convergence zone accompanied by similarly
 150 rapid circulation changes can be interpreted as a monsoon, aquaplanets provide a sim-
 151 ple tool for exploring the lowest-order processes at work. This represents a significant
 152 change in perspective from the classical view of monsoon wind reversal as driven by land-
 153 sea thermal contrast (Halley, 1686), towards a view of monsoons as local and seasonal
 154 manifestations of the Hadley circulation.

155 Different theoretical approaches have been used to interpret the results from these
 156 idealized simulations, primarily using large-scale budgets of energy and angular momen-
 157 tum. The momentum budget gives insight into the drivers and regimes of the overturn-
 158 ing circulation, and how these relate to monsoon onset. The energy budget provides a

¹ NB Mixed layer depths here are corrected from Bordoni and Schneider (2008), (S. Bordoni, pers. com., 2020).

² For further discussion of the use of idealized models and the model hierarchy see (Held, 2005; Jeevanjee, Hassanzadeh, Hill, & Sheshadri, 2017; Maher et al., 2019)

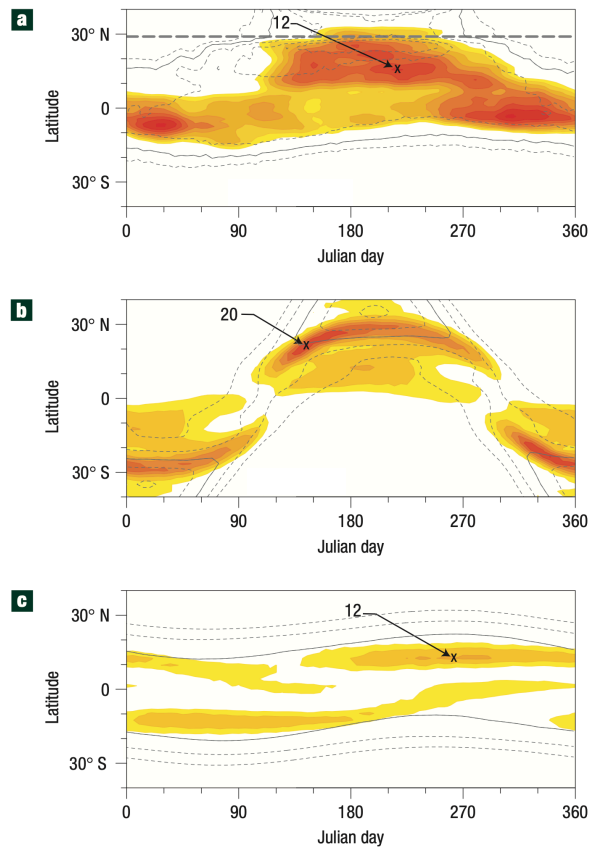


Figure 4. Seasonal cycle of zonal- and pentad- mean precipitation (color contours, data from GPCP 1999-2005) and sea-level air temperature (gray contours, data from the ERA-40 reanalysis (Uppala et al., 2005)) for (a) observations in the Asian monsoon sector (70-100°E), and for aquaplanet simulations with ocean mixed-layer heat capacity equivalent to (b) 0.5m and (c) 50m of water.¹ The precipitation contour interval is 1mm/day in (a) and 2mm/day in (b) and (c), and maxima are indicated by crosses. For sea-level air temperature, the contour interval is 2°C in all panels, and the solid gray line indicates the 24°C isotherm. The thick dashed line in (a) shows the latitude at which the zonal-mean topography in the Asian monsoon sector rises above 3 km. Figure taken from Bordoni and Schneider (2008).

159 framework for understanding the controls on the latitude of the zonally averaged ITCZ,
 160 and its meridional migration. In a real world context, this is useful in interpreting the
 161 latitude of tropical rainfall bands, and the meridional extent of Earth's monsoons. These
 162 complementary approaches are discussed in Sections 2.1 and 2.2, respectively.

163 2.1 Dynamical constraints

164 One important constraint on the atmospheric circulation is conservation of angular
 165 momentum. Recent results from aquaplanet simulations suggest that this can help
 166 to explain controls on the latitude of the convergence zone, the extent of the Hadley cir-
 167 culation, and the rapidity of monsoon onset.

168 The axial component of the angular momentum associated with the atmospheric
 169 circulation is

$$M = \Omega a^2 \cos^2 \phi + ua \cos \phi, \quad (1)$$

170 where Ω and a are Earth's rotation rate and radius, u is the zonal wind speed, and ϕ is
 171 latitude. Equation 1 states that the atmosphere's angular momentum comprises a plan-
 172 etary contribution from Earth's rotation, and a contribution from the zonal wind rela-
 173 tive to this. In the absence of torques, e.g. from friction, zonal pressure gradients or orog-
 174 raphy (see Egger, Weickmann, & Hoinka, 2007), M is conserved by an air parcel as it
 175 moves meridionally. Above orography, in the zonal mean we can approximate

$$\frac{DM}{Dt} = 0. \quad (2)$$

176 Substituting Eq. 1 into Eq. 2, linearising about the zonal and time mean state, and con-
 177 sidering upper-level flow where viscous damping is weak and can be neglected gives

$$\bar{v} \left(f - \frac{1}{a \cos \phi} \frac{\partial(\bar{u} \cos \phi)}{\partial \phi} \right) - \bar{\omega} \frac{\partial \bar{u}}{\partial p} = \frac{1}{a \cos^2 \phi} \frac{\partial(\overline{u'v'} \cos^2 \phi)}{\partial \phi} + \frac{\partial \overline{u'\omega'}}{\partial p}. \quad (3)$$

178 Overbars indicate the time and zonal mean, and primes deviations from the time mean.³
 179 v and ω are the meridional and vertical wind components, respectively. Terms relating
 180 to the mean flow have been grouped on the left hand side, while terms relating to eddy
 181 fluxes of momentum are grouped on the right. Assuming that meridional streamlines are
 182 near horizontal in the upper branch of the Hadley circulation, so that the vertical deriva-
 183 tives can be neglected, and utilising the definition of relative vorticity, $\zeta = \hat{\mathbf{k}} \cdot \nabla \times \mathbf{u}$, Eq.
 184 3 can be expressed in terms of a local Rossby number, $Ro = -\bar{\zeta}/f$, (cf. Schneider & Bor-
 185 doni, 2008) as

$$f(1 - Ro)\bar{v} = \frac{1}{a \cos^2 \phi} \frac{\partial(\overline{u'v'} \cos^2 \phi)}{\partial \phi}. \quad (4)$$

186 2.1.1 The axisymmetric case

187 Considering first the case of an axisymmetric atmosphere, in which there are no
 188 eddies, Eq. 4 has two classes of solution. Firstly, the zonal averaged meridional and (by
 189 continuity) vertical velocities may be zero everywhere. This corresponds to a radiative-
 190 convective equilibrium (RCE) solution. Alternatively, Ro may be equal to 1 and an ax-
 191 isymmetric circulation may exist, so that the zonal and time mean flow conserves an-
 192 gular momentum. Plumb and Hou (1992) and Emanuel (1995) explored the conditions
 193 under which either of these cases might occur in dry and moist atmospheres, respectively.
 194 Importantly, the RCE solution is not viable if the resulting zonal wind in thermal wind
 195 balance with the RCE temperatures violates Hide's theorem (Hide, 1969) by giving rise
 196 to a local extremum in angular momentum. Plumb and Hou (1992) demonstrate that
 197 for an off-equatorial forcing, this implies the existence of a threshold curvature of the lower-
 198 level RCE temperature, above which the RCE solution cannot exist and an overturn-
 199 ing circulation will develop. They also speculate that this threshold behavior in the ax-
 200 isymmetric model might be related to the rapid onset of Earth's monsoons. The over-
 201 all argument is as follows.

202 Taking the RCE case, in which v and ω vanish, gradient wind and hydrostatic bal-
 203 ance can be expressed in pressure coordinates as

$$\frac{\partial}{\partial p} \left[f u_e + \frac{u_e^2 \tan \phi}{a} \right] = \frac{1}{a} \left(\frac{\partial \alpha}{\partial \phi} \right)_p, \quad (5)$$

204 where α is specific volume and u_e is a RCE zonal wind profile. For a dry atmosphere,
 205 assuming the zonal wind speed at the surface is zero, the above can be integrated down

³ Since we are in an aquaplanet framework, there are no eddy transports by stationary waves.

206 to the surface for a given upper-level wind profile to give an associated RCE lower-level
 207 temperature distribution (cf. Lindzen & Hou, 1988; Plumb & Hou, 1992).

208 In modeling Earth’s atmosphere, moist processes must also be accounted for. In
 209 the tropics, frequent, intense moist convection means that in the time mean, the lapse
 210 rate is approximately maintained as a moist adiabat, so that the saturation moist en-
 211 tropy of the free atmosphere is nearly equal to the subcloud moist entropy, s_b (the b de-
 212 noting subcloud values) (e.g. Arakawa & Schubert, 1974; Emanuel, Neelin, & Brether-
 213 ton, 1994). This is known as *convective quasi-equilibrium* (CQE). Assuming the tropi-
 214 cal atmosphere to be in CQE, Emanuel (1995) uses Eq. 5 to derive a relation between
 215 the angular momentum at the tropopause, M_t , and subcloud equivalent potential tem-
 216 perature, θ_{eb} :

$$c_p(T_s - T_t) \frac{\partial \ln \theta_{eb}}{\partial \phi} = -\frac{1}{a^2} \frac{\tan \phi}{\cos^2 \phi} (M_t - \Omega^2 a^4 \cos^4 \phi), \quad (6)$$

217 where T_s and T_t are the RCE temperatures at the surface and tropopause respectively,
 218 c_p is the heat capacity of dry air at constant pressure and θ_{eb} is related to moist entropy
 219 as $s_b = c_p \ln \theta_{eb}$. The condition that no local maximum in angular momentum exist gives
 220 a critical curvature of θ_{eb} :

$$-\left[\frac{\partial}{\partial \phi} \left(\frac{\cos^2 \phi}{\tan \phi} c_p (T_s - T_t) \frac{\partial \ln \theta_{eb}}{\partial \phi} \right) \right]_{crit} = 4\Omega^2 a^2 \cos^3 \phi \sin \phi. \quad (7)$$

221 In an axisymmetric atmosphere, if the left hand side of Eq. 7 is less than than the right
 222 hand side, the RCE solution is viable and there is no meridional overturning cell. Where
 223 this condition is violated, so that the profile of θ_{eb} is supercritical, the RCE solution is
 224 not viable and a meridional flow must exist (cf. Emanuel, 1995; Hill et al., 2019; Plumb
 225 & Hou, 1992). This condition is illustrated graphically in Fig. 5, which shows the pro-
 226 files of RCE zonal wind, angular momentum, and absolute vorticity (proportional to the
 227 meridional gradient of angular momentum) that result from a range of forcings with a
 228 subtropical maximum and flat elsewhere, Fig. 5a.⁴ For weak forcing (blue lines), no ex-
 229 trema of M are produced, illustrated by the fact that absolute vorticity (Fig. 5d) is pos-
 230 itive everywhere. At the critical forcing profile (gray lines) a saddle point in M is pro-
 231 duced, where absolute vorticity is 0. Beyond this point, the profiles of u that are in gra-
 232 dient wind balance with the forcing are such as to produce extrema in M , and are in vi-
 233 olation of Hide’s theorem (Hide, 1969) so that a Hadley circulation must develop.

234 The above arguments assess the conditions under which a Hadley circulation will
 235 exist in an axisymmetric atmosphere. Privé and Plumb (2007a) further showed that this
 236 framework can give some insight into the controls on the latitude of the ITCZ. They noted
 237 that, for an overturning circulation that conserves angular momentum in the free tro-
 238 posphere, the circulation boundary for a vertical streamline must be located in a region
 239 of zero vertical wind shear. Where CQE applies, so that free tropospheric temperatures
 240 are coupled to lower-level θ_{eb} , this implies that the zero streamfunction contour must oc-
 241 cur in a region of zero horizontal gradient of θ_{eb} (i.e. where θ_{eb} maximizes). Most of the
 242 ascent in the circulation ascending branch, and consequently the precipitation, will oc-
 243 cur just equatorward of this maximum. They additionally noted that either the max-
 244 imum in θ_{eb} or the maximum in moist static energy (MSE), h , could also be used to es-
 245 timate the latitude of the convergence zone (see their Section 5)⁵, as the two variables
 246 are related by

$$\partial \theta_{eb} \approx \frac{1}{T_b} \partial h_b, \quad (8)$$

⁴ NB. This figure, taken from Hill et al. (2019), corresponds to a dry atmosphere, cf. Plumb and Hou (1992), but the behavior is equivalent to that for Eq. 7.

⁵ θ_e is useful due to its relationship to moist entropy, which for example allows the substitution of a Maxwell relation into Eq. 5 (Emanuel, 1995). However, MSE is a linear quantity that is straightforward to calculate, and so is more widely used.

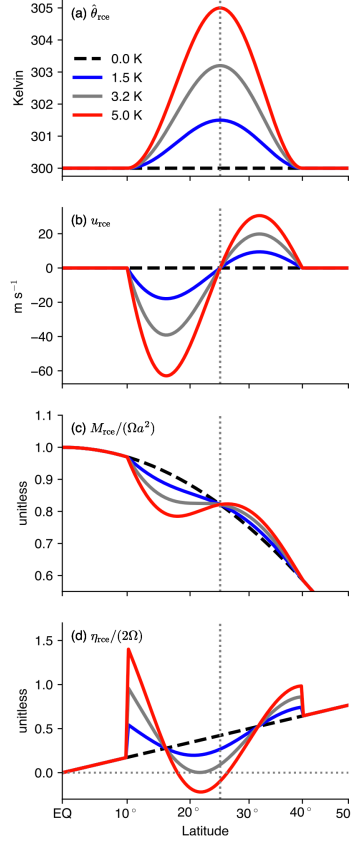


Figure 5. Illustration of the effects of a subcritical (blue lines), critical (gray lines) or supercritical (red lines) RCE potential temperature profile. Forcing profiles, shown in (a), are based on those used by Plumb and Hou (1992). The remaining panels show (b) zonal wind (m s^{-1}), (c) absolute angular momentum, normalized by the planetary angular momentum at the Equator, (d) absolute vorticity, normalized by twice the planetary rotation rate. Figure taken from Hill et al. (2019). ©American Meteorological Society. Used with permission.

$$h = c_p T + L_v q + gz. \quad (9)$$

247 In the above, T is temperature, q is specific humidity, z is height, L_v is the latent heat
 248 of vaporisation of water, and g is gravitational acceleration.

249 **2.1.2 Eddy-permitting solutions**

250 These theoretical considerations, based on conservation of angular momentum, pro-
 251 vide important constraints on the existence and extent of axisymmetric overturning cir-
 252 culations. However, it is now well known that extratropical eddies, generated in midlat-
 253 itude baroclinic zones, propagate into the subtropics where they break, and have non-
 254 negligible impact on the Hadley circulation (e.g. Becker, Schmitz, & Geprägs, 1997; C. C. Walker
 255 & Schneider, 2006). In particular, as transport of angular momentum by large-scale ed-
 256 dies becomes non-negligible, the associated eddy momentum flux convergence in Eq. 4
 257 can no longer be neglected. In the limit of small Ro , the mean meridional advection of
 258 zonal momentum is negligible, and the dominant balance is between the Coriolis effect
 259 on the zonal mean meridional flow and the eddy momentum flux divergence. This regime
 260 is linear, in that the mean advection term is negligible, and eddy driven, in that the strength

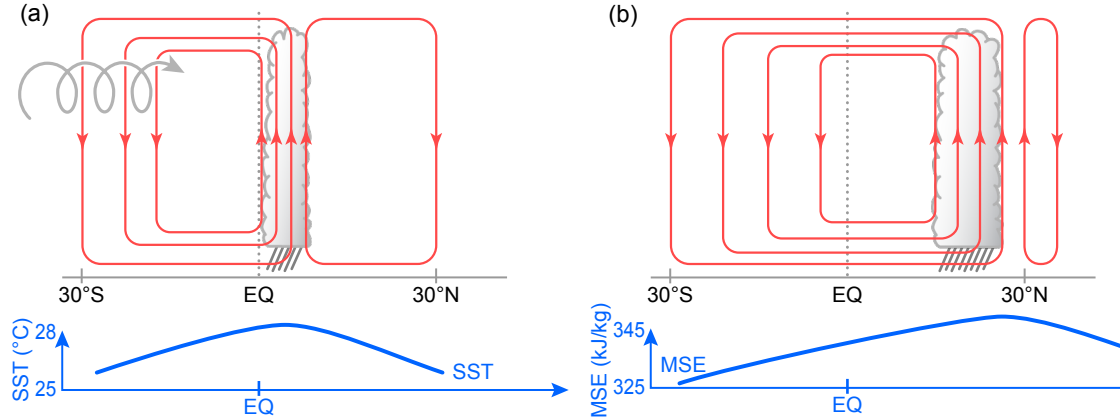


Figure 6. Schematic illustration of the two regimes of the meridional overturning circulation identified in aquaplanets (Bordoni & Schneider, 2008; Schneider & Bordoni, 2008). The gray cloud denotes clouds and precipitation, red contours denote streamfunction. (a) Convergence zone is an ITCZ located near to the Equator, and approximately co-located with the peak SST. Hadley cells are significantly eddy driven, as indicated by the helical arrow. (b) Convergence zone is monsoon-like, located farther from the Equator, with the mid-tropospheric zero contour of the streamfunction aligned with the moist static energy (MSE) maximum (Privé & Plumb, 2007b) and precipitation falling just equatorward of this. The winter Hadley cell crosses the Equator and is near angular-momentum conserving, with eddies only weakly influencing the overturning strength. The summer Hadley cell is comparatively weak, if present at all. Illustration by Beth Tully.

261 of the circulation is constrained by the eddy momentum fluxes. In reality, intermediate
 262 cases, with $Ro \sim 0.5$, are also observed, where both nonlinear zonal mean advection
 263 and eddy terms are important.

264 Transitions from regimes with small Ro to regimes with Ro approaching unity have
 265 been connected to the rapid changes in the tropical circulation that occur during mon-
 266 soon onset. Examining the upper-level momentum budget in aquaplanet simulations with
 267 a seasonal cycle, Bordoni and Schneider (2008) found that, while the Hadley cells in the
 268 two hemispheres are roughly symmetric and the associated convergence zone is near the
 269 Equator, $Ro \lesssim 0.5$, and the circulation strength is governed by eddies (e.g. Fig. 6a). As
 270 the insolation maximum starts moving into the summer hemisphere, the winter Hadley
 271 cell starts becoming cross equatorial. The zonal mean ascent and precipitation move to
 272 an off-equatorial location in the summer hemisphere (e.g. Fig. 4), and upper-level tropi-
 273 cal easterlies develop. These limit the ability of eddies from the winter hemisphere to
 274 propagate into the low latitudes, and the circulation shifts quickly towards the $Ro \sim 1$
 275 angular momentum conserving flow regime, at the same time strengthening and expand-
 276 ing rapidly (e.g. Fig. 6b). As the cross-equatorial circulation approaches conservation
 277 of angular momentum, the dominant balance becomes between the terms on the left hand
 278 side of Eq. 3, with the eddy terms a small residual. Once in this regime, the circulation
 279 is no longer constrained by the zonal momentum budget, which becomes a trivial bal-
 280 ance, but is constrained by the energy budget, and so responds strongly to the thermal
 281 forcing.

282 The rapid meridional migrations of the convergence zone in the aquaplanet are a
 283 result of a positive feedback relating to advection of cooler and drier air up the MSE gra-

284 dent in the lower branch of the winter Hadley cell, illustrated in Fig. 7. As summer be-
 285 gins the summer hemisphere warms via diabatic fluxes of MSE into the air column (Fig.
 286 7b). This pulls the lower-level peak in MSE and, in accordance with the arguments of
 287 Privé and Plumb (2007a), pulls the ITCZ off of the Equator. Simultaneously, the win-
 288 ter Hadley circulation begins to redistribute MSE, advecting cooler and drier air up the
 289 MSE gradient (Figs. 7c&d). The gradient of MSE steepens, amplifying the thermal forc-
 290 ing of the cell. The overturning circulation strengthens in response to this, further in-
 291 creasing the lower-level advection of cooler air, and expanding the upper-level easterlies,
 292 allowing the circulation to become further shielded from the eddies. It is important to
 293 note that in this view land is necessary for monsoon development only insofar as it pro-
 294 vides a lower boundary with low enough thermal inertia for the MSE to adjust rapidly
 295 and allowing the feedbacks described above to act on intraseasonal timescales. Behav-
 296 ior consistent with these feedbacks has been observed in Earth’s monsoons, and will be
 297 discussed in more detail in Section 3.

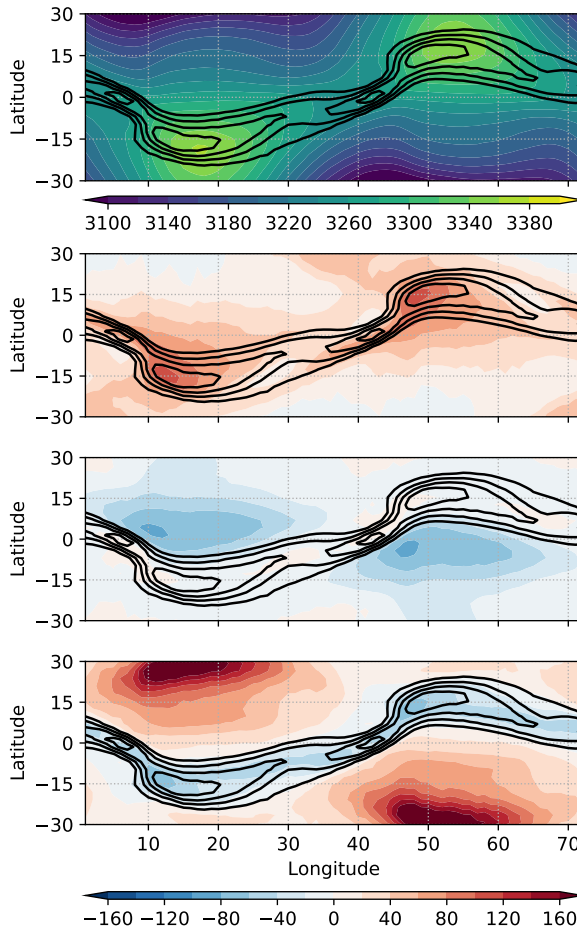


Figure 7. Zonal- and pentad- mean of terms in the moist static energy budget for an aqua-planet simulation with surface heat capacity equivalent to 10m of water. Panel (a) shows the vertically integrated moist static energy, h (units MJm^{-2}). Panels (b-d) show F_{net} (the net flux of radiation, latent and sensible heat into the column), $-v\partial h/\partial y$, and $-\omega\partial h/\partial p$ respectively (units Wm^{-2}).

