Feedback of Atmosphere-Ocean Coupling on Shifts of the Intertropical Convergence Zone

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Abstract. It is well known that the intertropical convergence zone (ITCZ) shifts in response to remote perturbations in the atmospheric energy balance, with shifts roughly in proportion to changes in the cross-equatorial atmospheric energy flux. However, atmospheric and oceanic energy fluxes in low latitudes are mechanically coupled, and the oceanic energy flux dominates the atmospheric energy flux. Here a quantitative framework is derived that shows how Ekman coupling of atmospheric and oceanic energy fluxes damps the perturbation response of the atmospheric energy flux, energy flux equator (EFE), and ITCZ. To first order, Ekman coupling alone mutes the response of EFE and ITCZ in the coupled atmosphere-ocean system by a factor $\gamma = 1 + O_0/\mathrm{NEI}_0$, where O_0 is the ocean energy uptake and NEI_0 is the net energy input into the atmosphere at the equator. In the current climate in the zonal and annual mean, this factor is about $\gamma \approx 3$.

Keypoints:

- Coupling of tropical atmospheric and oceanic energy transport through Ekman balance damps ITCZ shifts in response to energetic perturbations
- Theory shows that the damping depends on the ratio of oceanic to atmospheric energy flux divergence near the equator, among other factors
- The theory accounts for the damping of ITCZ shifts seen in recent simulation studies

1. Introduction

It is by now well established that the intertropical convergence zone (ITCZ) can shift in response to perturbations in the atmospheric energy balance that can be remote, such as perturbations caused by changes in insolation, high-latitude ice cover, or in extratropical or tropical cloud cover (see *Chiang and Friedman* [2012] and *Schneider et al.* [2014] for reviews). For example, if high-latitude albedo increases as a result of expanding Arctic ice sheets, the net energy input into the northern hemisphere diminishes, and the northern hemisphere cools relative to the southern. The atmosphere responds with a partially compensating northward energy flux across the equator [e.g., *Chiang and Bitz*, 2005; *Broccoli et al.*, 2006; *Kang et al.*, 2008, 2009]. The added northward energy flux across the equator is accomplished by a perturbation in the Hadley circulation and a southward ITCZ shift, so that the energy flux by the perturbation component of the Hadley circulation is directed northward across the equator. This way of thinking about how ITCZ shifts relate to the atmospheric energy balance has met considerable quantitative success on time scales longer than a season. For example, shifts in the ITCZ are typically proportional to perturbations in the cross-equatorial atmospheric energy flux, with a proportionality coefficient that can itself be related to the atmospheric energy balance [e.g., *Kang et al.*, 2008, 2009; *Frierson and Hwang*, 2012; *Donohoe et al.*, 2013, 2014; *Bischoff and Schneider*, 2014; *McGee et al.*, 2014; *Adam et al.*, 2016a, b; *Bischoff et al.*, 2017].

However, recent simulation studies have shown that when a dynamic ocean is coupled to the atmosphere, ITCZ shifts in response to perturbations in the atmospheric energy balance are much weaker than they are in an uncoupled atmosphere [*Kay et al.*, 2016;

Hawcroft et al., 2016; *Tomas et al.*, 2016; *Green and Marshall*, 2017]. Such damping of ITCZ shifts was postulated to occur [*Schneider et al.*, 2014], on the basis of the mechanical coupling of atmospheric and oceanic energy fluxes in low latitudes [*Klinger and Marotzke*, 2000; *Held*, 2001]. A similar mechanical coupling of atmospheric and oceanic energy fluxes in the Indian Ocean also damps sea surface temperature (SST) variations during the Asian monsoon [*Lee and Marotzke*, 1998; *Webster et al.*, 1999; *Loschnigg and Webster*, 2000; *Jayne and Marotzke*, 2001; *Webster*, 2006]. The underlying mechanisms have been demonstrated to act in simulations with an idealized coupled general circulation model (GCM) [*Green and Marshall*, 2017]. What is missing is a quantitative theory that links energetic perturbations and ITCZ shifts in the coupled atmosphere-ocean system. This paper offers one component of such a theory, focused on the energy balance.

The point of departure of the theory is a re-casting of known results of how ITCZ shifts are linked to the atmospheric energy balance, a re-casting that makes more explicit the role of oceanic energy fluxes (section 2). A discussion of how perturbations in atmospheric and oceanic energy fluxes in low latitudes are coupled (section 3) then motivates simple quantitative expressions for the damping of ITCZ shifts that results from atmosphere-ocean coupling (section 4). Finally, the results are summarized and their limitations and applicability are discussed (section 5).