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2.1.3 Hadley cell regimes and the latitude of the ITCZ

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The idealized modeling work discussed above indicates that the Hadley cells in an aquaplanet change their circulation regime over the course of the year, varying rapidly between an eddy-driven ‘ITCZ’ regime and a near angular-momentum conserving ‘monsoon’ regime. In addition, that the cross-equatorial Hadley cell approaches angular momentum conservation suggests that axisymmetric theories (e.g. Equation 7) might not be applicable to the understanding of the zonal and annual mean Hadley cell, but might provide important constraints on monsoonal circulations, which do approach an angular momentum conserving state. The relationship between these two regimes and the latitude of the ITCZ raises further questions: How far into the summer hemisphere must the Hadley cell extend for the regime transition, and associated rapid shift in ITCZ latitude, to occur? Does the latitude at which the convergence zone shifts from being governed by ‘ITCZ’ to ‘monsoon’ dynamics in aquaplanets relate to the observed latitudes of the ITCZs and monsoons? If the Hadley cell boundary follows the peak in MSE (Privé & Plumb, 2007a), what governs the extent of the cross-equatorial cell, e.g. is a pole-to-pole cell possible?

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Geen et al. (2019) investigate the first of the above questions. By running aquaplanet simulations under a wide range of conditions, including different slab ocean depths, year lengths, and rotation rates, they investigated how the ITCZ latitude and migration rate were related, and how these factors varied over the year. They found that, at Earth’s rotation rate, the ITCZ appeared least stable (migrated poleward fastest) at a latitude of 7° , suggesting that this may be the poleward limit of an eddy-driven overturning circulation fuelled by an ITCZ. Beyond this latitude there is a rapid transition to a monsoon circulation characterized by an overturning circulation with a rising branch far off the Equator and weak eddy momentum transports. This ‘transition latitude’ does not vary significantly with surface heat capacity or year length, but it does increase with decreasing planetary rotation rate. Although the mechanism governing the transition latitude is not yet fully understood, this 7° threshold might give a guideline for where the tropical precipitation is dynamically associated with a near-equatorial ‘ITCZ’ vs. a monsoon system.

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Consistent with these results, simulations with zonally symmetric continents extending poleward from southern boundaries with different latitudes show that monsoonal convergence zones extending into the subtropics and featuring a rapid onset can only develop over a continent extending to tropical latitudes (up to 20°). For continents with more poleward southern boundaries, the main precipitation zone remains close to the Equator and moves more gradually into the summer hemisphere. The absence of regions of low thermal inertia in the tropical region in this second case prevents the establishment of a reversed meridional MSE gradient and, with it, the poleward displacement of the circulation ascending branch and convergence zone (Hui & Bordoni, in prep.).

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While the results from this idealized work can help differentiate between oceanic ITCZs vs land monsoons, it is worth noting that the observed ITCZs over the Atlantic and Pacific are in fact located away from the Equator, and migrate between 4° and 12° over the year (see Figs. 1 and 8). There is considerable evidence that the off-equatorial position of these ITCZs is a result of a symmetric instability in the boundary layer flow (Levy & Battisti, 1995; Stevens, 1983; Tomas & Webster, 1997). Symmetric instability is a two-dimensional (latitude-height) instability that results from the joint criteria of conservation of angular momentum and potential temperature (potential vorticity).⁶ The instability in the boundary layer flow is set up by cross-equatorial pressure gradients, driven by equatorially asymmetric boundary layer heating. This may result from either a zon-

⁶ For motion on a constant potential temperature (angular momentum) surface, the criteria reduces to the criteria for inertial (convective) instability (Emanuel, 1988; Tomas & Webster, 1997).

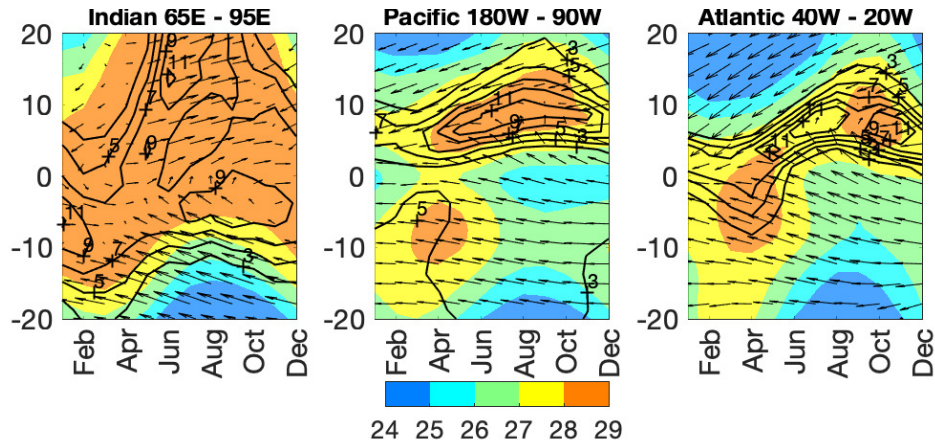


Figure 8. Hovmöller diagram of the climatological SST and 10m wind (1979–2017), averaged across the (left) Indian, (center) eastern half of the Pacific and (right) Atlantic basin. SST is shaded (in °C) and precipitation is contoured (contour interval 2 mm/day). The wind vectors are relative to the maximum in each panel. Precipitation data are from CMAP 1979–2017 (Xie & Arkin, 1997a), SST data are from HADISST 1870–2017 (Rayner et al., 2003), and wind data are from ERA-Interim 1979–2017 (Dee et al., 2011). From Battisti et al. (2019). ©American Meteorological Society. Used with permission.

347 ally asymmetric sea surface temperature (SST) distribution (see Back & Bretherton, 2009;
 348 J. C. H. Chiang, Zebiak, & Cane, 2001; Lindzen & Nigam, 1987, and references therein)
 349 or from asymmetric land heating (e.g., the West African summer monsoon). The result
 350 of the instability is a band of divergence in the boundary layer that lies between the Equa-
 351 tor and the latitude of neutral stability, flanked by a narrow zone of convergence that
 352 lies just poleward and provides the moisture convergence that fuels the ITCZ convec-
 353 tion. The role of symmetric instability was not considered by Geen et al. (2019), who
 354 ran only simulations where the annual mean ITCZ location was on the Equator. Fur-
 355 ther aquaplanet simulations, with for example ocean heat transport prescribed such as
 356 to set up an annual mean asymmetric boundary layer pressure gradient, may help in un-
 357 derstanding how these various instabilities and feedbacks interact during the seasonal
 358 transitions of the Hadley cells.

359 The application of the theoretical concepts discussed in Sections 2.1.1 and 2.1.2 to
 360 Hadley cell extent has been addressed in recent work by Faulk et al. (2017), Hilgenbrink
 361 and Hartmann (2018), Hill et al. (2019) and Singh (2019). Faulk et al. (2017) performed
 362 a series of simulations using an eddy-permitting aquaplanet model in which they varied
 363 rotation rate under seasonally varying insolation. They found that, at Earth’s rota-
 364 tion, the MSE maximized at the summer pole, but the ITCZ did not migrate poleward
 365 of $\sim 25^\circ$ from the Equator even in perpetual solstice simulations, contrary to expecta-
 366 tions from Privé and Plumb (2007a). The influence of eddies on the cross-equatorial cir-
 367 culation was found to be weak, consistent with the suppression of eddies by upper-level
 368 easterlies (Bordoni & Schneider, 2008; Schneider & Bordoni, 2008) and justifying the use
 369 of axisymmetric based considerations as a starting point for understanding the cell ex-
 370 tent. Faulk et al. (2017) found that a Hadley circulation existed over the latitudes where
 371 the curvature of θ_{eb} was supercritical (see Eq. 7), with the curvature subcritical in the
 372 extratropics.

373 Other authors have used axisymmetric theory to examine what determines the max-
 374 imum Hadley cell extent. For example, Hill et al. (2019) applies axisymmetric arguments

375 to show that an angular momentum conserving cell must extend at least as far into the
 376 winter hemisphere as the summer hemisphere. In eddy-permitting atmospheres, the Hadley
 377 cell edge is often viewed as related to the latitude at which the upper-level wind shear
 378 becomes baroclinically unstable (e.g. Held & Coauthors, 2000). Hilgenbrink and Hart-
 379 mann (2018) found that the reduced upper-level zonal velocity associated with an ITCZ
 380 located further from the Equator tends to reduce vertical wind shear, and consequently
 381 baroclinicity, over the winter hemisphere cell edge, so that the cell extends deeper into
 382 the winter hemisphere.

383 While these studies have provided novel insight into important features of cross-
 384 equatorial Hadley cells, prognostic theories for their poleward boundary in the summer
 385 hemisphere have yet to emerge. Singh (2019) investigated the limitations of CQE-based
 386 predictions based on the lower-level MSE maximum. The vertical instability addressed
 387 by CQE is not the only form of convective instability in the atmosphere. If vertical wind
 388 shear is strong, CQE predicts an unstable state in which potential energy is released when
 389 saturated parcels move along slantwise paths, along angular momentum surfaces (Emanuel,
 390 1983a, 1983b). Singh (2019) showed that the perpetual solstitial cell extent can be ac-
 391 curately estimated by assuming that the large-scale circulation adjusts the atmosphere
 392 towards a state that is neutral to this slantwise convection. When the peak in subcloud
 393 moist entropy is relatively close to the Equator, the cell boundary is near vertical and
 394 the atmosphere is near CQE, and this reduces to the condition of Privé and Plumb (2007a).

395 Notably, this developing body of literature indicates that the planetary rotation
 396 rate determines the latitudinal extent of the Hadley cell, potentially limiting the max-
 397 imum latitudinal extent of a monsoon circulation. This might provide a guideline for dis-
 398 tinguishing a monsoon associated with a cross-equatorial Hadley cell and governed by
 399 tropical dynamics, where the convergence zone is located with $\sim 25^\circ$ of the Equator (e.g.
 400 South Asia) from a monsoon strongly influenced by extratropical processes, where sum-
 401 mer rainfall is observed at higher latitudes (e.g. East Asia).

402 **2.2 Energetic constraints**

403 The regional monsoons are an integral part of the ITCZ, here interpreted as the
 404 longitudinal band of maximal tropical rainfall. One important step in understanding mon-
 405 soon dynamics has therefore been an exploration of the mechanisms governing the be-
 406 havior of the zonally averaged ITCZ. For example, in the annual and zonal mean, the
 407 ITCZ is located in the Northern Hemisphere (at 1.7°N if estimated by the precipitation
 408 centroid; Donohoe, Marshall, Ferreira, and Mcgee (2013), or $\sim 6^\circ\text{N}$ if judged by the pre-
 409 cipitation maximum; e.g. Gruber, Su, Kanamitsu, and Schemm (2000)). While it is usu-
 410 ally the case that the ITCZ is co-located with SST maxima, both paleoclimate proxies
 411 (e.g. Fig. 3; Arbuszewski, Demenocal, Cléroux, Bradtmiller, & Mix, 2013; Lea, Pak, Pe-
 412 terson, & Hughen, 2003; McGee, Donohoe, Marshall, & Ferreira, 2014) and model sim-
 413 ulations (Broccoli, Dahl, & Stouffer, 2006; J. C. H. Chiang & Bitz, 2005; R. Zhang &
 414 Delworth, 2005) indicate that the ITCZ responds to extratropical forcing, that is, to forc-
 415 ing remote from its location. Analysis of the atmospheric and oceanic energy budget has
 416 helped to explain these behaviors.

417 Not surprisingly, aquaplanet simulations have been used to examine systematically
 418 controls on the ITCZ latitude by imposing a prescribed hemispherically asymmetric forg-
 419 ing in the extratropics and varying its strength. Kang, Held, Frierson, and Zhao (2008)
 420 found that the atmospheric energy transport associated with the Hadley cell largely com-
 421 pensates for changes in hemispherically asymmetric extratropical surface heating. The
 422 Hadley cell diverges energy away from its ascending branch, i.e. away from the ITCZ,
 423 and generally transports energy in the direction of the upper-level meridional flow. Hence
 424 a hemispherically asymmetric atmospheric heating will cause the ITCZ to shift towards
 425 the hemisphere with the greater heating, as illustrated in Fig. 9. Kang et al. (2008) fur-

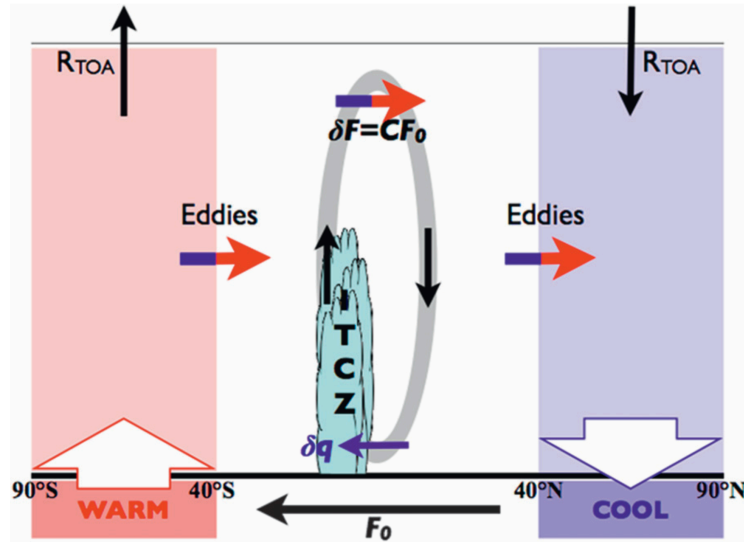


Figure 9. Schematic illustrating the energetics framework to determine the tropical response to extratropical thermal forcing (Kang et al., 2009). Warming is applied to the southern extratropical slab ocean, giving an implied ocean heat transport anomaly F_o . The atmosphere compensates for the additional warming by altering the top-of-atmosphere net radiative flux (R_{TOA}) and horizontal energy transports by the atmosphere. In the tropics, the gray oval indicates the anomalous Hadley circulation response, the direction of which is represented by black arrows. The blue (red) part of the colored arrow indicates regions where energy transports act to (anomalously) cool (warm) the atmosphere. These energy transports are due to midlatitude eddies and the Hadley circulation. The clockwise anomalous Hadley circulation transports energy northward to cool (warm) the southern (northern) subtropics and largely compensates the warming (cooling) by eddies. This figure is taken from Kang et al. (2009). ©American Meteorological Society. Used with permission.

426 ther noted that the ITCZ latitude was approximately colocated with the ‘Energy Flux
 427 Equator’ (EFE), the latitude at which the vertically integrated MSE flux is zero, and
 428 that it varied proportionally to the strength of the asymmetric forcing. This anticorrelation
 429 between the ITCZ latitude and the cross-equatorial atmospheric energy transport
 430 in the tropics has since been observed in aquaplanet models with different physical parametrisations
 431 (Kang et al., 2009), and in models with realistic continental configurations under global warming and
 432 paleoclimate scenarios (Donohoe et al., 2013; D. M. W. Frierson & Hwang, 2012). In the zonal average,
 433 the latitudinal location of tropical precipitation appears more tightly controlled by changes in hemispheric
 434 asymmetric surface energy flux forcing than by changes in SST (Kang & Held, 2012), highlighting the power
 435 of this energetic framework. However, the degree of compensation between the imposed heating and the
 436 atmospheric energy transport is sensitive to the parameterisations of convection, clouds, and ice
 437 (D. M. W. Frierson & Hwang, 2012; Kang et al., 2009, 2008), and to changes in energy transport by the
 438 ocean, which has been shown to play a significant role in the energy transport response to an imposed
 439 perturbation (Green & Marshall, 2017; Levine & Schneider, 2011; Schneider, 2017).

442 The relationship between the ITCZ, EFE, and tropical atmospheric energy transport can be understood
 443 more quantitatively using the steady state, zonally averaged, ver-

444 tically integrated energy budget,

$$\bar{\mathcal{S}} - \bar{\mathcal{L}} - \bar{\mathcal{O}} = \frac{\partial \langle \overline{vh} \rangle}{\partial y}. \quad (10)$$

445 In the above, \mathcal{S} is the net downward top of atmosphere shortwave radiation, \mathcal{L} the out-
 446 going longwave radiation and \mathcal{O} represents any net energy uptake at the surface. An-
 447 gular brackets denote a vertical integral, and overbars a time and zonal mean. Eq. 10
 448 states that net energy input into the atmospheric column through top-of-atmosphere ra-
 449 diative fluxes and surface energy fluxes must be in balance with meridional convergence
 450 or divergence of MSE into the atmospheric column. For small meridional displacements,
 451 δ , this equation can be Taylor expanded around the Equator to 3rd order as (Bischoff
 452 & Schneider, 2014, 2016)

$$\langle \overline{vh} \rangle_\delta = \langle \overline{vh} \rangle_0 + a \partial_y \langle \overline{vh} \rangle_0 \delta + \frac{1}{2} a^2 \partial_{yy} \langle \overline{vh} \rangle_0 \delta^2 + \frac{1}{6} a^3 \partial_{yyy} \langle \overline{vh} \rangle_0 \delta^3, \quad (11)$$

453 where the $_0$ subscript denotes quantities evaluated at the Equator. At the EFE, by def-
 454 inition, the vertically integrated, zonal mean MSE flux, $\langle \overline{vh} \rangle$, is zero. Taking δ as the lat-
 455 itude of the EFE, and substituting in from Eq. 10, gives

$$0 = \langle \overline{vh} \rangle_0 + a(\bar{\mathcal{S}} - \bar{\mathcal{L}} - \bar{\mathcal{O}})_0 \delta + \frac{1}{2} a^2 \partial_y (\bar{\mathcal{S}} - \bar{\mathcal{L}} - \bar{\mathcal{O}})_0 \delta^2 + \frac{1}{6} a^3 \partial_{yy} (\bar{\mathcal{S}} - \bar{\mathcal{L}} - \bar{\mathcal{O}})_0 \delta^3. \quad (12)$$

456 The net energy input ($\bar{\mathcal{S}} - \bar{\mathcal{L}} - \bar{\mathcal{O}}$) is approximately symmetric about the Equator, so
 457 the quadratic term is small relative to the other terms (Bischoff & Schneider, 2016), and
 458 can be neglected. Hence, to a good approximation, Eq. 12 can be written as

$$\delta = -\frac{\langle \overline{vh} \rangle_0}{a(\bar{\mathcal{S}} - \bar{\mathcal{L}} - \bar{\mathcal{O}})_0}. \quad (13)$$

459 Eq. 13 has been shown to give a good estimate of the EFE latitude under a range of warm-
 460 ing scenarios in aquaplanets (Bischoff & Schneider, 2014), and over the annual cycle in
 461 reanalysis (Adam, Bischoff, & Schneider, 2016b). The EFE in turn acts as an indicator
 462 of the ITCZ latitude. More broadly, (Bischoff & Schneider, 2016) found that the first
 463 order approximation is adequate when the net energy input at the Equator is large and
 464 positive, but that the cubic term is needed when this is small or negative. Notably the
 465 negative case corresponds to a double ITCZ.

466 Unfortunately, the ITCZ and EFE latitudes do not covary on all timescales. In par-
 467 ticular these can deviate from one another significantly over the seasonal cycle (e.g. Adam
 468 et al., 2016b; Wei & Bordoni, 2018). While the EFE denotes the latitude at which the
 469 meridional MSE flux changes sign, the ITCZ is associated with the ascending branch of
 470 the tropical meridional overturning circulation, which is close to the latitude where the
 471 mass flux changes sign. The energy flux and overturning circulation are related via the
 472 gross moist stability (Δ_M , defined here following e.g. D. M. W. Frierson, 2007; Hill, Ming,
 473 & Held, 2015; Wei & Bordoni, 2018):

$$\Delta_M = \frac{\langle \overline{vh} \rangle}{\Psi_{max}} = \frac{\langle \overline{vh} \rangle}{g^{-1} \int_0^{p_m} \bar{v} dp}. \quad (14)$$

474 In the above, Ψ_{max} is the maximum of the overturning streamfunction, corresponding
 475 to the mass flux by the Hadley cell, and p_m is the pressure level at which this maximum
 476 occurs. Considering Eq. 14 at the Equator, and combining with Eq. 13, we see that the
 477 strength of the Hadley circulation (and hence the ITCZ) will therefore covary with the
 478 EFE provided that the efficiency with which the Hadley cell transports energy, as cap-
 479 tured by Δ_M , remains approximately constant. However, recent aquaplanet simulations
 480 indicate that over the seasonal cycle Δ_M varies significantly, and in fact at times becomes
 481 negative, allowing the EFE and ITCZ to sit in opposite hemispheres (Wei & Bordoni,

482 2018). Gross moist stability has also been observed to vary significantly under changes
 483 to orbital precession in aquaplanet simulations (Merlis, Schneider, Bordoni, & Eisenman,
 484 2013). It is also worth noting that, in addition to variations in Δ_M , the zonal mean en-
 485 ergy flux compensating an energetic forcing may be achieved by transient eddies, rather
 486 than by changes to the zonal mean overturning circulation (Xiang, Zhao, Ming, Yu, &
 487 Kang, 2018).

488 Despite these limitations, the energetic framework has been a major advance, and
 489 has given insight into variations in tropical rainfall over both the observational and paleo
 490 record (see review by Schneider et al., 2014, and references therein). One attractive
 491 feature of this perspective is that it provides a simple explanation for why, in the annual
 492 and zonal mean, the ITCZ sits in the Northern Hemisphere (Donohoe et al., 2013; Gru-
 493 ber et al., 2000). The energetic framework neatly shows that the ITCZ latitude can be
 494 understood as a result of the net flux of energy into the Northern Hemisphere by the ocean,
 495 in particular due to asymmetry introduced by the Drake passage (D. M. Frierson et al.,
 496 2013; Fućkar, Xie, Farneti, Maroon, & Frierson, 2013; Marshall, Donohoe, Ferreira, &
 497 McGee, 2014). Efforts to extend this framework to account for zonal asymmetry in the
 498 boundary conditions (the ‘Energy Flux Prime Meridian’ Boos & Korty, 2016) are dis-
 499 cussed in Section 3.2.

500 **3 Interpreting observations and modeled response to forcings**

501 In parallel with the theoretical developments described in Section 2, observational
 502 and reanalysis datasets have allowed more detailed analysis of the behavior of Earth’s
 503 monsoons. As discussed in Section 1, one major step has been moving from a perspec-
 504 tive of monsoons as individual, unrelated systems, to a perspective of a global monsoon
 505 manifesting itself into several regional systems (P. X. Wang et al., 2014, 2017, and refer-
 506 ences therein). In this section, we look at the insight into the dynamics of Earth’s mon-
 507 soons gained from observations and Earth System models, and at how it connects to the
 508 theoretical ideas developed using idealized model simulations discussed in Section 2. First,
 509 we give an overview of the characteristics of Earth’s regional monsoons and ITCZs, and
 510 their role as part of the global monsoon. We then discuss the extent to which theory, par-
 511 ticularly that from aquaplanet models, may help us understand the behavior of these sys-
 512 tems in current climate, and offer possible constraints on future climate.

513 **3.1 The global and regional monsoons**

514 The magenta line in Fig. 1a-c marks out the regional monsoons, indicating areas
 515 where the local difference between summer and winter precipitation exceeds 2.5 mm/day,
 516 and where summer precipitation accounts for the majority of the annual total. Six re-
 517 gions can be identified: Asia, West Africa, Southern Africa, South America, North Amer-
 518 ica and Australia (cf. S. Zhang & Wang, 2008). The Asian monsoon is the most intense
 519 and largest in scale of these, and is often further divided into several subregions: the South
 520 Asian, East Asian, and Western North Pacific monsoons, as shown in Fig. 10. (B. Wang
 521 & LinHo, 2002).

522 ***3.1.1 Regional monsoon and ITCZ characteristics***

523 *South Asian Monsoon*

524 The South Asian monsoon features a wind reversal from winter easterlies to sum-
 525 mer westerlies at low levels (e.g. B. Wang & LinHo, 2002). Onset spreads from the south
 526 to the north, with the earliest onset of the system over the Southern Bay of Bengal, be-
 527 tween late April and mid-May (Mao & Wu, 2007), reaching Kerala between mid-May
 528 and mid-June (Ananthkrishnan & Soman, 1988; J. M. Walker & Bordoni, 2016; B. Wang,

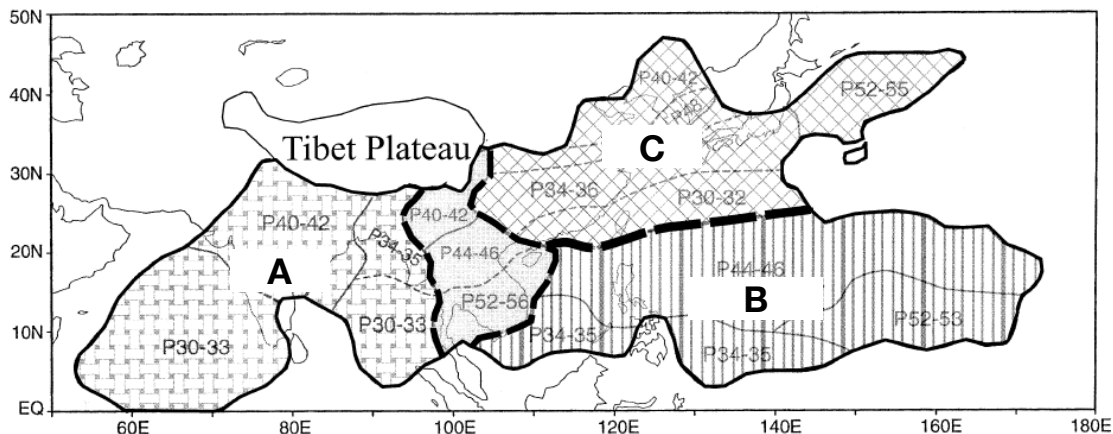


Figure 10. Map showing the division of the Asian monsoon into three subregions. The South Asian monsoon (A) and Western North Pacific monsoon (B) are tropical monsoon regions. A broad corridor in the Indochina Peninsula separates them. The subtropical monsoon region is identified as the East Asian monsoon (C), north of the Western North Pacific monsoon. Numbers indicate the pentad range during which the peak monsoon rainfall occurs. Adapted from B. Wang and LinHo (2002). ©American Meteorological Society. Used with permission.

529 Ding, & Joseph, 2009). Onset occurs over the South China Sea between early May and
 530 mid-June (B. Wang, LinHo, Zhang, & Lu, 2004).

531 The wet season over India generally lasts from June to September, during which
 532 time about 78% of the total annual rain falls over India (Parthasarathy, Munot, & Kothawale,
 533 1994). The rain band withdraws towards the Equator between late September and early
 534 November (B. Wang & LinHo, 2002).

535 *East Asian Monsoon*

536 While the South Asian monsoon is confined to latitudes southward of $\sim 30^\circ\text{N}$, the
 537 East Asian monsoon extends north of this into the extratropics. Although the monsoon
 538 onset over the South China Sea has been considered a precursor to the East Asian mon-
 539 soon onset (Martin et al., 2019; B. Wang et al., 2004), some authors (e.g. B. Wang &
 540 LinHo, 2002) consider it as an entirely subtropical system, defining the East Asian mon-
 541 soon region to include $\sim 20\text{--}45^\circ\text{N}$ and $110\text{--}140^\circ\text{E}$. A key element of the East Asian mon-
 542 soon is an east-west oriented band of precipitation, known as Meiyu in China and Baiu
 543 in Japan, that is accompanied by a wind reversal from winter northerlies to summer souther-
 544 lies. The Meiyu front brings intense rainfall to the Yangtze River valley and Japan from
 545 mid-June to mid-July, after which it breaks down and allows rainfall to extend into north-
 546 ern China and Korea. Prior to the onset of the Meiyu, South China experiences peri-
 547 ods of rainfall in the form of the South China spring rain, intensifying from mid-March
 548 to May (LinHo, Huang, & Lau, 2008). A thorough review of the East Asian monsoon's
 549 characteristics, including its onset and development, governing processes and telecon-
 550 nections is given in Ding and Chan (2005); see also Section 4.1.1.

551 *Western North Pacific Monsoon*

552 Monsoon rains arrive later over the Western North Pacific region (see Fig. 10) than
 553 the South and East Asian sectors, and last from July to October/November (B. Wang
 554 & LinHo, 2002). The monsoon advances from the south-west to north-east in a stepwise
 555 pattern associated with shifts in the Western North Pacific subtropical high (R. Wu &

556 Wang, 2001), while withdrawal occurs from the north-west to south-east (S. Zhang &
 557 Wang, 2008). A predominantly zonally oriented change in wind direction is seen between
 558 winter and summer, associated with a weakening of the low-latitude easterly flow as the
 559 Western North Pacific subtropical high shifts eastward (e.g. Fig. 1).

560 *Indonesian-Australian Monsoon*

561 The Indonesian-Australian monsoon develops over Java in October-November, and
 562 progresses southeastward, reaching northern Australia in late December (Hendon & Lieb-
 563 mann, 1990; S. Zhang & Wang, 2008). During austral summer, the low-latitude easter-
 564 lies over the western Maritime Continent reverse to a southwesterly flow, as seen in Fig.
 565 1b and c. Monsoon withdrawal occurs over northern Australia and the southeastern Mar-
 566 itime Continent through March, with the wet season persisting into April over Java (S. Zhang
 567 & Wang, 2008).

568 *West African Monsoon*

569 The West African monsoon begins near the Equator, with intense rainfall over the
 570 Gulf of Guinea in April. This continues through to the end of June, with a second max-
 571 imum developing near 10°N in late May. The peak precipitation is observed to jump rapidly
 572 to this second maximum in late June, accompanied by a reversal of the wind direction
 573 from north-easterly to south-westerly to the south of this maximum (Sultan & Janicot,
 574 2003). Precipitation weakens from August to September and the peak rainfall migrates
 575 back towards the Equator. Over the Sahel, the monsoon precipitation accounts for 75-
 576 90% of the total annual rainfall (Lebel, 2003).

577 *Southern African Monsoon*

578 The Southern African monsoon is offset longitudinally to the east of its Northern
 579 Hemisphere counterpart. The global monsoon onset metric of S. Zhang and Wang (2008)
 580 indicates that the rainy season begins in November over Angola and the southern DRC,
 581 and extends southeastward over the continent, progressing over southern Tanzania, Zam-
 582 bia and out over the ocean over northern Madagascar through December, and reaching
 583 Zimbabwe, Mozambique, and as far as the northeast of South Africa by January. The
 584 system extends out over the Southwestern Indian Ocean through January and Febru-
 585 ary. Withdrawal occurs directed towards the north and west from February to April. In
 586 austral winter, the prevailing wind is southeasterly, but in summer this reverses to a weak
 587 northeasterly flow, with stronger northeasterly flow to the north of the region, over the
 588 Horn of Africa, as seen in Fig. 1c.

589 In the same longitudinal region, an East African monsoon is also discussed over the
 590 Greater Horn of Africa. Near the Equator, there are two wet seasons: the long rains from
 591 March to May, and the short rains from October-November (Nicholson, 2017). Farther
 592 to the north, from June to September a boreal summer wet season occurs in Ethiopia
 593 and Eritrea, with lower-level westerlies developing in June and persisting through to Au-
 594 gust (Segele & Lamb, 2005).

595 *North American Monsoon*

596 The North American monsoon is observed as a marked increase in precipitation
 597 over the southwestern United States and Mexico, beginning in June-July, and withdraw-
 598 ing through September and October (Adams & Comrie, 1997; Barlow, Nigam, & Berbery,
 599 1998; Ellis, Saffell, & Hawkins, 2004). S. Zhang and Wang (2008) observed that onset
 600 (withdrawal) over this area occurs in a northward (southward) moving band. There is
 601 no large-scale reversal of the winds in this region (see Figs. 1b and 1c). However, the
 602 northwesterly flow down the coast of California observed in boreal winter weakens in bo-
 603 real summer, the southeasterly flow over the east coast of Mexico strengthens, and the
 604 low-latitude easterlies over the eastern Pacific weaken in the Northern Hemisphere (e.g.

Fig. 7, Barlow et al., 1998). In addition, at a smaller scale, the low-level wind direction reverses over the Gulf of California from northerly to southerly flow (Bordoni, Ciesielski, Johnson, McNoldy, & Stevens, 2004).

South American Monsoon

The monsoon season in South America begins in October, with an abrupt shift of convection southward over the Amazon river basin (Marengo et al., 2012). The precipitation progresses southeastward through November and December (S. Zhang & Wang, 2008). Withdrawal occurs from March to May, with the rain-band returning northward. During austral winter, the prevailing 850-hPa winds over the continent are predominantly easterly between 10°S and 10°N, but in summer the flow becomes northeasterly and cross equatorial, and a northwesterly jet, the South American Low-Level jet, develops along the east side of the Andes (Marengo et al., 2012). An upper-level anticyclone is observed over Bolivia, and a lower-level cyclone develops over northern Argentina (Rao, Cavalcanti, & Hada, 1996). Central Brazil receives over 70% of its annual rainfall during the monsoon season, between September and February (Rao et al., 1996).

The Atlantic and Pacific ITCZs

The latitudinal position of the ITCZs in the Atlantic and Pacific also has a distinct seasonal cycle, as can be seen from the north-south dipole in the October/November-April/May precipitation difference, shown in Fig. 1b. Over the continents, the latitudinal extremes in the position of the zonal average precipitation follows the Sun into the summer hemisphere and lags insolation by about a month due to the small heat capacity of the land-atmosphere system that participates in the seasonal cycle. In contrast, precipitation associated with the Atlantic and Pacific ITCZs reaches farthest north in October and farthest south (but still north of the Equator; see Section 4.1.3) in March about three months after the boreal and austral solstice, respectively (Fig. 8). These ITCZs tend to be located just equatorward of the warmest water in subtropical Northern Hemisphere, and the latter lags insolation by about three months due to the large heat capacity of the upper ocean that participates in the seasonal cycle.

The ITCZs are not a passive response to changes in SST, however: ocean-atmosphere interactions play a role in regulating the seasonal cycle of Atlantic and Pacific ITCZs (see Battisti, Vimont, & Kirtman, 2019, and references therein). As the water warms the sea level pressure drops and the ITCZ moves northward (a hydrostatic response; see Lindzen & Nigam, 1987); the cross-equatorial sea level pressure gradient increases, driving an increase in southerly winds near and just south of the Equator (Fig. 8) that increases the turbulent energy fluxes from the ocean to the atmosphere, cooling the ocean south of the Equator. This further increases the cross-equatorial pressure gradient and the southeasterly winds, which brings more moisture into the ITCZs.

3.1.2 The Global Monsoon

The regional monsoons exhibit a diverse range of behaviors, but some common features can be identified. From Fig. 1a, it can be seen that most monsoon regions feature anomalous westerly low-level flow in their summer season, with a cross-equatorial component directed into the summer hemisphere. However, comparing Figs. 1b and c shows that these anomalies are not always sufficient to cause a local reversal of the wind direction. Onset generally occurs as a poleward advancement of rainfall off of the Equator, often with an eastward directed progression. Onset also sometimes features sudden jumps or steps in the latitude (poleward) and longitude of precipitation, as observed over South Asia, West Africa, the Western North Pacific, and South America.

Paleoclimate reconstructions, present-day observations, and model simulations have begun to elucidate how the monsoons vary under a range of external and internal forc-

ings. These findings were recently summarized comprehensively by P. X. Wang et al. (2014, 2017), An et al. (2015) and Seth et al. (2019). Forcings that preferentially warm or cool one hemisphere relative to the other, such as Heinrich events and changes in Earth’s axial precession, are found to intensify the monsoons of the warmer hemisphere, and to weaken the monsoons of the cooler hemisphere (e.g. Battisti, Ding, & Roe, 2014; H. Cheng, Sinha, Wang, Cruz, & Edwards, 2012; Eroglu et al., 2016; Liu & Battisti, 2015; Pausata, Battisti, Nisancioglu, & Bitz, 2011; P. X. Wang et al., 2014). On interannual timescales, ENSO has also been found to modulate monsoon precipitation in both hemispheres, with El Niño associated with drying and La Niña associated with enhanced rainfall (B. Wang, Liu, Kim, Webster, & Yim, 2012).

The above similarities in the climatology and variability of regional monsoons have motivated a view of the local monsoon systems as components of a global monsoon, which responds to forcings in a coherent manner. The global monsoon might be interpreted as the seasonal migration of the ITCZ from north to south throughout the year, with regional monsoons corresponding to locations where this migration is enhanced (e.g. Gadgil, 2018).

3.2 Aquaplanet-like monsoons

Aquaplanet-based theoretical work, as discussed in Section 2, has used symmetric boundary conditions to study the fundamental processes governing the zonal mean ITCZ, Hadley cells, and global monsoon. In contrast, the bulk of studies using observations, reanalysis, and Earth System models have tended to focus on the mechanisms controlling regional monsoons. While local factors are of great importance in determining the seasonal evolution and the variability of the individual monsoon systems, we argue here that aquaplanet results can inform us of unanticipated commonalities in the dynamics of the monsoons, and help us interpret the behaviors observed. Of the two perspectives discussed in Section 2 the energetic approach has received more attention (Bisutti et al., 2018; Schneider et al., 2014), perhaps due to the relative ease with which the relevant diagnostics can be evaluated and the intuitive picture it presents (Fig. 9). In this section we discuss how both approaches provide useful insights into the dynamics of Earth’s monsoons.

3.2.1 *Insight from the momentum budget and CQE considerations*

For an aquaplanet, the momentum framework, combined with the assumption of CQE, indicates that:

1. The poleward boundary of a monsoon circulation lies just equatorward of the peak in subcloud MSE or θ_{eb} (Emanuel, 1995; Privé & Plumb, 2007a, 2007b).
2. Cross-equatorial ‘monsoon’ Hadley cells approach conservation of angular momentum more than equinoctial cells, and consequently are more strongly coupled to meridional MSE gradients (Schneider & Bordoni, 2008).
3. Rapid transitions can occur between an ITCZ regime (with two eddy-driven Hadley cells) and a monsoon regime with one dominant cell that extends into the summer hemisphere (cf. Figs. 6a and 6b). These transitions are mediated by feedbacks relating to advection of MSE in the lower branch of the Hadley circulation, and suppression of eddies by upper-level easterlies. (Bordoni & Schneider, 2008, 2010; Schneider & Bordoni, 2008).
4. At Earth’s rotation rate, the transition from the eddy-driven to angular momentum conserving Hadley cell regime appears to occur at $\sim 7^\circ$ latitude on an aquaplanet with zonally symmetric boundary conditions (Geen et al., 2019).

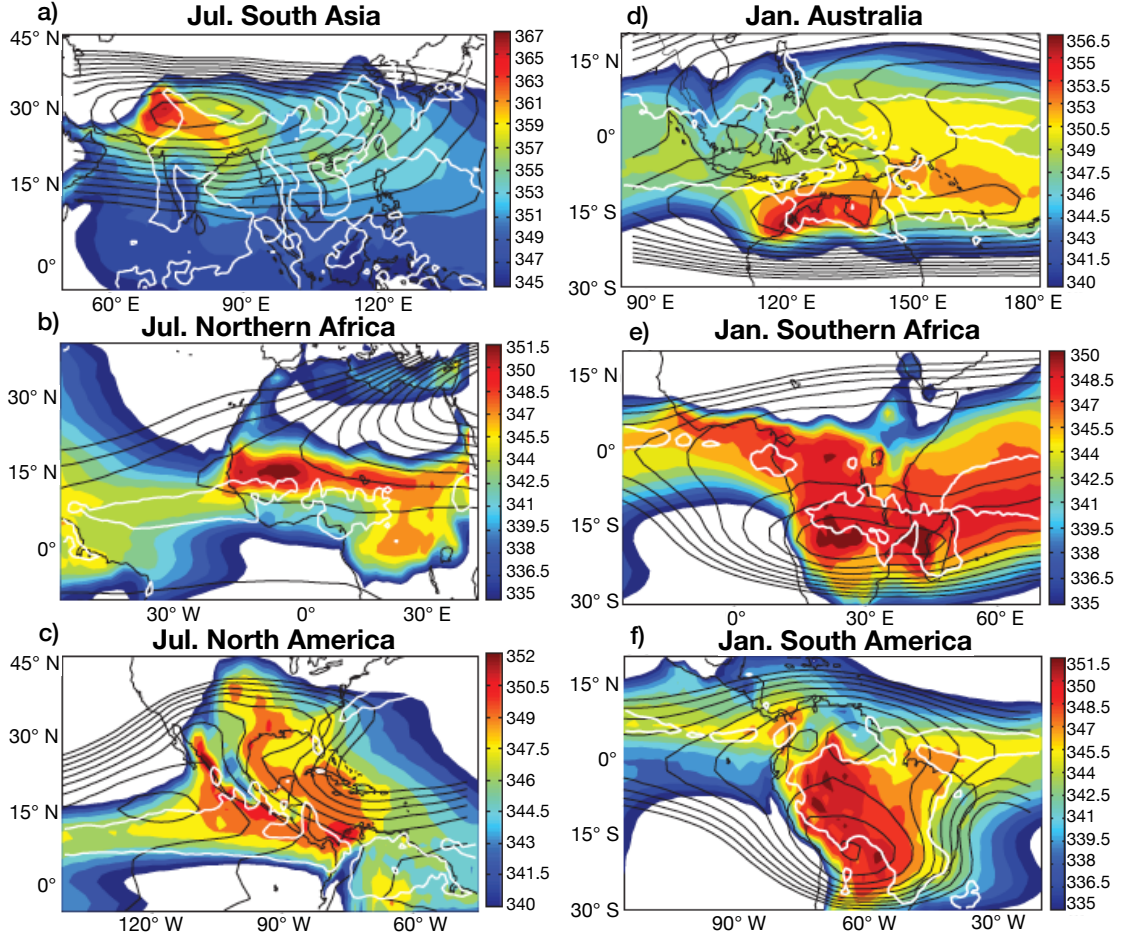


Figure 11. Evaluation of CQE for the (a) South Asia, (b) northern Africa, (c) North America, (d) Australia, (e) southern Africa and (f) South America monsoons. Colors show subcloud equivalent potential temperature, θ_{eb} . The black contour is the free tropospheric saturation equivalent potential temperature, θ_e^* , averaged from 200 to 400 hPa. The white contour indicates the region that has precipitation greater than 6 mm day^{-1} . The θ_e^* contours start from (a) 345 K, (b) 340 K, (c) 340 K, (d-f) 341 K and the respective interval is (a) 1 K, (b) 1 K, and (c-f) 0.5 K. Adapted from Nie et al. (2010). ©American Meteorological Society. Used with permission.

701 5. At Earth’s rotation rate, convergence zones within the ascending branches of seasonal cross-equatorial Hadley cells appear to be unable to migrate farther than
 702 $\sim 30^\circ$ from the Equator (Faulk et al., 2017; Hill et al., 2019; Singh, 2019).
 703

704 The above ideas were developed in a very idealized framework, but consistent behavior
 705 has been observed on Earth. Nie et al. (2010) investigated whether the CQE assumption
 706 was relevant locally in the regional monsoons. By analysing ERA-40 and Tropical
 707 Rainfall Measuring Mission (TRMM) data, they demonstrated that, in the South Asian,
 708 Australian, and African monsoons, maxima of θ_{eb} and free-troposphere saturation equivalent
 709 potential temperature are approximately collocated, and peak precipitation indeed
 710 lies just equatorward of the peak in subcloud MSE, consistent with CQE predictions (Fig.
 711 11). Also consistent with the idealized modeling work, seasonal changes in the character
 712 of the overturning circulation have been observed in the regional monsoons. The Hadley
 713 circulation over the South Asian monsoon region in particular has been highlighted as

714 showing rapid transitions between an eddy-driven and an angular momentum conserving
 715 Hadley circulation that are similar to those seen in aquaplanet simulations. In this
 716 region, precipitation migrates rapidly off the Equator and the summertime circulation
 717 is nearly angular momentum conserving (Bordoni & Schneider, 2008; Geen et al., 2018;
 718 J. M. Walker & Bordoni, 2016). To give an indication of other regions where angular mo-
 719 mentum conservation may apply, Fig. 12 shows the local overturning circulation, defined
 720 using the divergent component of the meridional wind (e.g. Schwendike et al., 2014; G. Zhang
 721 & Wang, 2013) for each of the monsoon regions marked in Fig. 1. Angular momentum
 722 contours are plotted in gray. The upper-level summertime overturning circulation be-
 723 comes roughly aligned with angular momentum contours in the Asian and African monsoon
 724 regions. These areas are consistent with areas in which CQE has been shown to ap-
 725 pply (Nie et al., 2010).

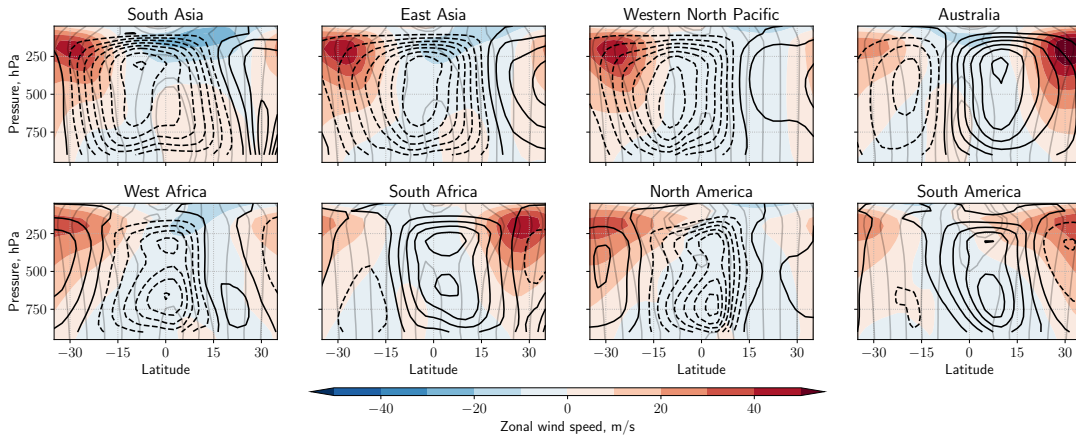


Figure 12. Black contours show local summer (June-September or December-March) meridional overturning circulations for: South Asia (70-100°E), East Asia (110-140°E), Western North Pacific (140-160°E), Australia (100-170°E), West Africa (35°W-45°E), South Africa (20-90°E), North America (80-120°W), South America (40-80°W). Shading shows zonal wind. Light gray contours indicate absolute angular momentum per unit mass, with contours at $\Omega a^2 \cos^2 \phi_i$ ($\phi_i = 0^\circ, \pm 5^\circ \pm 10^\circ, \dots$).

726 The climatological monsoons show consistency with the aquaplanet results. Aware-
 727 ness of the relevance of the lower-level MSE and upper-level wind structures to the merid-
 728 ional overturning circulation may additionally help in understanding present day vari-
 729 ability of the monsoons and model projections of future climate. For example, Hurley
 730 and Boos (2013) used reanalysis and observational datasets to explore whether variabil-
 731 ity in monsoon precipitation could be connected to variability in θ_{eb} . Even removing the
 732 signal of variability linked to ENSO, they found that positive precipitation anomalies
 733 in the American, African, South Asian and Australian monsoons were associated with
 734 enhanced θ_{eb} , consistent with previous findings over West Africa (Eltahir & Gong, 1996).
 735 In addition, variability in θ_{eb} was found to be due primarily to variability in moisture
 736 rather than in temperature, with strong monsoon years associated with enhanced spec-
 737 ific humidity near the climatological θ_{eb} maximum, with temperature anomalies of the
 738 opposite sign (see also J. M. Walker, Bordoni, & Schneider, 2015). This clearly contra-
 739 dicts the classical sea-breeze view of the monsoons, but is consistent with the CQE per-
 740 spective. Shaw and Voigt (2015) showed that the CQE perspective can help to explain
 741 the weak response of the Asian monsoons to global warming seen in climate model pro-
 742 jections. Using data from the Atmosphere Model Intercomparison Project (AMIP) ex-
 743 periments, they compared the circulation response to a quadrupling of CO₂ with fixed

744 SSTs (AMIP4xCO₂) with the response to a uniform 4K increase in SST (expected due
 745 to a 4x increase in CO₂), but with no CO₂ increase (AMIP4K). They found that the CO₂
 746 forcing led to θ_{eb} changes that supported a more intense monsoon, but the SST forcing
 747 led to opposite θ_{eb} changes which, they argued, led to a weak net response to an increase
 748 in CO₂.

749 The connection of precipitation to lower-level MSE anomalies predicted by CQE
 750 is an intuitive focus for understanding monsoon variability and future change. The con-
 751 nection to the upper-level momentum budget has not yet been so comprehensively in-
 752 vestigated. However, it has been observed that anomalous upper-level easterlies and west-
 753 erlies are associated with anomalous upper-level divergence and convergence in monsoon
 754 regions in a sense that is consistent with the aquaplanet angular momentum conserv-
 755 ing monsoon regime. Enhanced upper-level tropical easterlies, consistent with a higher
 756 Rossby number at upper levels (cf. Eq. 4), accompany more intense precipitation over
 757 West Africa via enhancement of upper-level divergence and meridional overturning (Nichol-
 758 son, 2009b). On intraseasonal and interannual timescales over South Asia and West Africa,
 759 and interannual timescales over the Western North Pacific, anomalously wet conditions
 760 are associated with easterly upper-level zonal wind anomalies, westerly lower-level zonal
 761 wind anomalies, and expansion and strengthening of the meridional overturning, with
 762 the opposite applying in dry phases (Goswami & Ajaya Mohan, 2001; Sultan & Janicot,
 763 2003; J. M. Walker et al., 2015; B. Wang, Wu, & Lau, 2001). Modulation of the mon-
 764 soons by anomalous upper-level flow may help in understanding teleconnections between
 765 regional monsoons, although more work is needed to explore the mechanisms involved
 766 and to ascertain the direction of causality between anomalous upper- and lower-level cir-
 767 culations.

768 The recent findings summarized in points (4) and (5) above suggest that planetary
 769 rotation constrains the maximum latitude that the ITCZ can reach, and the latitude at
 770 which the overturning circulation tends to transition from an eddy-driven to an angu-
 771 lar momentum conserving regime. The implications for Earth's tropical circulations re-
 772 main to be explored. However, one could imagine that these latitudinal bounds might
 773 provide information on what circulation regime we expect to be associated with a ver-
 774 tical mass flux at a given latitude. Fig. 13 shows the mass fluxes associated with merid-
 775 ional and zonal overturning circulations for May to September and November to March
 776 (cf. Schwendike et al., 2014). Gray shading indicates the region between 10-30° from the
 777 Equator. Consistent with the findings of Faulk et al. (2017) for the aquaplanet circula-
 778 tion, the upward mass fluxes associated with the Hadley cell are confined to within 30°
 779 of the Equator. One might further speculate that circulations for which the upward mass
 780 flux peaks between 10-30° from the Equator (Asia, Southern Africa) might bear simi-
 781 larities to the aquaplanet angular momentum conserving regime, while those where as-
 782 cent remains equatorward of 10° (Australia, North America, the Atlantic and Pacific ITCZs)
 783 might behave more like the aquaplanet eddy-driven regime. Figs. 12 and 13 suggest this
 784 idea shows promise, with, for example, the summer overturning circulation over Australia
 785 remaining in an eddy-driven regime, while the circulation over areas such as South Asia
 786 and Southern Africa becomes more aligned with angular momentum contours. These cat-
 787 egorisations of the various flow regimes associated with tropical rainfall could be of use
 788 in interpreting the responses of different regions to external forcings.

789 While the aquaplanet results provide a simple framework for interpreting regional
 790 monsoons and their variability, some caveats must be remembered. The regional mon-
 791 soons are local, stationary wave governed systems, with overturning associated with both
 792 meridional and zonal flow (e.g. Fig. 13). Simple symmetric theories do not necessarily
 793 extend straightforwardly to these cases, with stationary waves modifying the momen-
 794 tum and energy budgets (Shaw, 2014). Also, in addition to the deep, moist convective
 795 overturning circulation, many of the regional monsoons also feature a shallow, dry cir-
 796 culation whose ascent is colocated with the peak in potential temperature (e.g. Hagos

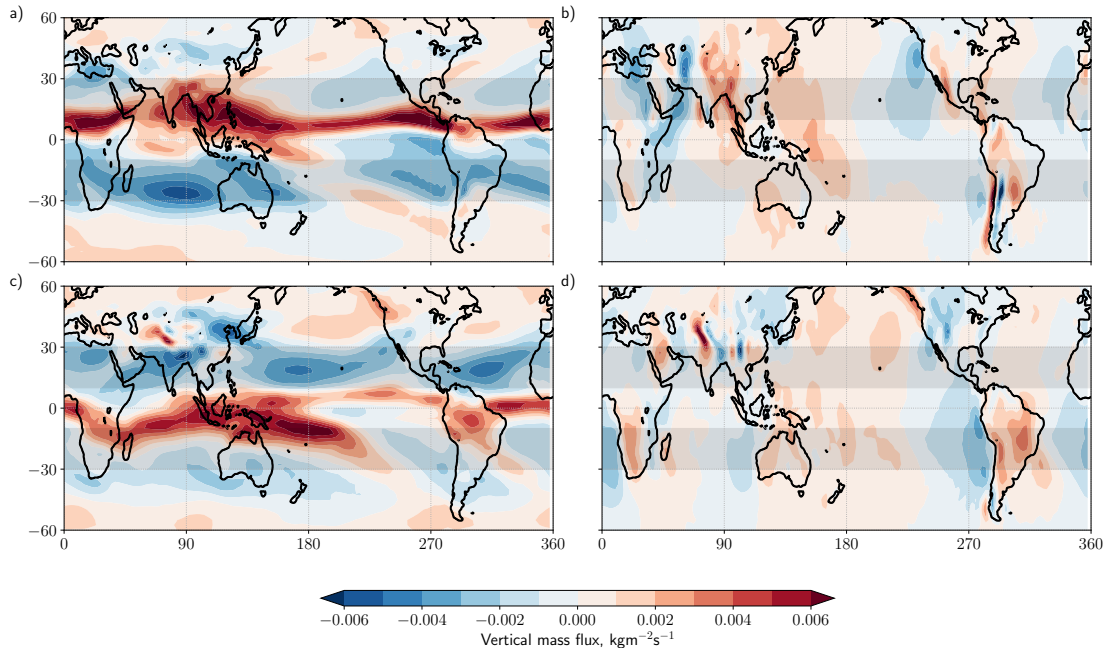


Figure 13. Vertical mass flux at 500 hPa, calculated from JRA-55, associated with (a) the divergent meridional circulation and (b) the divergent zonal circulation (cf. Schwendike et al., 2014) in boreal summer, defined as in Fig. 12. (c) and (d) are as (a) and (b) but for austral summer. Gray shading highlights the regions between 10 and 30°N/S, see discussion in text.

797 & Cook, 2007; Nie et al., 2010; Trenberth, Stepaniak, & Caron, 2000; C. Zhang, Nolan,
 798 Thorncroft, & Nguyen, 2008); advective drying by this shallow circulation appears to sup-
 799 press monsoon precipitation (Zhai & Boos, 2017). Accounting for this behavior requires
 800 some adjustments to the CQE assumption (Shekhar & Boos, 2016).

801 *3.2.2 Insight from energetic considerations*

802 As reviewed in Section 2.2., the vertically integrated atmospheric energy budget
 803 provides a complementary approach to understanding constraints on tropical rainfall.
 804 An elegant finding from applying this in aquaplanets is that the ITCZ approximately
 805 follows the EFE, so that changes in zonal mean ITCZ latitude can be linked to changes
 806 in net forcing not only in the tropics, but also at higher latitudes (see Section 2.2 and
 807 e.g. Bischoff & Schneider, 2014; Kang et al., 2008). Additionally, the MSE budget al-
 808 lows for a more mechanistic understanding of the local response to such changes. Re-
 809 cent reviews have discussed the energetic perspective of the ITCZ (Schneider et al., 2014)
 810 and its application to Earth’s monsoons (Biasutti et al., 2018), and so only a brief dis-
 811 cussion is given here.

812 The zonally averaged ITCZ latitude is strongly anticorrelated with the meridional
 813 atmospheric energy transport at the Equator, and correlated with the EFE latitude. This
 814 relation holds in both observations and under a range of modeled forcing scenarios, al-
 815 though it breaks down where the ITCZ shifts far from the Equator over the seasonal cy-
 816 cle (Adam et al., 2016b; Bischoff & Schneider, 2014; Donohoe et al., 2013). This rela-
 817 tionship helps to explain why the annual mean position of the zonal mean ITCZ is north
 818 of the Equator (Marshall et al., 2014).

819 Extending this framework to local cases has proved more challenging. Adam, Bischoff,
 820 and Schneider (2016a) defined the zonally varying EFE as the latitude at which the merid-
 821 ionally divergent column integrated MSE flux vanishes and has positive meridional gra-
 822 dient. This was found to approximate the seasonal cycle of ITCZ migrations over Africa,
 823 Asia and the Atlantic. However, the influence of the Walker cell limited the local EFE’s
 824 usefulness over the Pacific, and the EFE deviates from the ITCZ in the solstitial seasons
 825 that are particularly relevant to the monsoons. Boos and Korty (2016) took this a step
 826 further, using the longitudes where the zonally divergent column integrated MSE flux
 827 vanishes, and has positive zonal gradient, to define ‘Energy Flux Prime Meridians’ (EF-
 828 PMs). Two EFPMs can be identified in each season: over the Bay of Bengal and Gulf
 829 of Mexico/Caribbean Sea in boreal summer, and over the Western Pacific and South Amer-
 830 ica in austral summer. They showed that this extended theory gives some basic insight
 831 into how localized shifts in precipitation with ENSO relate to anomalous energy trans-
 832 ports.

833 As with the momentum budget framework, while the EFE framework is valuable
 834 in explaining some features of the overturning circulation, limitations must be remem-
 835 bered. Relating changes in the latitude of the ITCZ to that of the zonally averaged EFE
 836 assumes that the response to forcing is via changes to the meridional overturning cir-
 837 culation, and neglects changes to the gross moist stability. Such changes have been shown
 838 to be non-negligible both over the seasonal cycle and in the response to orbital and green-
 839 house gas forcings (Merlis et al., 2013; Seo, Kang, & Merlis, 2017; Smyth, Hill, & Ming,
 840 2018; Wei & Bordoni, 2018). In addition, Biasutti et al. (2018) noted that while the EFE
 841 predicts changes to the ITCZ latitude once the net energy imbalance is known, changes
 842 in ocean energy transport, and feedbacks internal to the atmosphere, can result in a net
 843 imbalance different to that expected from an imposed external forcing, including orbital
 844 forcing (Liu, Battisti, & Donohoe, 2017).

845 *3.2.3 Reconciling dynamics and energetics perspectives*

846 The two perspectives discussed so far in this review have emerged via separate con-
 847 sideration of the momentum and energy budgets, and have yet to be fully connected. Com-
 848 mon to both is consideration of processes that can alter the distribution of MSE either
 849 in the boundary layer or in a vertically integrated sense, and this might provide a bridge
 850 to fill the gaps between these two frameworks.

851 The local, vertically integrated MSE budget has long been used in explaining the
 852 distribution of tropical precipitation. Chou and Neelin (2001) and Chou and Neelin (2003)
 853 analysed the column integrated MSE budget in the South American and North Amer-
 854 ican, Asian and African monsoon regions respectively. They identified three key processes
 855 governing the MSE distribution and thus determining the extent of tropical rainfall over
 856 land: advection of high or low MSE air into the region, soil-moisture feedbacks, and the
 857 interaction between the convergence zone and the Rossby wave induced subsidence, which
 858 occurs to the west of monsoon heating (the interactive Rodwell-Hoskins mechanism; see
 859 Rodwell and Hoskins (2001)). The column integrated MSE budget has also allowed in-
 860 vestigation of the mechanisms determining the differing responses of models to intuitively
 861 similar forcing scenarios (e.g D’Agostino, Bader, Bordoni, Ferreira, & Jungclaus, 2019),
 862 and the different responses of model variants to the same forcing (e.g Hill, Ming, Held,
 863 & Zhao, 2017; Hill, Ming, & Zhao, 2018).

864 Provided CQE holds, so that the tropical atmosphere is near a moist neutral state,
 865 the horizontal distribution of column integrated moist static energy will be strongly tied
 866 to the distribution of subcloud moist static energy: for example, in Fig. 7 it can be seen
 867 that precipitation lies just equatorward of the column integrated MSE. This may allow
 868 connections to be made between the constraints arising from the momentum and ener-

869 getic frameworks, at least in the zonal mean. However, a unified theory for monsoon cir-
 870 culations remains an outstanding challenge (e.g. Biasutti et al., 2018; Hill, 2019).

871 **4 Beyond the aquaplanet perspective**

872 The theories that have emerged from the aquaplanet perspective have begun to prove
 873 useful in interpreting the climatology and variability of the tropical monsoon systems
 874 on both regional and global scales, particularly where their dynamics show similarities
 875 to that of the ITCZ in an aquaplanet. Synthesising idealized modeling work with obser-
 876 vational and realistic modeling studies suggests a picture that is consistent with a view
 877 of the monsoons and ITCZs as local migrations of the tropical convergence zone:

- 878 1. In the zonal mean, the ITCZ latitude is set by energetic constraints (Fig. 9).
- 879 2. Locally and seasonally, the convergence zone location appears governed by the MSE
 880 distribution, which can be understood via the regional MSE budget (Fig. 11).
- 881 3. When the ITCZ is near the Equator, the overturning circulation is strongly in-
 882 fluenced by extratropical eddies. Once it is far from the Equator, the cross-equatorial
 883 (winter) Hadley cell may approach an angular momentum conserving regime (Fig.
 884 12).
- 885 4. It may be possible to categorize the dynamics of the various convergence zones
 886 as ‘monsoons’ and ‘ITCZs’ by their circulation regime, e.g. Figs. 6 and 13, and
 887 so deduce where and when extratropical eddies vs. thermal forcing are likely to
 888 exert a stronger or weaker control over regional rainfall.
- 889 5. Regional variability in monsoon precipitation on interannual timescales (and per-
 890 haps subseasonal timescales) appears related to local variations in MSE which, where
 891 CQE applies, is connected to variations in the Hadley circulation.
- 892 6. Global variability in the latitude of the zonal mean convergence zone on interdecadal
 893 and longer timescales is driven by variations in the hemispheric energy budgets,
 894 with consequences for regional monsoon rainfall.

895 However, there are important influences on the regional monsoons and ITCZs that
 896 are not well accounted for by the above, in particular, the roles of zonal asymmetries and
 897 transient activity in the tropics.

898 **4.1 Asymmetries in the boundary conditions**

899 Zonal asymmetries, such as land-sea contrast, orography, and the ocean circula-
 900 tion, introduce complications unaccounted for by the simple aquaplanet framework. One
 901 key feature of the regional monsoons that cannot be captured by the symmetric picture
 902 is the presence of the subtropical frontal zones that occur on the eastern side of the mon-
 903 soon lows: the Meiyu-Baiu frontal zone, the South Pacific Convergence Zone (SPCZ),
 904 the South Atlantic Convergence Zone, and the South Indian Convergence Zone, which
 905 extends off the southeast coast of Southern Africa (Cook, 2000; Kodama, 1992). In par-
 906 ticular, the monsoons of East Asia and South America require us to step beyond the per-
 907 spective of angular momentum conserving monsoons and eddy-driven ITCZs.

908 **4.1.1 East Asia - a frontal monsoon**

909 While the South Asian monsoon fits well with the theoretical paradigm emerging
 910 from idealized work, the circulation over East Asia behaves very differently. Here, wind
 911 reversal is predominantly meridional, and monsoon precipitation extends north into the
 912 subtropics (Fig. 10). Summer precipitation is concentrated into the Meiyu-Baiu front,
 913 a rainband which forms on the northern boundary of the high MSE air mass centered
 914 over South Asia and the Bay of Bengal (Ding & Chan, 2005, and references therein). This
 915 front migrates northward in steps over the summer season, as detailed in Section 3.1.

916 Unlike in tropical monsoon regions, in the Meiyu-Baiu region the net energy in-
 917 put into the atmospheric column is negative. Vertical upward motion and convection in
 918 the front (with associated energy export) require MSE convergence, which is provided
 919 by horizontal advection, with interactions between the Tibetan Plateau and the west-
 920 erly jet playing a key role (Chen & Bordoni, 2014; J. Chiang, Wu, Kong, & Battisti, 2020;
 921 Molnar, Boos, & Battisti, 2010; Sampe & Xie, 2010). Comparing the monsoon season
 922 precipitation in this region in numerical experiments with and without the Tibetan Plateau
 923 indicates that, when the plateau is removed, precipitation is weakened and is no longer
 924 focused into the front (Chen & Bordoni, 2014). Analysis of the MSE budget of these sim-
 925 ulations suggests that the plateau chiefly reinforces convergence into the Meiyu-Baiu re-
 926 gion by strengthening the southerly stationary eddy downstream. The subtropical west-
 927 erly jet off the eastern flank of the Plateau additionally appears to act as an anchor for
 928 transient precipitating weather systems, focussing precipitation along the front (Molnar
 929 et al., 2010; Sampe & Xie, 2010).

930 Over the summer season, the East Asian Summer monsoon features two abrupt north-
 931 ward jumps of the precipitation, with three stationary periods (Ding & Chan, 2005). This
 932 intraseasonal evolution of the monsoon has also been suggested to relate to interactions
 933 between the plateau and westerly jet, with the migration of westerlies from the south
 934 of the plateau to the north causing the first abrupt jump and the development of Meiyu,
 935 and the northward migration of westerlies away from the plateau causing the second (Kong
 936 & Chiang, 2020; Molnar et al., 2010). A series of recent papers has examined implica-
 937 tions of this interaction for interpretation of changes to the East Asian summer mon-
 938 soon over the paleoclimate record (J. C. H. Chiang et al., 2015) and the Holocene (Kong,
 939 Swenson, & Chiang, 2017), and for interannual variability of the East Asian summer mon-
 940 soon (J. C. H. Chiang, Swenson, & Kong, 2017), with the hypothesis appearing able to
 941 explain all cases.

942 *4.1.2 South America - a zonal monsoon*

943 Similarities have been noted between the South American and East Asian mon-
 944 soons; however, studies indicate that diabatic heating over land is most important in gen-
 945 erating the upper-level monsoon anticyclone over South America (Lenters & Cook, 1997).
 946 One important difference is that the Andes form a narrow, meridionally oriented bar-
 947 rier from the tropics to subtropics. This acts to divert the easterly flow from the Atlantic
 948 to the south, concentrating it into the South American Low Level Jet (Byerle & Pae-
 949 gle, 2002; Campetella & Vera, 2002) and inducing adiabatic ascent (Rodwell & Hoskins,
 950 2001). In austral summer, the result is a zonally convergent mass flux of similar mag-
 951 nitude to the meridionally convergent component (Fig. 13), which extends the summer
 952 precipitation southward.

953 *4.1.3 The Pacific and Atlantic ITCZs*

954 Energy and momentum considerations are useful in understanding the behavior of
 955 the zonal mean ITCZ, especially in an aquaplanet where the boundary conditions are
 956 symmetric. However, to understand the regional positions and seasonal cycle of the ITCZ,
 957 the effects of zonal asymmetries must also be considered.

958 Except for in the far western tropical Atlantic where the ITCZ dips slightly south
 959 of the Equator in March and April, the ITCZ is north of the Equator year-round in the
 960 Atlantic and Pacific (Fig. 8). This appears to be due to the land monsoon heating and
 961 the geometrical asymmetry in tropical Africa and South America (Rodwell & Hoskins,
 962 2001). Specifically, the austral summer monsoon in southern Africa forces subsidence to
 963 the west and causes a subtropical high to build over the southern subtropical tropical
 964 Atlantic, increasing the southeasterly trade winds which act to cool the ocean by enhanced
 965 turbulent energy fluxes. Together, the subsidence and cool water suppress convection

966 south of the Equator in the austral summer and fall. In addition, in boreal summer the
 967 north African monsoon forces a strong local Hadley circulation that also causes subsi-
 968 dence in the sub-tropical south Atlantic which supports the formation of stratus clouds
 969 that further cool the ocean during austral winter. Hence, the ITCZ does not transit into
 970 the Southern Hemisphere in austral summer.

971 Subsidence is also seen off the west coast of South America in austral summer. This
 972 descent can be attributed to a range of factors. SSTs over the western coast of South
 973 America are cooler due to coastal upwelling (e.g. Takahashi, 2005), but this cooling is
 974 largely confined to within 50km of the coast. Equatorward flow is generated to the west
 975 in response to heating over the Amazon (Rodwell & Hoskins, 2001), which descends adi-
 976 abatically. The extratropical mid-level westerlies incident on the Andes are also diverted
 977 equatorward, contributing to further descent and evaporative cooling of the ocean by the
 978 dry subsiding air (e.g. Fig. 8; Rodwell & Hoskins, 2001; Takahashi & Battisti, 2007). The
 979 large-scale descent causes an inversion to form that allows for the development of large-
 980 scale stratus clouds that cool the ocean for thousands of kilometers offshore, which sup-
 981 presses convection over the eastern Pacific, particularly in austral summer. The large-
 982 scale descent forced by the Andes causes the Pacific ITCZ to be located exceptionally
 983 far north of the Equator throughout the year (Maroon, Frierson, & Battisti, 2015; Taka-
 984 hashi & Battisti, 2007). It may also partially account for the large seasonal contrast in
 985 precipitation in the North American monsoon, which involves an eastward extension of
 986 the Pacific ITCZ (Figs. 1 and 13).

987 4.2 The role of transients

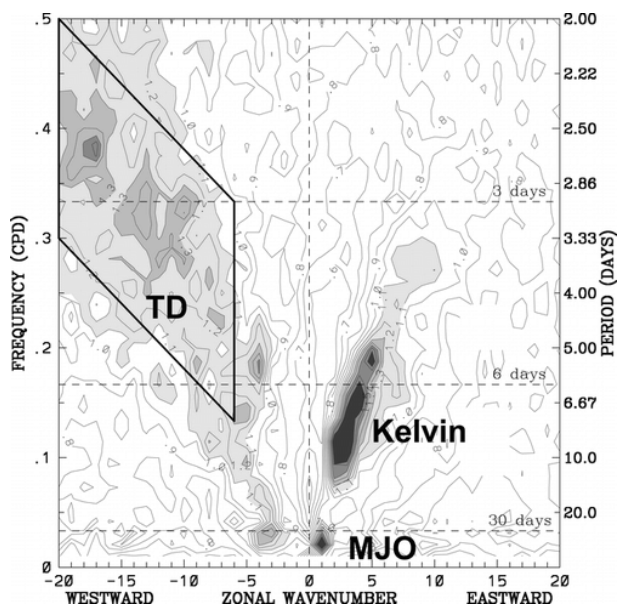


Figure 14. Wavenumber-frequency power spectrum of the symmetric component of OLR for June-August 1979-2003, averaged from 15°N to 15°S, plotted as the ratio of the raw OLR spectrum against a smooth red noise background (see Wheeler & Kiladis, 1999, for details). Contour interval is 0.1. Shading begins at 1.1, where the signal is statistically significant at approximately the 95% level. Peaks associated with the MJO, tropical depressions, and Kelvin waves are identified. Taken from Kiladis et al. (2006). ©American Meteorological Society. Used with permission.

988 Even in an aquaplanet, tropical rainfall does not occur in a zonally uniform, con-
 989 tinuously raining band. For simplicity, theoretical studies like those discussed in Section
 990 2 tend to consider time and zonal means and neglect transient activity except for its con-
 991 tribution to the momentum and energy budgets via eddy fluxes from the extratropics.
 992 However, many types of transient activity occur in the tropics. Wheeler and Kiladis (1999)
 993 produced wavenumber-frequency spectra of tropical outgoing longwave radiation (OLR),
 994 which is used as a proxy for deep convection, and showed that the spectral peaks that
 995 emerge are similar to wave modes of the shallow water equations on the beta plane (Mat-
 996 suno, 1966), providing clear evidence for a strong influence of convectively coupled waves
 997 on tropical precipitation.

998 Fig. 14 shows a Wheeler-Kiladis wavenumber-frequency spectrum for Northern Hemi-
 999 sphere summer (Kiladis et al., 2006). In this season, the spectrum of the symmetric com-
 1000 ponent of tropical OLR exhibits three dominant peaks: eastward propagating waves as-
 1001 sociated with the Madden-Julian Oscillation (MJO) and Kelvin waves, and westward prop-
 1002 agating waves classed as tropical depressions. Here we focus on the organisation of pre-
 1003 cipitation by these features, followed by a discussion of the transients observed in the
 1004 ITCZs.⁷

1005 *4.2.1 Intraseasonal timescales*

1006 Fig. 14 shows intense activity associated with low wavenumbers and a period of
 1007 30-60 days. This has been shown to correspond to the MJO; a convectively coupled, large-
 1008 scale equatorially trapped wave that propagates slowly eastward from the east coast of
 1009 Africa to the western-central Pacific, whereafter it continues eastward as a Kelvin wave
 1010 (Madden & Julian, 1971, 1972; C. Zhang, 2005, and references therein). The structure
 1011 of the MJO is illustrated in Fig. 15. The oscillation has strong influences on tropical rain-
 1012 fall, particularly in the Indo-Pacific region (see examples below), but the precise mech-
 1013 anism responsible remains a topic of ongoing research.

1014 *The Madden Julian Oscillation and the tropical Indian Ocean ‘ITCZ’*

1015 Precipitation in the Indian Ocean sector in austral summer is found between 10°N
 1016 and 15°S, but is concentrated slightly south of the Equator. It can be seen from Fig. 8
 1017 that, unlike the Atlantic and Pacific ITCZs, precipitation in the Indian Ocean is not or-
 1018 ganized into a narrow zonal band, and it is due to a different physics than is described
 1019 in Section 2.1.3 (the zonal asymmetry in SST is insufficient to drive a symmetrically un-
 1020 stable flow). Estimates show that between 30 and 40% of the annual precipitation in the
 1021 Indian Ocean and Maritime continent (10°N and 10°S, 70°E and 150°E) is associated
 1022 with the MJO (Kerns & Chen, 2020).

1023 *Intraseasonal variability in the Indo-Pacific*

1024 In the Indo-Pacific region, the MJO features trailing Rossby waves with enhanced
 1025 shear zones that angle polewards and westwards from the precipitation center near the
 1026 Equator and support precipitation. Hence, along a fixed longitude, bands of precipita-
 1027 tion appear to propagate poleward from the Equator to about 20°N over India as the
 1028 MJO propagates eastward over the maritime continent (Hartmann & Michelsen, 1989;
 1029 Wallace, Battisti, Thompson, & Hartmann, 2020).

1030 In boreal summer, in addition to the MJO, the climate in this region appears to
 1031 be modulated by propagating ‘Boreal Summer Intraseasonal Oscillations’ (BSISO), ob-
 1032 served to have dominant timescales of 10-20 and 30-60 days, and to propagate northward

⁷ A more detailed discussion of equatorial waves can be found in Roundy and Frank (2004), who de-
 velop a climatology, and in a review of the subject by Kiladis, Wheeler, Haertel, Straub, and Roundy
 (2009).

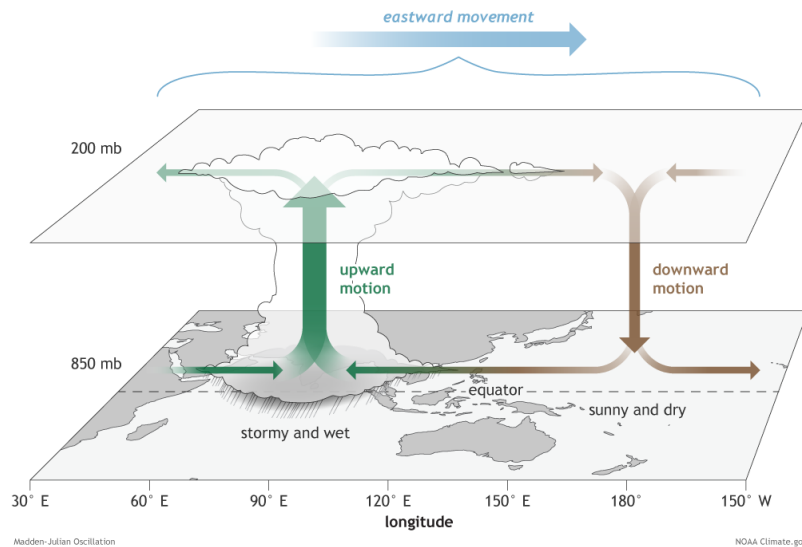


Figure 15. The surface and upper-atmosphere structure of the MJO for a period when the enhanced convective phase is centered across the Indian Ocean and the suppressed convective phase is centered over the west-central Pacific Ocean. Horizontal arrows pointing left represent wind departures from average that are easterly, and arrows pointing right represent wind departures from average that are westerly. The entire system shifts eastward over time. From NOAA Climate.gov (Gottschalck, 2014), drawing by Fiona Martin.

1033 over the continent (Annamalai & Slingo, 2001; Goswami & Ajaya Mohan, 2001; Hartmann & Michelsen, 1989; Lee et al., 2013). These oscillations modulate the active and
 1034 break phases of the Indian monsoon, with the tropical convergence zone and associated
 1035 Hadley circulation oscillating between an off-equatorial ‘monsoon’ location, and a near
 1036 equatorial ‘ITCZ’ location (e.g. Annamalai & Slingo, 2001; Goswami & Ajaya Mohan,
 1037 2001; D. R. Sikka & Gadgil, 1980). Like the MJO, the propagation mechanism and precise
 1038 drivers of the BSISOs remain unclear, and are the subject of ongoing research. Some
 1039 authors argue that the BSISOs are distinct from the MJO (e.g. Lee et al., 2013; B. Wang
 1040 & Xie, 1997), while others identify them as associated with the MJO (e.g. Hartmann &
 1041 Michelsen, 1989; Jiang, Adames, Zhao, Waliser, & Maloney, 2018).

1043 4.2.2 Monsoon lows, depressions & easterly waves

1044 In addition to the influence of oscillations on intraseasonal timescales, regional monsoon
 1045 precipitation has long been observed to be organized by westward propagating synoptic-
 1046 scale low-pressure systems, including African Easterly Waves, observed over West Africa
 1047 (e.g. Burpee, 1974; Reed, Norquist, & Recker, 1977), and monsoon depressions, observed
 1048 in the Indian and Australian monsoon regions (e.g. Godbole, 1977; Mooley, 1973; D. Sikka,
 1049 1978). Hurley and Boos (2015) produced a global climatology of monsoon lows. They
 1050 found that the behavior over India, the western Pacific and northern Australia showed
 1051 strong similarities, with a deep warm-over-cold core (e.g. Fig. 16a). A second class of
 1052 systems was seen over West Africa and western Australia, with a shallower warm core
 1053 (e.g. Fig. 16b). They estimated that monsoon low-pressure systems are responsible for
 1054 at least 40% of precipitation in monsoon regions (Fig. 16c,d).

1055 Fig. 16 confirms that the monsoon depressions observed over South Asia are distinct
 1056 from the easterly waves observed over Africa. The mechanisms for both are still

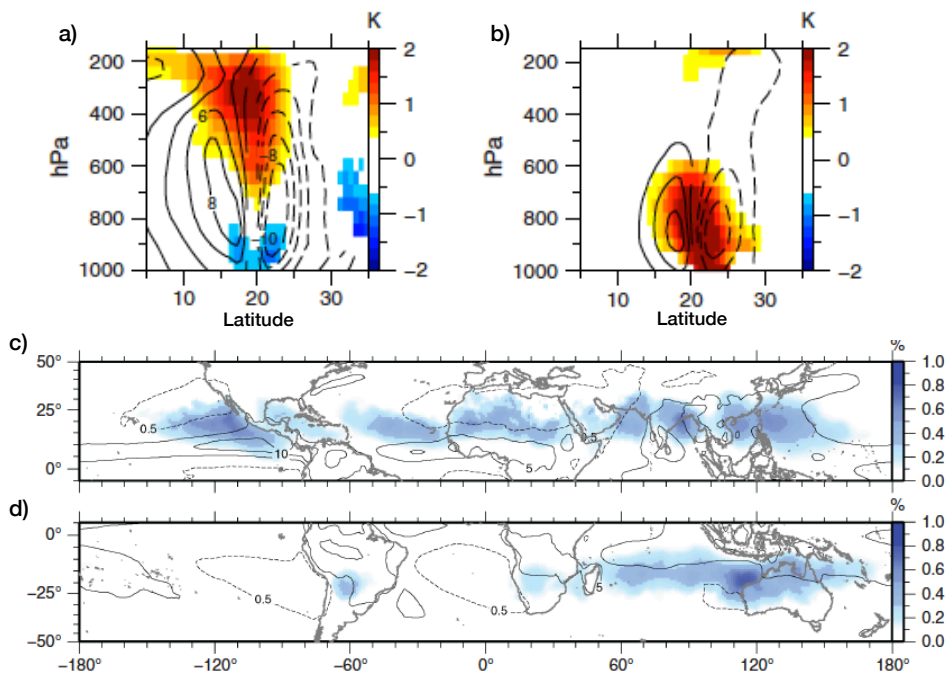


Figure 16. Northern Hemisphere summer (May-September) regional composites of monsoon depressions from ERA-Interim (1979-2012). Composite vertical sections through the storm center of potential temperature (K, shading) and zonal wind (ms^{-1} , contours) anomalies are shown for (a) India and (b) West Africa. Dashed contours are negative. Values are shaded or contoured where a t-test indicates significance at the 5% level. (c) and (d) show the fraction (shading) of total summer precipitation that can be attributed to monsoon lows and monsoon depressions in May-September and November-March respectively. Shading indicates the ratio of precipitation within 500 km of all tracked lows and depressions to the total summer precipitation. Contours reflect the summer climatological precipitation rate. Dashed contours surround dry regions, where precipitation is on average less than 0.5 mm/day. Solid contours indicate wet regions, where precipitation is greater than 5 mm/day (5 mm/day contour interval). Adapted from Figs. 9 and 12 of Hurley and Boos (2015). ©2014 Royal Meteorological Society. Used with permission.

1057 being investigated. Recent work indicates that monsoon depressions form over South Asia
 1058 from moist barotropic instability due to the meridional shear of the monsoon trough, and
 1059 are intensified by latent heating (Diaz & Boos, 2019a, 2019b). The occurrence of depres-
 1060 sions can be modulated by variations in the background flow. For example, ENSO causes
 1061 large-scale changes in the summertime environment that has a modest statistical effect
 1062 on the strength of synoptic scale tropical depressions that propagate from the Bay of Ben-
 1063 gal to the northwest over India (Hunt, Turner, Inness, Parker, & Levine, 2016), with La
 1064 Niña (El Niño) conditions favoring tropical depressions with enhanced (weakened) pre-
 1065 cipitation.

1066 Over Africa and the Atlantic, strong surface heating of the Sahara in summer forces
 1067 a monsoon circulation that is barotropically and baroclinically unstable (Burpee, 1972;
 1068 M.-L. C. Wu, Reale, Schubert, Suarez, & Thorncroft, 2012, and references therein). In-
 1069 stabilities in the flow give rise to African Easterly Waves. The precise dynamics govern-
 1070 ing the amplification, propagation and variability of these waves remain unclear. Sea-
 1071 sons with strong African Easterly Wave activity have been found to be associated with
 1072 a strong upper level easterly jet (Nicholson, 2009a) and an enhancement of other equa-

1073 torial waves, specifically Rossby and westward-moving mixed Rossby-gravity wave modes
 1074 (Y.-M. Cheng, Thorncroft, & Kiladis, 2019; Yang, Methven, Woolnough, Hodges, & Hoskins,
 1075 2018).

1076 **4.2.3 Transients in the ITCZ**

1077 Easterly waves and other organized synoptic disturbances are important contribu-
 1078 tors to the ITCZs in the tropical Atlantic and Pacific. An upper bound on their con-
 1079 tribution to ITCZ precipitation can be estimated by assuming that all precipitation events
 1080 lasting more than 24 hours are related to organized synoptic disturbances, in which case
 1081 the fraction of the total precipitation in the Atlantic and Pacific ITCZs that is due to
 1082 large-scale organized waves is about 40% (White, Battisti, & Skok, 2017). Alternatively,
 1083 about half of the total precipitation in these ITCZs is in the form of stratiform precip-
 1084 itation (Schumacher & Houze Jr, 2003), which is overwhelmingly in the form of long-lived
 1085 mesoscale convective systems (Houze Jr, 2018).

1086 African Easterly Waves propagate westward into the Atlantic Ocean and are the
 1087 primary precursors of tropical cyclones in the Atlantic. The associated rainfall contributes
 1088 to the summer precipitation in the Atlantic ITCZ. Easterly waves are also found in the
 1089 tropical east and central Pacific, although the dynamics of these systems is different from
 1090 their Atlantic counterparts. Many easterly waves in the Pacific are Mixed Rossby-Gravity
 1091 waves: antisymmetric equatorially trapped waves with low pressure centered on 5-10°
 1092 latitude (Kiladis et al., 2009; Matsuno, 1966). Friction acts to cause convergence in the
 1093 low pressure centers and in the Northern Hemisphere this leads to moisture convergence
 1094 and precipitation (Holton, Wallace, & Young, 1971; Liebmann & Hendon, 1990) (the low
 1095 pressure center in the Southern Hemisphere does not feature precipitation because the
 1096 water is cold and there is strong subsidence). Other convectively coupled equatorial waves
 1097 that contribute to the ITCZ in the central Pacific include Kelvin waves (in which con-
 1098 vection is not symmetric about the Equator) and inertio-gravity waves (Kiladis et al.,
 1099 2009).

1100 **5 Conclusions and outlook**

1101 This article has reviewed the theory of monsoons that has resulted in large part
 1102 from idealized models and discussed the behavior of Earth’s monsoons in light of the the-
 1103 ory. While the regional monsoons have a diverse range of individual features, they also
 1104 have much in common, including enhancement of cross-equatorial and westerly flow in
 1105 the summer season, rapid onset, and development in an off-equatorial direction. In ad-
 1106 dition, regional monsoons often covary as components of a global monsoon, both under
 1107 changes to orbital forcing, and to internal variations in the climate state such as ENSO
 1108 (P. X. Wang et al., 2014). The theoretical considerations outlined in Section 2 are start-
 1109 ing to provide explanations for these behaviors, as presented in Section 3, but many open
 1110 problems remain in connecting theoretical ideas to observations (Section 4). We conclude
 1111 the review by discussing these successes and challenges.

1112 **5.1 Successes**

1113 Insight from theory has caused a shift in the understanding of monsoon dynam-
 1114 ics – from that of primarily land-sea contrast driven, sea-breeze like circulations, to lo-
 1115 calized variations of the convergence zone and overturning circulations, strongly governed
 1116 by the momentum and energy budgets.

1117 The momentum budget, Eq. 4, indicates three classes of solution for the Hadley
 1118 circulation: a ‘radiative-convective equilibrium’ regime, $v = \omega = 0$; an ‘angular mo-
 1119 mentum conserving’ regime, in which the Rossby number Ro approaches 1 and eddies
 1120 have a negligible effect; and an ‘eddy-driven’ regime, where Ro is much less than 1 and

1121 eddies strongly influence the overturning circulation. Our understanding of monsoon dy-
 1122 namics has been greatly advanced by considering the transitions between these regimes,
 1123 and the controls on the latitude of the ascending branch of the circulation.

1124 Constraints on the zonal mean ITCZ latitude have been identified by considering
 1125 the energetics of the circulation, in addition to the momentum budget. If the atmosphere
 1126 is in CQE then, for an angular momentum conserving overturning circulation, the ITCZ
 1127 is expected to lie just equatorward of the peak in subcloud moist static energy (see Privé
 1128 and Plumb (2007a, 2007b), and Section 2.1). The subcloud distribution of MSE there-
 1129 fore strongly constrains the circulation. A related energetic constraint is obtained from
 1130 considering the vertically integrated MSE budget (Eq. 10). The latitude of the EFE has
 1131 been found to be approximately collocated with the ITCZ latitude, allowing the zonal
 1132 mean ITCZ location to be related to the meridional cross-equatorial energy flux and net
 1133 energy input at the Equator (e.g. Bischoff & Schneider, 2016).

1134 The latitude of the convergence zone is also related to the dynamics that govern
 1135 the Hadley circulation. It has been demonstrated in aquaplanets that when the conver-
 1136 gence zone is on or near the Equator the circulation is more eddy-driven (i.e., an ITCZ),
 1137 while when the convergence zone is far from the Equator the circulation is near angu-
 1138 lar momentum conserving (Bordoni & Schneider, 2008, 2010; Schneider & Bordoni, 2008).
 1139 These ‘*ITCZ*’ and ‘*monsoon*’ regimes are illustrated schematically in Figs. 6a and b re-
 1140 spectively. In aquaplanet simulations with a low thermal inertia surface, similar to land,
 1141 a fast transition between these two regimes is observed, with the zonal mean ITCZ jump-
 1142 ing off the Equator into the summer hemisphere at the start of the summer season. This
 1143 fast transition is mediated by two feedbacks. Firstly, as the ITCZ shifts off the Equa-
 1144 tor and the Hadley circulation becomes cross equatorial, the lower branch of the Hadley
 1145 cell advects cooler, drier air up the meridional MSE gradient. Combined with the con-
 1146 tinued diabatic warming of the summer hemisphere by the insolation, this has the ef-
 1147 fect of pushing the MSE peak poleward and so shifting the convergence zone farther off
 1148 the Equator (Fig. 7). Secondly, as a result of angular momentum conservation, the equa-
 1149 torward upper-level meridional flow gives rise to upper-level easterlies. These easterlies
 1150 suppress propagation of extratropical eddies into the low latitudes (Charney & Drazin,
 1151 1961) and help to kick the Hadley cell into the angular momentum conserving regime,
 1152 so that the meridional overturning is strongly responsive to the thermal forcing and strength-
 1153 ens and broadens further.

1154 Recent results suggest that in an aquaplanet, the transition between an eddy-driven
 1155 and angular momentum conserving Hadley circulation occurs when the ITCZ migrates
 1156 to $\sim 7^\circ$, regardless of slab ocean characteristics (Geen et al., 2019). In this review, we
 1157 have argued that the former case is relevant to the dynamics of the observed ITCZs, while
 1158 the latter is appropriate for understanding the monsoon circulations. Another recent strand
 1159 of research has explored the maximum limits on the migrations of the ITCZ away from
 1160 the Equator: in aquaplanets, the convergence zone does not migrate more than 30° pole-
 1161 ward, even when the MSE maximum is at the poles (Faulk et al., 2017). Current work
 1162 (Hill et al., 2019; Singh, 2019) is exploring this poleward limit of monsoons using con-
 1163 straints relating the Hadley circulation regime to the curvature of the subcloud equiv-
 1164 alent potential temperature.

1165 Analysis of observations has demonstrated that the South Asian, Australian and
 1166 African monsoons show behavior similar to that described by the above theoretical work.
 1167 In these monsoons, the peak precipitation is located just equatorward of the peak in sub-
 1168 cloud MSE (Nie et al., 2010) and the convergence zones migrate in line with the EFE
 1169 (Adam et al., 2016b; Boos & Korty, 2016). In monsoons where the ascending branch mi-
 1170 grates far from the Equator, such as the South Asian and South African monsoons, the
 1171 summertime overturning circulation becomes aligned with angular momentum contours,
 1172 suggesting a strongly thermally driven cross-equatorial flow regime (e.g. Bordoni & Schnei-
 1173 der, 2008; J. M. Walker & Bordoni, 2016). In addition, while a more thorough analy-

1174 sis is needed, we note from Fig. 13 that a change from an eddy-driven ('ITCZ') to an
 1175 angular momentum conserving ('monsoon') overturning regime indeed appears to occur
 1176 when ascent moves poleward of $\sim 10^\circ$ (Geen et al., 2019). For Earth's planetary rota-
 1177 tion and insolation characteristics, the tropics, and the tropical overturning circulations,
 1178 are confined within $\sim 30^\circ$ of the Equator (Faulk et al., 2017). Current work is explor-
 1179 ing conditions under which monsoon-like, cross-equatorial cell can extend further away
 1180 from the Equator and what sets their poleward boundary under these more exotic con-
 1181 ditions (Hill et al., 2019; Singh, 2019).

1182 Awareness of these mechanisms can help motivate work investigating sources of in-
 1183 terannual variability, and the response to external forcings, with one clear goal being a
 1184 better mechanistic understanding of model projections forced by future warming scenar-
 1185 ios. On this front, some success has already been achieved. For example, interannual vari-
 1186 ability in monsoon precipitation has been found to be correlated to variability in sub-
 1187 cloud MSE (Hurley & Boos, 2013). Migrations of the zonal mean ITCZ under histor-
 1188 ical forcings have been examined in relation to migrations of the EFE (Donohoe et al.,
 1189 2013). Interannual variability of the West African monsoon has been connected to the
 1190 strength of the upper-level easterly jet (Nicholson, 2009b). The weak changes to the Asian
 1191 monsoon in simulations of future climate appear to be explained by opposing responses
 1192 to increased CO_2 levels and surface warming (Shaw & Voigt, 2015). Further exploration
 1193 of the observations, informed by theory, could prove fruitful for improved understand-
 1194 ing of model biases, or for identifying sources of seasonal predictability.

1195 5.2 Challenges

1196 Constraints on the zonal and time mean ITCZ and overturning circulation are be-
 1197 ginning to emerge from theory and have now been successfully applied to aquaplanets
 1198 and to some features of the observations. This represents a significant step in our un-
 1199 derstanding of the tropical circulation. However, asymmetries that arise from land-sea
 1200 contrast and orography introduce a zoo of additional complications that these simple the-
 1201 ories do not account for, and some care must therefore be taken in applying aquaplanet
 1202 theories to reality. For example, while the monsoon circulation in an aquaplanet is char-
 1203 acterized by an angular momentum conserving Hadley circulation, stationary waves can
 1204 be important when zonal asymmetries are included in the boundary conditions (Shaw,
 1205 2014). However, as we show here in Fig. 12, in individual monsoon sectors (South Asia,
 1206 Africa and Australia) advection of momentum by the mean circulation causes the over-
 1207 turning circulation to approach an angular momentum conserving state.

1208 As discussed in Section 4, the pattern of precipitation in the South American mon-
 1209 soon and the intensity of the East Asian monsoon in particular are strongly influenced
 1210 by orography. The interaction of the westerly jet with the orography of Tibet generates
 1211 a stationary wave downstream over East Asia that gives rise to the Meiyu front and gov-
 1212 erns the duration of the stages of the East Asian summer monsoon. In South America,
 1213 the Andes divert the tropical easterly and subtropical westerly flow, resulting in strong
 1214 equatorward descending flow to the west of the mountains, and poleward ascending flow
 1215 to the east. In austral summer, the South American low-level jet develops to the east
 1216 of the Andes and extends the South American monsoon flow southward. This results in
 1217 precipitation that is displaced far from the Equator, but without the formation of an an-
 1218 gular momentum conserving Hadley cell of the kind seen in aquaplanets. The descend-
 1219 ing flow to the west of the Andes suppresses precipitation year-round off the coast of South
 1220 and Central America, over the East Pacific and helps to push the ITCZ north of the Equa-
 1221 tor year round. Overall, we conclude that aquaplanet theories do not appear applica-
 1222 ble to the American monsoons, or to the East Asian monsoon. Hierarchical modeling
 1223 work, where complexity is introduced in a progressive way, might be the path forward
 1224 to make conceptual progress in the understanding of similarities and differences between
 1225 individual monsoon systems. Initial steps on this hierarchy are already being taken, by

1226 introducing continents into idealized models (J. Chiang et al., 2020; Geen et al., 2018;
 1227 Hui & Bordoni, in prep.; Zhou & Xie, 2018), or removing orography from more complete
 1228 models (Baldwin, Vecchi, & Bordoni, 2019; Boos & Kuang, 2010; Wei & Bordoni, 2016).

1229 Transients make a non-negligible contribution to precipitation in the regional mon-
 1230 soons and ITCZs. These phenomena are not accounted for in the theoretical framework
 1231 reviewed in Section 2. In some cases, parallel ideas exist, which should be possible to syn-
 1232 thesize. For example, as discussed in Section 4, monsoon onset over South Asia and the
 1233 South China Sea has been suggested to relate to the arrival of the moist phase of a tran-
 1234 sient Intraseasonal Oscillation (ISO), with active and break phases throughout the sea-
 1235 son then arising due to further ISOs (e.g. Lee et al., 2013). Aquaplanet based model-
 1236 ing work has instead led to development of a zonal- and climatological-mean view of mon-
 1237 soon onset as a regime change of the Hadley circulation (see Section 2.1 and Bordoni &
 1238 Schneider, 2008; Schneider & Bordoni, 2008). These ideas appear tantalisingly recon-
 1239 cilable; for example the arrival of an ISO might act as the trigger for the regime change
 1240 of the circulation, and the active and break phases of the Indian monsoon could be con-
 1241 nected to intraseasonal changes in Hadley cell regime. However, evidence for this has yet
 1242 to be examined in idealized models or in reanalysis. The MSE budget has been used to
 1243 investigate the propagation of the MJO (Andersen & Kuang, 2012; Jiang et al., 2018;
 1244 Sobel & Maloney, 2013), and may provide a way to bridge these two perspectives.

1245 Perhaps the greatest challenge for theory and modeling is to determine how the mon-
 1246 soon systems will change in future climates. The current consensus from CMIP5 mod-
 1247 els is that the precipitation in the global monsoon is likely to increase under anthropogenic
 1248 forcings, though the monsoon circulation is likely to weaken (Christensen et al., 2013).
 1249 However, there is a significant spread in model projections (e.g. Seth et al., 2019, and
 1250 references therein), and models show varying degrees of skill in capturing the present-
 1251 day climatology of the monsoon and its variability (e.g. Jourdain et al., 2013; Roehrig,
 1252 Bouniol, Guichard, Hourdin, & Redelsperger, 2013; Sperber et al., 2013). As discussed
 1253 in Section 3.2, future predictability depends on direct and indirect responses to radia-
 1254 tive forcing, which may oppose one another (Shaw & Voigt, 2015). However, we suggest
 1255 that a more complete understanding of the constraints on the monsoons will provide op-
 1256 portunities to identify model biases affecting their simulation of present climate, and may
 1257 allow future projections to be better constrained.

1258 Glossary

- 1259 **AMIP** Atmospheric Model Intercomparison Project. A project comparing the behav-
 1260 iors of atmospheric general circulation models forced by realistic sea surface tem-
 1261 peratures and sea ice.
- 1262 **BSISO** Boreal Summer Intraseasonal Oscillation. Describes the dominant modes of trop-
 1263 ical intraseasonal variability over Asia during boreal summer.
- 1264 **CQE** Convective Quasi-Equilibrium. A theoretical framework for the tropical atmosphere
 1265 that assumes the atmospheric lapse rate is maintained close to a moist adiabat
 1266 due to the occurrence frequent, intense moist convection. See discussion in Sec-
 1267 tion 2.1.
- 1268 **Earth System model** A comprehensive model of the Earth System, simulating the
 1269 fluid motions and thermodynamics of the atmosphere and ocean, as well as inter-
 1270 actions with ice, the land surface and vegetation, and ocean biogeochemistry.
- 1271 **EFE** Energy Flux Equator. The latitude at which the vertically integrated MSE flux
 1272 is zero.
- 1273 **EFPM** Energy Flux Prime Meridian. Defined as the longitudes at which the zonally
 1274 divergent column integrated MSE flux vanishes and has positive zonal gradient.
- 1275 **ENSO** The El Niño-Southern Oscillation. A recurring climate pattern involving changes
 1276 to the temperature of the waters in the Pacific Ocean. El Niño (La Niña) phases

- 1277 are associated with warmer (cooler) than usual SSTs in the central and eastern
 1278 tropical Pacific Ocean.
- 1279 **GCM** Global Circulation Model. A numerical model for the circulation of the atmo-
 1280 sphere and/or ocean.
- 1281 **Heinrich event** A natural phenomenon featuring the collapse of Northern Hemisphere
 1282 ice shelves and consequently the release of large numbers of icebergs.
- 1283 **Idealized model** A model in which only some elements of the Earth System are in-
 1284 cluded to allow testing of theories in a more conceptually simple and computa-
 1285 tionally affordable framework.
- 1286 **ISO** Intra-seasonal Oscillation
- 1287 **ITCZ** Intertropical Convergence Zone. The upward branch of the overturning circu-
 1288 lation that is driven primarily by eddy momentum fluxes and located with $\sim 10^\circ$
 1289 of the Equator, and characterized by low-level moisture convergence and precip-
 1290 itation.
- 1291 **Monsoon** The rainy summer season of a tropical region, in which tropical precipita-
 1292 tion extends far from the Equator, the low-level prevailing wind changes direction
 1293 or strength, and the overturning circulation approaches the angular momentum
 1294 conserving (eddy-less) limit.
- 1295 **MJO** Madden-Julian Oscillation
- 1296 **MSE** Moist static energy, defined in Eq. 9.
- 1297 **RCE** Radiative-convective equilibrium. Describes the balance between the radiative cool-
 1298 ing of the atmosphere and the heating via latent heat release resulting from con-
 1299 vection.
- 1300 **Sea breeze** A wind that blows from a large body of water onto a landmass due to dif-
 1301 ferences in surface temperature, and consequently air pressure, between the land
 1302 and water.
- 1303 **SPCZ** South Pacific Convergence Zone. The band convergence observed over the South-
 1304 West Pacific, e.g. Fig. 1e.
- 1305 **SST** Sea Surface Temperature

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1313 References

- 1314 Adam, O., Bischoff, T., & Schneider, T. (2016a). Seasonal and interannual
 1315 variations of the energy flux equator and ITCZ. Part II: Zonally vary-
 1316 ing shifts of the ITCZ. *Journal of Climate*, 29(20), 7281–7293. doi:
 1317 10.1175/JCLI-D-15-0710.1
- 1318 Adam, O., Bischoff, T., & Schneider, T. (2016b). Seasonal and interannual varia-
 1319 tions of the energy flux equator and ITCZ. Part I: Zonally averaged ITCZ po-
 1320 sition. *Journal of Climate*, 29(9), 3219–3230. doi: 10.1175/JCLI-D-15-0512.1
- 1321 Adams, D. K., & Comrie, A. C. (1997). The North American Monsoon. *Bulletin*
 1322 *of the American Meteorological Society*, 78(10), 2197–2214. doi: 10.1175/1520-
 1323 -0477(1997)078(2197:TNAM)2.0.CO;2
- 1324 An, Z., Guoxiong, W., Jianping, L., Youbin, S., Yimin, L., Weijian, Z., . . . Juan,
 1325 F. (2015). Global Monsoon Dynamics and Climate Change. *Annual*

- 1326 *Review of Earth and Planetary Sciences*, 43(1), 29–77. doi: 10.1146/
 1327 annurev-earth-060313-054623
- 1328 Ananthakrishnan, R., & Soman, M. K. (1988). The Onset of the Southwest Mon-
 1329 soon over Kerala: 1901-1980. *Journal of Climatology*, 8, 283–296. doi: 10
 1330 .1002/joc.3370080305
- 1331 Andersen, J. A., & Kuang, Z. (2012). Moist Static Energy Budget of MJO-like Dis-
 1332 turbances in the Atmosphere of a Zonally Symmetric Aquaplanet. *Journal of*
 1333 *Climate*, 25(8), 2782-2804. Retrieved from <https://doi.org/10.1175/JCLI-D-11-00168.1> doi: 10.1175/JCLI-D-11-00168.1
- 1334 Annamalai, H., & Slingo, J. M. (2001). Active/break cycles: Diagnosis of the in-
 1335 traseasonal variability of the Asian Summer Monsoon. *Climate Dynamics*,
 1336 18(1), 85–102. Retrieved from <https://doi.org/10.1007/s003820100161>
 1337 doi: 10.1007/s003820100161
- 1338 Arakawa, A., & Schubert, W. H. (1974). Interaction of a Cumulus Cloud Ensemble
 1339 with the Large-Scale Environment, Part I. *J. Atmos. Sci.*, 31, 674–701. doi: 10
 1340 .1175/1520-0469(1974)031%3C0674:IOACCE%3E2.0.CO;2
- 1341 Arbuszewski, J. A., Demenocal, P. B., Cléroux, C., Bradtmiller, L., & Mix, A.
 1342 (2013). Meridional shifts of the Atlantic Intertropical Convergence Zone
 1343 since the Last Glacial Maximum. *Nature Geoscience*, 6(11), 959–962. doi:
 1344 10.1038/ngeo1961
- 1345 Back, L. E., & Bretherton, C. S. (2009). On the relationship between SST gradi-
 1346 ents, boundary layer winds, and convergence over the tropical oceans. *Journal*
 1347 *of Climate*, 22(15), 4182–4196.
- 1348 Baldwin, J. W., Vecchi, G. A., & Bordoni, S. (2019). The direct and ocean-mediated
 1349 influence of Asian orography on tropical precipitation and cyclones. *Climate*
 1350 *Dynamics*, 1–20. Retrieved from <http://dx.doi.org/10.1007/s00382-019-04615-5>
 1351 -04615-5 <http://link.springer.com/10.1007/s00382-019-04615-5> doi: 10
 1352 .1007/s00382-019-04615-5
- 1353 Barlow, M., Nigam, S., & Berbery, E. H. (1998). Evolution of the North American
 1354 Monsoon System. *Journal of Climate*, 11(9), 2238–2257. doi: 10.1175/1520
 1355 -0442(1998)011(2238:EOTNAM)2.0.CO;2
- 1356 Battisti, D. S., Ding, Q., & Roe, G. H. (2014). Coherent pan-Asian climatic and iso-
 1357 topic response to orbital forcing of tropical insolation. *Journal of Geophysical*
 1358 *Research*, 119(21), 11,997–12,020. doi: 10.1002/2014JD021960
- 1359 Battisti, D. S., Vimont, D. J., & Kirtman, B. P. (2019). 100 years of progress in un-
 1360 derstanding the dynamics of coupled atmosphere-ocean variability. *Meteorolog-
 1361 ical Monographs*. doi: 10.1175/AMSMONOGRAPHS-D-18-0025.1
- 1362 Becker, E., Schmitz, G., & Geprägs, R. (1997). The feedback of midlatitude waves
 1363 onto the Hadley cell in a simple general circulation model. *Tellus A*, 49, 182–
 1364 199.
- 1365 Biasutti, M., Voigt, A., Boos, W. R., Braconnot, P., Hargreaves, J. C., Harrison,
 1366 S. P., ... Xie, S.-P. (2018). Global energetics and local physics as drivers of
 1367 past, present and future monsoons. *Nature Geoscience*, 11(6), 392–400. doi:
 1368 10.1038/s41561-018-0137-1
- 1369 Bischoff, T., & Schneider, T. (2014). Energetic constraints on the position of the In-
 1370 tertropical Convergence Zone. *Journal of Climate*, 27(13), 4937–4951. doi: 10
 1371 .1175/JCLI-D-13-00650.1
- 1372 Bischoff, T., & Schneider, T. (2016). The equatorial energy balance, ITCZ position,
 1373 and double-ITCZ bifurcations. *Journal of Climate*, 29(8), 2997–3013. doi: 10
 1374 .1175/JCLI-D-15-0328.1
- 1375 Bond, G., Broecker, W., Johnsen, S., McManus, J., Labeyrie, L., Jouzel, J., & Bo-
 1376 nani, G. (1993). Correlations between climate records from North Atlantic
 1377 sediments and Greenland ice. *Nature*, 365, 143-147.
- 1378 Boos, W. R., & Korty, R. L. (2016). Regional energy budget control of the In-
 1379 tertropical Convergence Zone and application to mid-Holocene rainfall. *Nature*

- 1381 *Geoscience*, 9(12), 892–897. doi: 10.1038/ngeo2833
- 1382 Boos, W. R., & Kuang, Z. (2010). Dominant control of the South Asian monsoon by
1383 orographic insulation versus plateau heating. *Nature*, 463(7278), 218–222. doi:
1384 10.1038/nature08707
- 1385 Bordoni, S., Ciesielski, P. E., Johnson, R. H., McNoldy, B. D., & Stevens, B.
1386 (2004). The low-level circulation of the North American Monsoon as re-
1387 vealed by QuikSCAT. *Geophysical Research Letters*, 31, L10109. doi:
1388 10.1029/2004GL020009
- 1389 Bordoni, S., & Schneider, T. (2008). Monsoons as eddy-mediated regime transitions
1390 of the tropical overturning circulation. *Nature Geoscience*, 1(8), 515–519. doi:
1391 10.1038/ngeo248
- 1392 Bordoni, S., & Schneider, T. (2010). Regime Transitions of Steady and Time-
1393 Dependent Hadley Circulations: Comparison of Axisymmetric and Eddy-
1394 Permitting Simulations. *Journal of the Atmospheric Sciences*, 67(5), 1643–
1395 1654. doi: 10.1175/2009JAS3294.1
- 1396 Broccoli, A. J., Dahl, K. A., & Stouffer, R. J. (2006). Response of the ITCZ to
1397 Northern Hemisphere cooling. *Geophysical Research Letters*, 33(1), 1–4. doi:
1398 10.1029/2005GL024546
- 1399 Burpee, R. W. (1972). The origin and structure of easterly waves in the lower tropo-
1400 sphere of North Africa. *Journal of the Atmospheric Sciences*, 29(1), 77–90.
- 1401 Burpee, R. W. (1974). Characteristics of North African easterly waves during the
1402 summers of 1968 and 1969. *Journal of the Atmospheric Sciences*, 31(6), 1556–
1403 1570.
- 1404 Byerle, L. A., & Paegle, J. (2002). Description of the Seasonal Cycle of Low-Level
1405 Flows Flanking the Andes and their Interannual Variability. *Meteorologica*, 27,
1406 71–88.
- 1407 Campetella, C. M., & Vera, C. S. (2002). The influence of the Andes mountains on
1408 the South American low-level flow. *Geophysical Research Letters*, 29(17), 7:1–
1409 4. doi: 10.1029/2002gl015451
- 1410 Carolin, S. A., Cobb, K. M., Adkins, J. F., Clark, B., Conroy, J. L., Lejau, S., ...
1411 Tuen, A. A. (2013). Varied Response of Western Pacific Hydrology to Climate
1412 Forcings over the Last Glacial Period. *Science*, 340(6140), 1564–1566. doi:
1413 10.1126/science.1233797
- 1414 Charney, J. G., & Drazin, P. G. (1961). Propagation of planetary-scale disturbances
1415 from the lower into the upper atmosphere. *Journal of Geophysical Research*,
1416 66(1), 83–109. Retrieved from [https://agupubs.onlinelibrary.wiley.com/
1417 doi/abs/10.1029/JZ066i001p00083](https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/JZ066i001p00083) doi: 10.1029/JZ066i001p00083
- 1418 Chen, J., & Bordoni, S. (2014). Orographic Effects of the Tibetan Plateau on the
1419 East Asian Summer Monsoon: An Energetic Perspective. *Journal of Climate*,
1420 27(8), 3052–3072. doi: 10.1175/JCLI-D-13-00479.1
- 1421 Cheng, H., Sinha, A., Wang, X., Cruz, F. W., & Edwards, R. L. (2012). The Global
1422 Paleomonsoon as seen through speleothem records from Asia and the Ameri-
1423 cas. *Climate Dynamics*, 39, 1045–1062. doi: 10.1007/s00382-012-1363-7
- 1424 Cheng, Y.-M., Thorncroft, C. D., & Kiladis, G. N. (2019). Two contrasting African
1425 easterly wave behaviors. *Journal of the Atmospheric Sciences*, 76(6), 1753–
1426 1768.
- 1427 Chiang, J., Wu, C., Kong, W., & Battisti, D. (2020). Origins of East Asian Summer
1428 Monsoon Seasonality. *Revised, Journal of Climate*.
- 1429 Chiang, J. C., & Friedman, A. R. (2012). Extratropical cooling, interhemispheric
1430 thermal gradients, and tropical climate change. *Annual Review of Earth and
1431 Planetary Sciences*, 40.
- 1432 Chiang, J. C. H., & Bitz, C. M. (2005). Influence of high latitude ice cover on the
1433 marine Intertropical Convergence Zone. *Climate Dynamics*, 25(5), 477–496.
1434 doi: 10.1007/s00382-005-0040-5
- 1435 Chiang, J. C. H., Fung, I. Y., Wu, C.-H., Cai, Y., Edman, Jacob, P., Liu, Y., ...

- 1436 Labrousse, C. A. (2015). Role of seasonal transitions and westerly jets in
 1437 East Asian paleoclimate. *Quaternary Science Reviews*, *108*, 111–129. doi:
 1438 10.1016/j.quascirev.2014.11.009
- 1439 Chiang, J. C. H., Swenson, L. M., & Kong, W. (2017). Role of seasonal transitions
 1440 and the westerlies in the interannual variability of the East Asian summer
 1441 monsoon precipitation. *Geophysical Research Letters*, *44*(8), 3788–3795. doi:
 1442 10.1002/2017GL072739
- 1443 Chiang, J. C. H., Zebiak, S. E., & Cane, M. A. (2001). Relative roles of elevated
 1444 heating and surface temperature gradients in driving anomalous surface winds
 1445 over tropical oceans. *Journal of the Atmospheric Sciences*, *58*(11), 1371–1394.
- 1446 Chou, C., & Neelin, J. D. (2001). Mechanisms limiting the southward extend of
 1447 the South American summer monsoon. *Geophysical Research Letters*, *28*(12),
 1448 2433–2436. doi: 10.1029/2000GL012138
- 1449 Chou, C., & Neelin, J. D. (2003). Mechanisms limiting the northward extent of the
 1450 northern summer monsoons over North America, Asia, and Africa. *Journal of*
 1451 *Climate*, *16*(3), 406–425. doi: 10.1175/1520-0442(2003)016<0406:MLTNEO>2.0
 1452 .CO;2
- 1453 Christensen, J., Krishna Kumar, K., Aldrian, E., An, S.-I., Cavalcanti, I., de Castro,
 1454 M., ... Zhou, T. (2013). Climate phenomena and their relevance for future
 1455 regional climate change [Book Section]. In T. Stocker et al. (Eds.), *Climate*
 1456 *Change 2013: The Physical Science Basis. Contribution of Working Group*
 1457 *I to the Fifth Assessment Report of the Intergovernmental Panel on Climate*
 1458 *Change* (p. 12171308). Cambridge, United Kingdom and New York, NY, USA:
 1459 Cambridge University Press. Retrieved from www.climatechange2013.org
 1460 doi: 10.1017/CBO9781107415324.028
- 1461 Cook, K. H. (2000). The South Indian Convergence Zone and Interannual Rainfall
 1462 Variability over Southern Africa. *Journal of Climate*, *13*(21), 3789–3804. doi:
 1463 10.1175/1520-0442(2000)013<3789:TSICZA>2.0.CO;2
- 1464 D’Agostino, R., Bader, J., Bordoni, S., Ferreira, D., & Jungclaus, J. (2019). North-
 1465 ern Hemisphere Monsoon Response to Mid-Holocene Orbital Forcing and
 1466 Greenhouse Gas-Induced Global Warming. *Geophysical Research Letters*, 1–11.
 1467 doi: 10.1029/2018GL081589
- 1468 Deplazes, G., Lckge, A., Peterson, L. C., Timmermann, A., Hamann, Y., Hughen,
 1469 K. A., ... Haug, G. H. (2013). Links between tropical rainfall and North
 1470 Atlantic climate during the last glacial period. *Nature Geosci*, *6*, 213–217. doi:
 1471 10.1038/ngeo1712
- 1472 Diaz, M., & Boos, W. R. (2019a). Barotropic growth of monsoon depressions. *Quar-*
 1473 *terly Journal of the Royal Meteorological Society*, *145*(719), 824–844.
- 1474 Diaz, M., & Boos, W. R. (2019b). Monsoon depression amplification by moist
 1475 barotropic instability in a vertically sheared environment. *Quarterly Journal of*
 1476 *the Royal Meteorological Society*, *145*(723), 2666–2684. doi: 10.1002/qj.3585
- 1477 Ding, Y., & Chan, J. C. L. (2005). The East Asian summer monsoon: An
 1478 overview. *Meteorology and Atmospheric Physics*, *89*(1), 117–142. doi:
 1479 10.1007/s00703-005-0125-z
- 1480 Donohoe, A., Marshall, J., Ferreira, D., & Mcgee, D. (2013). The relationship be-
 1481 tween ITCZ location and cross-equatorial atmospheric heat transport: From
 1482 the seasonal cycle to the last glacial maximum. *Journal of Climate*, *26*(11),
 1483 3597–3618. doi: 10.1175/JCLI-D-12-00467.1
- 1484 Egger, J., Weickmann, K., & Hoinka, K.-P. (2007). Angular momentum in the
 1485 global atmospheric circulation. *Reviews of Geophysics*, *45*(4). Retrieved
 1486 from [https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/](https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2006RG000213)
 1487 [2006RG000213](https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2006RG000213) doi: 10.1029/2006RG000213
- 1488 Ellis, A. W., Saffell, E. M., & Hawkins, T. W. (2004). A method for defining mon-
 1489 soon onset and demise in the Southwestern USA. *International Journal of Cli-*
 1490 *matology*, *24*(2), 247–265. doi: 10.1002/joc.996

- 1491 Eltahir, E. A., & Gong, C. (1996). Dynamics of Wet and Dry Years in West Africa.
1492 *Journal of Climate*, *9*, 1030–1042. doi: 10.1175/1520-0442(1996)009%3C1030:
1493 DOWADY%3E2.0.CO;2
- 1494 Emanuel, K. A. (1983a). The Lagrangian parcel dynamics of moist symmetric insta-
1495 bility. *Journal of the Atmospheric Sciences*, *40*, 2368–2376.
- 1496 Emanuel, K. A. (1983b). On assessing local conditional symmetric instability from
1497 atmospheric soundings. *Monthly weather review*, *111*, 2016–2033.
- 1498 Emanuel, K. A. (1988). Observational evidence of slantwise convective adjustment.
1499 *Monthly weather review*, *116*(9), 1805–1816.
- 1500 Emanuel, K. A. (1995). On Thermally Direct Circulations in Moist Atmo-
1501 spheres. *Journal of the Atmospheric Sciences*, *52*(9), 1529–1534. doi:
1502 10.1175/1520-0469(1995)052<1529:OTDCIM>2.0.CO;2
- 1503 Emanuel, K. A., Neelin, J. D., & Bretherton, C. S. (1994). On large-scale circula-
1504 tions in convecting atmospheres. *Quarterly Journal of the Royal Meteorological*
1505 *Society*, *120*(519), 1111–1143. doi: 10.1002/qj.49712051902
- 1506 Eroglu, D., McRobie, F. H., Ozken, I., Stemler, T., Wyrwoll, K. H., Breitenbach,
1507 S. F., ... Kurths, J. (2016). See-saw relationship of the Holocene East
1508 Asian-Australian summer monsoon. *Nature Communications*, *7*, 1–7. doi:
1509 10.1038/ncomms12929
- 1510 Faulk, S., Mitchell, J., & Bordoni, S. (2017). Effects of Rotation Rate and Seasonal
1511 Forcing on the ITCZ Extent in Planetary Atmospheres. *Journal of the Atmo-*
1512 *spheric Sciences*, *74*(3), 665–678. doi: 10.1175/JAS-D-16-0014.1
- 1513 Frierson, D. M., Hwang, Y. T., Fućkar, N. S., Seager, R., Kang, S. M., Donohoe, A.,
1514 ... Battisti, D. S. (2013). Contribution of ocean overturning circulation to
1515 tropical rainfall peak in the Northern Hemisphere. *Nature Geoscience*, *6*(11),
1516 940–944. doi: 10.1038/ngeo1987
- 1517 Frierson, D. M. W. (2007). The Dynamics of Idealized Convection Schemes and
1518 Their Effect on the Zonally Averaged Tropical Circulation. *Journal of the At-*
1519 *mospheric Sciences*, *64*(6), 1959–1976. doi: 10.1175/JAS3935.1
- 1520 Frierson, D. M. W., & Hwang, Y. T. (2012). Extratropical influence on ITCZ shifts
1521 in slab ocean simulations of global warming. *Journal of Climate*, *25*(2), 720–
1522 733. doi: 10.1175/JCLI-D-11-00116.1
- 1523 Fućkar, N. S., Xie, S.-P., Farneti, R., Maroon, E. A., & Frierson, D. M. (2013). In-
1524 fluence of the extratropical ocean circulation on the intertropical convergence
1525 zone in an idealized coupled general circulation model. *Journal of Climate*,
1526 *26*(13), 4612–4629. doi: 10.1175/JCLI-D-12-00294.1
- 1527 Gadgil, S. (2018). The monsoon system: Land-sea breeze or the ITCZ? *Journal of*
1528 *Earth System Science*, *127*(1), 1–29. doi: 10.1007/s12040-017-0916-x
- 1529 Geen, R., Lambert, F. H., & Vallis, G. K. (2018). Regime Change Behavior Dur-
1530 ing Asian Monsoon Onset. *Journal of Climate*, *31*, 3327–3348. Retrieved
1531 from <http://journals.ametsoc.org/doi/10.1175/JCLI-D-17-0118.1> doi:
1532 10.1175/JCLI-D-17-0118.1
- 1533 Geen, R., Lambert, F. H., & Vallis, G. K. (2019). Processes and Timescales in On-
1534 set and Withdrawal of Aquaplanet Monsoons?. *Journal of the Atmospheric Sci-*
1535 *ences*.
- 1536 Godbole, R. V. (1977). The composite structure of the monsoon depression. *Tellus*,
1537 *29*(1), 25–40.
- 1538 Goswami, B. N., & Ajaya Mohan, R. S. (2001). Intraseasonal Oscillations and In-
1539 terannual Variability of the Indian Summer Monsoon. *Journal of Climate*, *14*,
1540 1180–1198. doi: 10.1175/1520-0442(2001)014<1180:IOAIVO>2.0.CO;2
- 1541 Gottschalck, J. (2014). *What is the MJO, and why do we care?* [https://www](https://www.climate.gov/news-features/blogs/enso/what-mjo-and-why-do-we-care)
1542 [.climate.gov/news-features/blogs/enso/what-mjo-and-why-do-we-care](https://www.climate.gov/news-features/blogs/enso/what-mjo-and-why-do-we-care).
1543 NOAA climate.gov. (Accessed: 2019-12-24)
- 1544 Green, B., & Marshall, J. (2017). Coupling of Trade Winds with Ocean Circulation
1545 Damps ITCZ Shifts. *Journal of Climate*, *30*(12), 4395–4411. Retrieved from

- 1546 <https://doi.org/10.1175/JCLI-D-16-0818.1> doi: 10.1175/JCLI-D-16-0818
1547 .1
- 1548 Gruber, A., Su, X., Kanamitsu, M., & Schemm, J. (2000). The Compari-
1549 son of Two Merged Rain Gauge-Satellite Precipitation Datasets. *Bul-*
1550 *letin of the American Meteorological Society*, 81(11), 2631-2644. doi:
1551 10.1175/1520-0477(2000)081<2631:TCOTMR>2.3.CO;2
- 1552 Hagos, S. M., & Cook, K. H. (2007). Dynamics of the West African monsoon jump.
1553 *Journal of Climate*, 20(21), 5264–5284. doi: 10.1175/2007JCLI1533.1
- 1554 Halley, E. (1686). An Historical Account of the Trade Winds, and Monsoons,
1555 Observable in the Seas between and Near the Tropicks, with an Attempt to
1556 Assign the Phisical Cause of the Said Winds, By E. Halley. *Philosophical*
1557 *Transactions of the Royal Society of London*, 16(179-191), 153–168. doi:
1558 10.1098/rstl.1686.0026
- 1559 Hartmann, D. L., & Michelsen, M. L. (1989). Intraseasonal periodicities in Indian
1560 rainfall. *Journal of the Atmospheric Sciences*, 46(18), 2838–2862.
- 1561 Held, I. M. (2005). The gap between simulation and understanding in climate
1562 modeling. *Bulletin of the American Meteorological Society*, 86(11), 1609-
1563 1614. Retrieved from <https://doi.org/10.1175/BAMS-86-11-1609> doi:
1564 10.1175/BAMS-86-11-1609
- 1565 Held, I. M., & Coauthors. (2000). *The General Circulation of the Atmosphere*.
1566 Woods Hole Oceanographic Institution.
- 1567 Hendon, H. H., & Liebmann, B. (1990). A Composite Study of Onset of the Aus-
1568 tralian Summer Monsoon. *Journal of Atmospheric Sciences*, 47(18), 2227–
1569 2240. doi: 10.1175/1520-0469(1990)047<2227:ACS000>2.0.CO;2
- 1570 Hide, R. (1969). Dynamics of the Atmospheres of the Major Planets with an Ap-
1571 pendix on the Viscous Boundary Layer at the Rigid Bounding Surface of an
1572 Electrically-Conducting Rotation Fluid in the Presence of a Magnetic Field.
1573 *J. Atmos. Sci.*, 26, 841–853. doi: 10.1175/1520-0469(1969)026%3C0841:
1574 DOTAOT%3E2.0.CO;2
- 1575 Hilgenbrink, C. C., & Hartmann, D. L. (2018). The response of Hadley circulation
1576 extent to an idealized representation of poleward ocean heat transport in an
1577 aquaplanet GCM. *Journal of Climate*, 31, 9753–9770.
- 1578 Hill, S. A. (2019). Theories for Past and Future Monsoon Rainfall Changes. *Current*
1579 *Climate Change Reports*, 5, 160–171.
- 1580 Hill, S. A., Bordoni, S., & Mitchell, J. L. (2019). Hadley cell emergence and extent
1581 in axisymmetric, nearly inviscid, planetary atmospheres. *J. Atmos. Sci.*
- 1582 Hill, S. A., Ming, Y., & Held, I. M. (2015). Mechanisms of forced tropical meridional
1583 energy flux change. *Journal of Climate*, 28, 1725–1742. doi: 10.1175/JCLI-D
1584 -14-00165.1
- 1585 Hill, S. A., Ming, Y., Held, I. M., & Zhao, M. (2017). A moist static energy budget-
1586 based analysis of the Sahel rainfall response to uniform oceanic warming. *Jour-*
1587 *nal of Climate*, 30(15), 5637–5660. doi: 10.1175/JCLI-D-16-0785.1
- 1588 Hill, S. A., Ming, Y., & Zhao, M. (2018). Robust Responses of the Hydrological Cy-
1589 cle to Global Warming. *Journal of Climate*, 1931, 9793–9814. doi: 10.1175/
1590 JCLI-D-18-0238.1
- 1591 Holton, J. R., Wallace, J. M., & Young, J. (1971). On boundary layer dynamics and
1592 the ITCZ. *Journal of the Atmospheric Sciences*, 28(2), 275–280.
- 1593 Houze Jr, R. A. (2018). 100 years of research on mesoscale convective systems. *Me-*
1594 *teorological Monographs*, 59, 17–1.
- 1595 Huffman, G. J., Adler, R. F., Morrissey, M. M., Bolvin, D. T., Curtis, S., Joyce, R.,
1596 ... Susskind, J. (2001). Global Precipitation at One-Degree Daily Resolution
1597 from Multisatellite Observations. *Journal of Hydrometeorology*, 2, 36–50.
- 1598 Hui, K., & Bordoni, S. (in prep.). Influence of Continental Geometry on the Onset
1599 and Spatial Distribution of Monsoonal Precipitation.
- 1600 Hunt, K. M., Turner, A. G., Inness, P. M., Parker, D. E., & Levine, R. C. (2016).

- 1601 On the structure and dynamics of Indian monsoon depressions. *Monthly*
 1602 *Weather Review*, 144(9), 3391–3416.
- 1603 Hurley, J. V., & Boos, W. R. (2013). Interannual variability of monsoon precipita-
 1604 tion and local subcloud equivalent potential temperature. *Journal of Climate*,
 1605 26(23), 9507–9527. doi: 10.1175/JCLI-D-12-00229.1
- 1606 Hurley, J. V., & Boos, W. R. (2015). A global climatology of monsoon low-pressure
 1607 systems. *Quarterly Journal of the Royal Meteorological Society*, 141(689),
 1608 1049–1064. doi: 10.1002/qj.2447
- 1609 Jeevanjee, N., Hassanzadeh, P., Hill, S., & Sheshadri, A. (2017). A perspective
 1610 on climate model hierarchies. *Journal of Advances in Modeling Earth Systems*,
 1611 9(4), 1760–1771. doi: 10.1002/2017MS001038
- 1612 Jiang, X., Adames, . F., Zhao, M., Waliser, D., & Maloney, E. (2018). A unified
 1613 moisture mode framework for seasonality of the Madden-Julian oscillation.
 1614 *Journal of Climate*, 31(11), 4215–4224. Retrieved from [https://doi.org/](https://doi.org/10.1175/JCLI-D-17-0671.1)
 1615 10.1175/JCLI-D-17-0671.1 doi: 10.1175/JCLI-D-17-0671.1
- 1616 Jourdain, N. C., Gupta, A. S., Taschetto, A. S., Ummenhofer, C. C., Moise, A. F., &
 1617 Ashok, K. (2013). The Indo-Australian monsoon and its relationship to ENSO
 1618 and IOD in reanalysis data and the CMIP3/CMIP5 simulations. *Climate*
 1619 *Dynamics*, 41(11), 3073–3102. Retrieved from [https://doi.org/10.1007/](https://doi.org/10.1007/s00382-013-1676-1)
 1620 s00382-013-1676-1 doi: 10.1007/s00382-013-1676-1
- 1621 Kang, S. M., Frierson, D. M. W., & Held, I. M. (2009). The Tropical Response
 1622 to Extratropical Thermal Forcing in an Idealized GCM: The Importance of
 1623 Radiative Feedbacks and Convective Parameterization. *Journal of the Atmo-*
 1624 *spheric Sciences*, 66(9), 2812–2827. doi: 10.1175/2009jas2924.1
- 1625 Kang, S. M., & Held, I. M. (2012). Tropical precipitation, SSTs and the surface
 1626 energy budget: A zonally symmetric perspective. *Climate Dynamics*, 38(9-10),
 1627 1917–1924. doi: 10.1007/s00382-011-1048-7
- 1628 Kang, S. M., Held, I. M., Frierson, D. M., & Zhao, M. (2008). The response of the
 1629 ITCZ to extratropical thermal forcing: Idealized slab-ocean experiments with a
 1630 GCM. *Journal of Climate*, 21(14), 3521–3532. doi: 10.1175/2007JCLI2146.1
- 1631 Kerns, B., & Chen, S. S. (2020). A 20-year climatology of Madden-Julian Oscillation
 1632 convection: Large-scale precipitation tracking from TRMM-GPM rainfall. *In*
 1633 *press, J. Geophys. Res. Atmos.*
- 1634 Kiladis, G. N., Thorncroft, C. D., & Hall, N. M. J. (2006). Three-Dimensional Struc-
 1635 ture and Dynamics of African Easterly Waves. Part I: Observations. *Journal of*
 1636 *the Atmospheric Sciences*, 63(9), 2212–2230. doi: 10.1175/JAS3741.1
- 1637 Kiladis, G. N., Wheeler, M. C., Haertel, P. T., Straub, K. H., & Roundy, P. E.
 1638 (2009). Convectively coupled equatorial waves. *Reviews of Geophysics*,
 1639 47(RG2003).
- 1640 Kobayashi, S., Ota, Y., Harada, Y., Ebata, A., Moriya, M., Onoda, H., . . . Taka-
 1641 hashi, K. (2015). The JRA-55 Reanalysis: General Specifications and
 1642 Basic Characteristics. *J. Meteorol. Soc. Jpn.*, 93(1), 5–48. doi: 10.2151/
 1643 jmsj.2015-001
- 1644 Kodama, Y. (1992). Large-Scale Common Features of Subtropical Precipitation
 1645 Zones (the Baiu Frontal Zone, the SPCZ, and the SACZ). Part I: Character-
 1646 istics of Subtropical Frontal Zones. *Journal of the Meteorological Society of*
 1647 *Japan*, 70(4), 813–836. doi: 10.2151/jmsj1965.70.4.813
- 1648 Kong, W., & Chiang, J. C. H. (2020). Interaction of the Westerlies with the Tibetan
 1649 Plateau in Determining the Mei-Yu Termination. *Journal of Climate*, 33(1),
 1650 339–363. doi: 10.1175/JCLI-D-19-0319.1
- 1651 Kong, W., Swenson, L. M., & Chiang, J. C. H. (2017). Seasonal transitions and the
 1652 westerly jet in the Holocene East Asian summer monsoon. *Journal of Climate*,
 1653 30(9), 3343–3365. doi: 10.1175/JCLI-D-16-0087.1
- 1654 Kothawale, D., & Kumar, K. R. (2002). Tropospheric temperature variation over In-
 1655 dia and links with the Indian summer monsoon: 1971–2000. *Mausam*, 53, 289–

- 1656 308.
- 1657 Lea, D. W., Pak, D. K., Peterson, L. C., & Hughen, K. A. (2003). Temperatures
1658 over the Last Glacial Termination Synchronicity of Tropical and High-Latitude
1659 Atlantic Temperatures over the Last Glacial Termination. *Science*, *301*, 1361–
1660 1364. doi: 10.1126/science.1088470
- 1661 Lebel, T. (2003). Seasonal cycle and interannual variability of the Sahelian rainfall
1662 at hydrological scales. *Journal of Geophysical Research*, *108*(D8), 8389. doi: 10
1663 .1029/2001JD001580,
- 1664 Lee, J.-Y., Wang, B., Wheeler, M. C., Fu, X., Waliser, D. E., & Kang, I.-S. (2013).
1665 Real-time multivariate indices for the boreal summer intraseasonal oscillation
1666 over the Asian summer monsoon region. *Climate Dynamics*, *40*(1), 493–
1667 509. Retrieved from <https://doi.org/10.1007/s00382-012-1544-4> doi:
1668 10.1007/s00382-012-1544-4
- 1669 Lenters, J. D., & Cook, K. H. (1997). On the Origin of the Bolivian High and Re-
1670 lated Circulation Features of the South American Climate. *Journal of the At-
1671 mospheric Sciences*, *54*, 656–678. doi: 10.1175/1520-0469(1997)054%3C0656:
1672 OTOOTB%3E2.0.CO;2
- 1673 Levine, X. J., & Schneider, T. (2011). Response of the Hadley Circulation to
1674 Climate Change in an Aquaplanet GCM Coupled to a Simple Representa-
1675 tion of Ocean Heat Transport. *Journal of the Atmospheric Sciences*, *68*(4),
1676 769–783. Retrieved from <https://doi.org/10.1175/2010JAS3553.1> doi:
1677 10.1175/2010JAS3553.1
- 1678 Levy, G., & Battisti, D. S. (1995). The symmetric stability and the low level equator-
1679 ial flow. *Global Atmosphere-Ocean System*, *3*(4), 341–354.
- 1680 Liebmann, B., & Hendon, H. H. (1990). Synoptic-scale disturbances near the equator.
1681 *Journal of the Atmospheric Sciences*, *47*(12), 1463–1479.
- 1682 Lindzen, R. S., & Hou, A. Y. (1988). Hadley Circulations for Zonally Averaged
1683 Heating Centred off the Equator. *J. Atmos. Sci.*, *45*(17), 2416–2427.
- 1684 Lindzen, R. S., & Nigam, S. (1987). On the role of sea surface temperature gradients
1685 in forcing low-level winds and convergence in the tropics. *Journal of the Atmo-
1686 spheric Sciences*, *44*(17), 2418–2436.
- 1687 Linho, L. H., Huang, X., & Lau, N. C. (2008). Winter-to-spring transition in east
1688 Asia: A planetary-scale perspective of the south China spring rain onset. *Jour-
1689 nal of Climate*, *21*(13), 3081–3096. doi: 10.1175/2007JCLI1611.1
- 1690 Liu, X., & Battisti, D. S. (2015). The influence of orbital forcing of tropical in-
1691 solation on the climate and isotopic composition of precipitation in South
1692 America. *Journal of Climate*, *28*(12), 4841–4862.
- 1693 Liu, X., Battisti, D. S., & Donohoe, A. (2017). Tropical precipitation and cross-
1694 equatorial ocean heat transport during the mid-Holocene. *Journal of Climate*,
1695 *30*(10), 3529–3547. doi: 10.1175/JCLI-D-16-0502.1
- 1696 Madden, R. A., & Julian, P. R. (1971). Detection of a 40–50 day oscillation in the
1697 zonal wind in the tropical Pacific. *Journal of the Atmospheric Sciences*, *28*(5),
1698 702–708.
- 1699 Madden, R. A., & Julian, P. R. (1972). Description of global-scale circulation cells
1700 in the tropics with a 40–50 day period. *Journal of the Atmospheric Sciences*,
1701 *29*(6), 1109–1123.
- 1702 Maher, P., Gerber, E. P., Medeiros, B., Merlis, T. M., Sherwood, S., Sheshadri, A.,
1703 ... ZuritaGotor, P. (2019). Model Hierarchies for Understanding Atmo-
1704 spheric Circulation. *Reviews of Geophysics*, 2018RG000607. Retrieved from
1705 <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018RG000607> doi:
1706 10.1029/2018RG000607
- 1707 Mao, J., & Wu, G. (2007). Interannual variability in the onset of the summer mon-
1708 soon over the Eastern Bay of Bengal. *Theoretical and Applied Climatology*,
1709 *89*(3-4), 155–170. doi: 10.1007/s00704-006-0265-1
- 1710 Marengo, J. A., Liebmann, B., Grimm, A. M., Misra, V., Silva Dias, P. L., Cav-

- 1711 alcanti, I. F., ... Alves, L. M. (2012). Recent developments on the South
 1712 American monsoon system. *International Journal of Climatology*, *32*(1), 1–21.
 1713 doi: 10.1002/joc.2254
- 1714 Maroon, E. A., Frierson, D. M., & Battisti, D. S. (2015). The tropical precipitation
 1715 response to Andes topography and ocean heat fluxes in an aquaplanet model.
 1716 *Journal of Climate*, *28*(1), 381–398. doi: 10.1175/JCLI-D-14-00188.1
- 1717 Marshall, J., Donohoe, A., Ferreira, D., & McGee, D. (2014). The ocean’s role in
 1718 setting the mean position of the Inter-Tropical Convergence Zone. *Climate Dy-*
 1719 *namics*, *42*(7-8), 1967–1979. doi: 10.1007/s00382-013-1767-z
- 1720 Martin, G. M., Chevuturi, A., Comer, R. E., Dunstone, N. J., Scaife, A. A., &
 1721 Zhang, D. (2019). Predictability of South China Sea Summer Mon-
- 1722 soon Onset. *Advances in Atmospheric Sciences*, *36*(3), 253–260. doi:
 1723 10.1007/s00376-018-8100-z
- 1724 Matsuno, T. (1966). Quasi-Geostrophic Motions in the Equatorial Area. *J. Meteor.*
 1725 *Soc. Japan*, *44*(1), 25–43. doi: 10.2151/jmsj1965.44.1.25
- 1726 McGee, D., Donohoe, A., Marshall, J., & Ferreira, D. (2014). Changes in ITCZ
 1727 location and cross-equatorial heat transport at the Last Glacial Maximum,
 1728 Heinrich Stadial 1, and the mid-Holocene. *Earth and Planetary Science Let-*
 1729 *ters*, *390*, 69–79. doi: 10.1016/j.epsl.2013.12.043
- 1730 Merlis, T. M., Schneider, T., Bordoni, S., & Eisenman, I. (2013). Hadley circu-
- 1731 lation response to orbital precession. Part I: Aquaplanets. *Journal of Climate*,
 1732 *26*, 740–753. doi: 10.1175/JCLI-D-11-00716.1
- 1733 Molnar, P., Boos, W. R., & Battisti, D. S. (2010). Orographic controls on cli-
- 1734 mate and paleoclimate of Asia: thermal and mechanical roles for the Tibetan
 1735 Plateau. *Annual Review of Earth and Planetary Sciences*, *38*(1), 77–102. doi:
 1736 10.1146/annurev-earth-040809-152456
- 1737 Mooley, D. A. (1973). Some aspects of Indian monsoon depression and associated
 1738 rainfall. *Monthly Weather Review*, *101*, 271–280.
- 1739 Nicholson, S. E. (2009a). On the factors modulating the intensity of the tropical
 1740 rainbelt over West Africa. *International Journal of Climatology*, *29*(5), 673–
 1741 689.
- 1742 Nicholson, S. E. (2009b). A revised picture of the structure of the "monsoon" and
 1743 land ITCZ over West Africa. *Climate Dynamics*, *32*(7-8), 1155–1171. doi: 10
 1744 .1007/s00382-008-0514-3
- 1745 Nicholson, S. E. (2017). Climate and climatic variability of rainfall over eastern
 1746 Africa. *Reviews of Geophysics*, 590–635. doi: 10.1002/2016RG000544
- 1747 Nie, J., Boos, W. R., & Kuang, Z. (2010). Observational evaluation of a convective
 1748 quasi-equilibrium view of monsoons. *Journal of Climate*, *23*(16), 4416–4428.
 1749 doi: 10.1175/2010JCLI3505.1
- 1750 Parthasarathy, B., Munot, A. A., & Kothawale, D. R. (1994). All-India monthly and
 1751 seasonal rainfall series: 1871–1993. *Theoretical and Applied Climatology*, *49*(4),
 1752 217–224. doi: 10.1007/BF00867461
- 1753 Pausata, F. S., Battisti, D. S., Nisancioglu, K. H., & Bitz, C. M. (2011). Chinese
 1754 stalagmite $\delta^{18}\text{O}$ controlled by changes in the Indian monsoon during a simu-
- 1755 lated Heinrich event. *Nature Geoscience*, *4*(7), 474.
- 1756 Plumb, R. A., & Hou, A. Y. (1992). The response of a zonally symmetric atmo-
- 1757 sphere to subtropical thermal forcing - Threshold behavior. *Journal of the*
 1758 *Atmospheric Sciences*, *49*(19), 1790–1799. doi: 10.1175/1520-0469(1992)
 1759 049<1790:TROAZS>2.0.CO;2
- 1760 Privé, N. C., & Plumb, R. A. (2007a). Monsoon Dynamics with Interactive For-
- 1761 cing. Part I: Axisymmetric Studies. *Journal of the Atmospheric Sciences*, *64*(5),
 1762 1417–1430. doi: 10.1175/JAS3916.1
- 1763 Privé, N. C., & Plumb, R. A. (2007b). Monsoon Dynamics with Interactive For-
- 1764 cing. Part II: Impact of Eddies and Asymmetric Geometries. *Journal of the At-*
 1765 *mospheric Sciences*, *64*(5), 1431–1442. doi: 10.1175/JAS3917.1

- 1766 Rao, V. B., Cavalcanti, I. F. A., & Hada, K. (1996). Annual variation of rain-
1767 fall over Brazil and water vapor characteristics over South America. *Jour-*
1768 *nal of Geophysical Research: Atmospheres*, *101*(D21), 26539–26551. doi:
1769 10.1029/96JD01936
- 1770 Reed, R. J., Norquist, D. C., & Recker, E. E. (1977). The structure and properties
1771 of African wave disturbances as observed during Phase III of GATE. *Monthly*
1772 *Weather Review*, *105*(3), 317–333.
- 1773 Rodwell, M. J., & Hoskins, B. J. (2001). Subtropical anticyclones and summer mon-
1774 soons. *Journal of Climate*, *14*(15), 3192–3211. doi: 10.1175/1520-0442(2001)
1775 014{3192:SAASM}2.0.CO;2
- 1776 Roehrig, R., Bouniol, D., Guichard, F., Hourdin, F., & Redelsperger, J.-L. (2013).
1777 The Present and Future of the West African Monsoon: A Process-Oriented
1778 Assessment of CMIP5 Simulations along the AMMA Transect. *Journal of*
1779 *Climate*, *26*(17), 6471–6505. Retrieved from [https://doi.org/10.1175/](https://doi.org/10.1175/JCLI-D-12-00505.1)
1780 [JCLI-D-12-00505.1](https://doi.org/10.1175/JCLI-D-12-00505.1) doi: 10.1175/JCLI-D-12-00505.1
- 1781 Roundy, P. E., & Frank, W. M. (2004). A climatology of waves in the equatorial re-
1782 gion. *Journal of the Atmospheric Sciences*, *61*(17), 2105–2132.
- 1783 Sampe, T., & Xie, S.-P. (2010). Large-scale dynamics of the Meiyu-Baiu rainband:
1784 Environmental forcing by the westerly jet. *Journal of Climate*, *23*(1), 113–134.
1785 doi: 10.1175/2009JCLI3128.1
- 1786 Schneider, T. (2017). Feedback of Atmosphere-Ocean Coupling on Shifts of the
1787 Intertropical Convergence Zone. *Geophysical Research Letters*, *44*(22), 11,644-
1788 11,653. Retrieved from [https://agupubs.onlinelibrary.wiley.com/doi/](https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2017GL075817)
1789 [abs/10.1002/2017GL075817](https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2017GL075817) doi: 10.1002/2017GL075817
- 1790 Schneider, T., Bischoff, T., & Haug, G. H. (2014). Migrations and dynamics of
1791 the Intertropical Convergence Zone. *Nature*, *513*(7516), 45–53. doi: 10.1038/
1792 nature13636
- 1793 Schneider, T., & Bordoni, S. (2008). Eddy-Mediated Regime Transitions in
1794 the Seasonal Cycle of a Hadley Circulation and Implications for Monsoon
1795 Dynamics. *Journal of the Atmospheric Sciences*, *65*(1), 915–934. doi:
1796 10.1175/2007JAS2415.1
- 1797 Schumacher, C., & Houze Jr, R. A. (2003). Stratiform rain in the tropics as seen by
1798 the TRMM precipitation radar. *Journal of Climate*, *16*(11), 1739–1756.
- 1799 Schwendike, J., Govekar, P., Reeder, M. J., Wardle, R., Berry, G. J., & Jakob, C.
1800 (2014). Local partitioning of the overturning circulation in the tropics and
1801 the connection to the Hadley and Walker circulations. *Journal of Geophysical*
1802 *Research*, *119*(3), 1322–1339. doi: 10.1002/2013JD020742
- 1803 Segele, Z. T., & Lamb, P. J. (2005). Characterization and variability of Kiremt rainy
1804 season over Ethiopia. *Meteorology and Atmospheric Physics*, *89*(1-4), 153–180.
1805 doi: 10.1007/s00703-005-0127-x
- 1806 Seo, J., Kang, S. M., & Merlis, T. M. (2017). A model intercomparison of the trop-
1807 ical precipitation response to a CO₂ doubling in aquaplanet simulations. *Geo-*
1808 *physical Research Letters*, *44*(2), 993–1000. Retrieved from [https://agupubs](https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2016GL072347)
1809 [.onlinelibrary.wiley.com/doi/abs/10.1002/2016GL072347](https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2016GL072347) doi: 10.1002/
1810 2016GL072347
- 1811 Seth, A., Giannini, A., Rojas, M., Rauscher, S. A., Bordoni, S., Singh, D., & Ca-
1812 margo, S. J. (2019, Jun 01). Monsoon responses to climate changes—
1813 connecting past, present and future. *Current Climate Change Reports*, *5*(2),
1814 63–79. Retrieved from <https://doi.org/10.1007/s40641-019-00125-y> doi:
1815 10.1007/s40641-019-00125-y
- 1816 Shaw, T. A. (2014). On the Role of Planetary-Scale Waves in the Abrupt Seasonal
1817 Transition of the Northern Hemisphere General Circulation. *Journal of the At-*
1818 *mospheric Sciences*, *71*(5), 1724–1746. doi: 10.1175/JAS-D-13-0137.1
- 1819 Shaw, T. A., & Voigt, A. (2015). Tug of war on summertime circulation between
1820 radiative forcing and sea surface warming. *Nature Geoscience*, *8*(7), 560–566.

- 1821 doi: 10.1038/ngeo2449
- 1822 Shekhar, R., & Boos, W. R. (2016). Improving energy-based estimates of monsoon
1823 location in the presence of proximal deserts. *Journal of Climate*, 29(13), 4741–
1824 4761. doi: 10.1175/JCLI-D-15-0747.1
- 1825 Sikka, D. (1978). Some aspects of the life history, structure and movement of mon-
1826 soon depressions. In *Monsoon dynamics* (pp. 1501–1529). Springer.
- 1827 Sikka, D. R., & Gadgil, S. (1980). On the Maximum Cloud Zone and the
1828 ITCZ over Indian, Longitudes during the Southwest Monsoon. *Monthly*
1829 *Weather Review*, 108(11), 1840-1853. Retrieved from [https://doi.org/
1830 10.1175/1520-0493\(1980\)108<1840:OTMCZA>2.0.CO;2](https://doi.org/10.1175/1520-0493(1980)108<1840:OTMCZA>2.0.CO;2) doi: 10.1175/
1831 1520-0493(1980)108(1840:OTMCZA)2.0.CO;2
- 1832 Simpson, G. (1921). The South-West monsoon. *Quarterly Journal of the Royal Me-*
1833 *teorological Society*, 47(199), 151–171.
- 1834 Singh, M. S. (2019). Limits on the extent of the solstitial Hadley Cell: The role of
1835 planetary rotation. *Journal of Atmospheric Sciences*.
- 1836 Smyth, J. E., Hill, S. A., & Ming, Y. (2018). Simulated Responses of the West
1837 African Monsoon and Zonal-Mean Tropical Precipitation to Early Holocene
1838 Orbital Forcing. *Geophysical Research Letters*(Figure 1), 49–57. doi:
1839 10.1029/2018GL080494
- 1840 Sobel, A., & Maloney, E. (2013). Moisture Modes and the Eastward Propagation
1841 of the MJO. *Journal of the Atmospheric Sciences*, 70(1), 187-192. Retrieved
1842 from <https://doi.org/10.1175/JAS-D-12-0189.1> doi: 10.1175/JAS-D-12
1843 -0189.1
- 1844 Sperber, K. R., Annamalai, H., Kang, I.-S., Kitoh, A., Moise, A., Turner, A., ...
1845 Zhou, T. (2013). The Asian summer monsoon: An intercomparison of CMIP5
1846 vs. CMIP3 simulations of the late 20th century. *Climate Dynamics*, 41(9),
1847 2711–2744. Retrieved from <https://doi.org/10.1007/s00382-012-1607-6>
1848 doi: 10.1007/s00382-012-1607-6
- 1849 Stevens, D. E. (1983). On symmetric stability and instability of zonal mean flows
1850 near the equator. *Journal of the Atmospheric Sciences*, 40(4), 882–893.
- 1851 Sultan, B., & Janicot, S. (2003). The West African monsoon dynamics. Part II: The
1852 preonset and onset of the summer monsoon. *Journal of climate*, 16(21), 3407–
1853 3427. doi: 10.1175/1520-0442(2003)016<3407:TWAMDP>2.0.CO;2
- 1854 Takahashi, K. (2005). The annual cycle of heat content in the Peru current region.
1855 *Journal of Climate*, 18(23), 4937–4954. doi: 10.1175/JCLI3572.1
- 1856 Takahashi, K., & Battisti, D. S. (2007). Processes Controlling the Mean Tropical
1857 Pacific Precipitation Pattern. Part I: The Andes and the Eastern Pacific ITCZ.
1858 *Journal of Climate*, 20(14), 3434–3451. doi: 10.1175/jcli4198.1
- 1859 Tomas, R. A., & Webster, P. J. (1997). The role of inertial instability in determin-
1860 ing the location and strength of near-equatorial convection. *Quarterly Journal*
1861 *of the Royal Meteorological Society*, 123(542), 1445–1482.
- 1862 Trenberth, K. E., Stepaniak, D. P., & Caron, J. M. (2000). The global mon-
1863 soon as seen through the divergent atmospheric circulation. *Journal of*
1864 *Climate*, 13(22), 3969–3993. doi: 10.1175/1520-0442(2000)013<3969:
1865 TGMASST>2.0.CO;2
- 1866 Uppala, S. M., Kållberg, P., Simmons, A., Andrae, U., Bechtold, V. D. C., Fiorino,
1867 M., ... others (2005). The ERA-40 re-analysis. *Q. J. R. Meteorol. Soc.*, 131,
1868 2961–3012.
- 1869 Walker, C. C., & Schneider, T. (2006). Eddy Influences on Hadley Circulations:
1870 Simulations with an Idealized GCM. *Journal of the Atmospheric Sciences*,
1871 63(12), 3333–3350. Retrieved from [http://journals.ametsoc.org/doi/abs/
1872 10.1175/JAS3821.1](http://journals.ametsoc.org/doi/abs/10.1175/JAS3821.1) doi: 10.1175/JAS3821.1
- 1873 Walker, J. M., & Bordoni, S. (2016). Onset and withdrawal of the large-scale South
1874 Asian monsoon: A dynamical definition using change point detection. *Geo-*
1875 *physical Research Letters*, 43(22), 11,815–11,822. doi: 10.1002/2016GL071026

- 1876 Walker, J. M., Bordoni, S., & Schneider, T. (2015). Interannual variability in the
1877 large-scale dynamics of the South Asian summer monsoon. *Journal of Climate*,
1878 *28*(9), 3731–3750. doi: 10.1175/JCLI-D-14-00612.1
- 1879 Wallace, J., Battisti, D., Thompson, D., & Hartmann, D. (2020). The Atmospheric
1880 General Circulation. *Under review, Cambridge University Press*.
- 1881 Wang, B., & Ding, Q. (2006). Changes in global monsoon precipitation over the
1882 past 56 years. *Geophysical Research Letters*, *33*(6), L06711. doi: 10.1029/
1883 2005GL025347
- 1884 Wang, B., Ding, Q., & Joseph, P. V. (2009). Objective definition of the Indian sum-
1885 mer monsoon onset. *Journal of Climate*, *22*(12), 3303–3316. doi: 10.1175/
1886 2008JCLI2675.1
- 1887 Wang, B., & LinHo. (2002). Rainy Season of the Asian-Pacific Summer Monsoon.
1888 *Journal of Climate*, *15*, 386–398. doi: 10.1175/1520-0442(2002)015%3C0386:
1889 RSOTAP%3E2.0.CO;2
- 1890 Wang, B., LinHo, Zhang, Y., & Lu, M. M. (2004). Definition of South China Sea
1891 monsoon onset and commencement of the East Asian summer monsoon. *Jour-
1892 nal of Climate*, *17*(4), 699–710. doi: 10.1175/2932.1
- 1893 Wang, B., Liu, J., Kim, H. J., Webster, P. J., & Yim, S. Y. (2012). Recent change of
1894 the global monsoon precipitation (1979-2008). *Climate Dynamics*, *39*(5), 1123–
1895 1135. doi: 10.1007/s00382-011-1266-z
- 1896 Wang, B., Wu, R., & Lau, K.-M. (2001). Interannual Variability of the Asian
1897 Summer Monsoon: Contrasts between the Indian and the Western North
1898 Pacific-East Asian Monsoons. *Journal of Climate*, *14*, 4073–4090. doi:
1899 10.1175/1520-0442(2001)014<4073:IVOTAS>2.0.CO;2
- 1900 Wang, B., & Xie, X. (1997). A model for the boreal summer intraseasonal oscilla-
1901 tion. *Journal of the Atmospheric Sciences*, *54*(1), 72-86. doi: 10.1175/1520
1902 -0469(1997)054(0072:AMFTBS)2.0.CO;2
- 1903 Wang, P. X., Wang, B., Cheng, H., Fasullo, J., Guo, Z. T., Kiefer, T., & Liu,
1904 Z. Y. (2014). The global monsoon across timescales: Coherent variabil-
1905 ity of regional monsoons. *Climate of the Past*, *10*(6), 2007–2052. doi:
1906 10.5194/cp-10-2007-2014
- 1907 Wang, P. X., Wang, B., Cheng, H., Fasullo, J., Guo, Z. T., Kiefer, T., & Liu,
1908 Z. Y. (2017). The global monsoon across time scales: Mechanisms and
1909 outstanding issues. *Earth-Science Reviews*, *174*, 84–121. doi: 10.1016/
1910 j.earscirev.2017.07.006
- 1911 Wei, H.-H., & Bordoni, S. (2016). On the Role of the African Topography in
1912 the South Asian Monsoon. *Journal of the Atmospheric Sciences*, *73*(8),
1913 3197–3212. Retrieved from [http://journals.ametsoc.org/doi/10.1175/
1914 JAS-D-15-0182.1](http://journals.ametsoc.org/doi/10.1175/JAS-D-15-0182.1) doi: 10.1175/JAS-D-15-0182.1
- 1915 Wei, H.-H., & Bordoni, S. (2018). Energetic Constraints on the ITCZ Position in
1916 Idealized Simulations With a Seasonal Cycle. *Journal of Advances in Modeling
1917 Earth Systems*, *10*(7), 1708–1725. doi: 10.1029/2018MS001313
- 1918 Wheeler, M., & Kiladis, G. N. (1999). Convectively coupled equatorial waves:
1919 Analysis of clouds and temperature in the wavenumber-frequency do-
1920 main. *Journal of the Atmospheric Sciences*, *56*(3), 374–399. doi: 10.1175/
1921 1520-0469(1999)056(0374:CCEWAO)2.0.CO;2
- 1922 White, R., Battisti, D., & Skok, G. (2017). Tracking precipitation events in time
1923 and space in gridded observational data. *Geophysical Research Letters*, *44*(16),
1924 8637–8646.
- 1925 Wolff, E. W., Chappellaz, J., Blunier, T., Rasmussen, S. O., & Svensson, A. (2010).
1926 Millennial-scale variability during the last glacial: The ice core record. *Quater-
1927 nary Science Reviews*, *29*, 2828–2838.
- 1928 Wu, M.-L. C., Reale, O., Schubert, S. D., Suarez, M. J., & Thorncroft, C. D. (2012).
1929 African easterly jet: Barotropic instability, waves, and cyclogenesis. *Journal of
1930 Climate*, *25*(5), 1489–1510.

- 1931 Wu, R., & Wang, B. (2001). Multi-stage onset of the summer monsoon over the
1932 western North Pacific. *Climate Dynamics*, *17*(4), 277–289. doi: 10.1007/
1933 s003820000118
- 1934 Xiang, B., Zhao, M., Ming, Y., Yu, W., & Kang, S. M. (2018). Contrasting impacts
1935 of radiative forcing in the Southern Ocean versus southern tropics on ITCZ po-
1936 sition and energy transport in one GFDL climate model. *Journal of Climate*,
1937 *31*(14), 5609–5628. doi: 10.1175/JCLI-D-17-0566.1
- 1938 Yang, G.-Y., Methven, J., Woolnough, S., Hodges, K., & Hoskins, B. (2018). Link-
1939 ing African easterly wave activity with equatorial waves and the influence of
1940 Rossby waves from the Southern Hemisphere. *Journal of the Atmospheric*
1941 *Sciences*, *75*(6), 1783–1809.
- 1942 Zhai, J., & Boos, W. R. (2017). The drying tendency of shallow meridional cir-
1943 culations in monsoons. *Quarterly Journal of the Royal Meteorological Society*,
1944 *143*(708), 2655–2664. doi: 10.1002/qj.3091
- 1945 Zhang, C. (2005). Madden-Julian oscillation. *Reviews of Geophysics*, *43*(2).
- 1946 Zhang, C., Nolan, D. S., Thorncroft, C. D., & Nguyen, H. (2008). Shallow merid-
1947 ional circulations in the tropical atmosphere. *Journal of Climate*, *21*(14),
1948 3453–3470. doi: 10.1175/2007JCLI1870.1
- 1949 Zhang, G., & Wang, Z. (2013). Interannual variability of the Atlantic Hadley circu-
1950 lation in boreal summer and its impacts on tropical cyclone activity. *Journal of*
1951 *Climate*, *26*(21), 8529–8544. doi: 10.1175/JCLI-D-12-00802.1
- 1952 Zhang, R., & Delworth, T. L. (2005). Simulated tropical response to a substan-
1953 tial weakening of the Atlantic thermohaline circulation. *Journal of Climate*,
1954 *18*(12), 1853–1860. doi: 10.1175/JCLI3460.1
- 1955 Zhang, S., & Wang, B. (2008). Global summer monsoon rainy seasons. *International*
1956 *Journal of Climatology*, *28*, 1563–1578. doi: 10.1002/joc.1659
- 1957 Zhou, W., & Xie, S.-P. (2018). A Hierarchy of Idealized Monsoons in an Interme-
1958 diate GCM. *Journal of Climate*, *31*, 9021–9036. Retrieved from [http://](http://journals.ametsoc.org/doi/10.1175/JCLI-D-18-0084.1)
1959 journals.ametsoc.org/doi/10.1175/JCLI-D-18-0084.1 doi: 10.1175/JCLI-
1960 -D-18-0084.1