2. Relation of ITCZ Latitude to Atmospheric Energy Balance

The atmospheric energy balance in a statistically steady state can be written as

$$\operatorname{div} F = \mathrm{NEI}, \tag{1}$$

where F is the atmospheric energy flux, whose divergence is balanced by the net energy input NEI $= S - L - O$ to the atmosphere on the right-hand side. NEI consists of the net downward shortwave radiation S at the top of the atmosphere, the outgoing longwave radiation L, and ocean energy uptake O. Averaged zonally over a longitude sector sufficiently wide that zonal fluxes into or out of the sector can be neglected, the flux divergence $\operatorname{div} F$ reduces to its meridional component. If, additionally, the atmospheric energy flux F varies approximately linearly with latitude near the equator, as it does in the zonal and annual mean (Fig. 1), first-order Taylor expansion about the equator gives $F(\phi) \approx F_0 + (\operatorname{div} F_0)a\phi$, where a is Earth's radius and the subscript 0 indicates evaluation at the equator. The energy flux equator (EFE) [*Broccoli et al.*, 2006; *Kang et al.*, 2008] is the latitude ϕ_E near the equator where $F(\phi_E) = 0$, which then follows as

$$\phi_E \approx -\frac{1}{a}\frac{F_0}{\mathrm{NEI}_0}. \tag{2}$$

That is, the EFE is proportional to the cross-equatorial atmospheric energy flux F_0 and inversely proportional to NEI at the equator, $\mathrm{NEI}_0 = \operatorname{div} F_0$ [*Bischoff and Schneider*, 2014]. (One could also have linearized the flux F about the EFE instead of the equator, leading to an expression like (2) but with NEI_0 replaced by $\mathrm{NEI}(\phi_E)$ to the same level of approximation.)

If one equates the EFE with the ITCZ latitude $\phi_I \approx \phi_E$ and when $\mathrm{NEI}_0 > 0$, the relation (2) captures how ITCZ shifts $\delta\phi_I \approx \delta\phi_E$ depend on the atmospheric energy balance, at least in a perturbative sense on sufficiently large space and long time scales [e.g., *Schneider et al.*, 2014; *Adam et al.*, 2016a]. However, modifications are necessary and the accuracy of the approximations involved decreases when zonal energy fluxes play a role, or when NEI_0 becomes small, so that the linear approximation breaks down [*Bischoff and Schneider*, 2016; *Adam et al.*, 2016b; *Boos and Korty*, 2016]. Equating the EFE and ITCZ latitude may also be inaccurate when eddy energy fluxes near the equator play a significant role in the energy balance, because one would expect the ITCZ to lie near a zero of the energy flux associated with the divergent mean circulation, not necessarily near a zero of the total energy flux. In such situations, one can improve the accuracy of the approximations by separating out the eddy energy flux from the total flux F and including its divergence in an effective net energy input that is balanced by the mean divergent energy flux [cf. *Adam et al.*, 2016b].

Beyond such potential limits of applicability, the relation (2) obscures how the cross-equatorial atmospheric energy flux F_0 depends on the oceanic energy flux G, whose divergence $\operatorname{div} G = O$ balances the ocean energy uptake in a statistically steady state (i.e., if ocean energy storage can be ignored). As a first step toward making this dependence more explicit, integrate the atmospheric energy balance (1) from the equator to some latitude ϕ_N in the northern hemisphere, and from the equator to some latitude ϕ_S in the southern hemisphere. Averaging the resulting two expressions for the cross-equatorial atmospheric energy flux F_0 gives

$$F_0 = \{(F+G)\cos\phi\}_S^N - G_0 - \left\{ \int_0^{\phi} (S-L)\, a\cos\phi\, d\phi \right\}_S^N, \tag{3}$$

where the braces $\{\cdot\}_S^N$ denote the arithmetic mean of $(\cdot)$ evaluated at ϕ_N and ϕ_S [*Bischoff and Schneider*, 2014]. Thus, if everything else stays fixed, the atmospheric energy flux F_0 across the equator increases as its oceanic counterpart G_0 decreases. The cross-equatorial atmospheric energy flux F_0 also increases with increasing net northward energy export $\{F+G\}_S^N$ by the atmosphere and oceans out of the latitude span $[\phi_S, \phi_N]$. (Note that $F+G$ usually has opposite signs in the northern and southern hemisphere, as it does in

Fig. 1, so $\{F+G\}_S^N > 0$ implies a net northward energy export out of $[\phi_S, \phi_N]$.) Similarly, the cross-equatorial atmospheric energy flux F_0 increases with increasing hemispheric asymmetry of net radiative fluxes $S - L$ at the top of the atmosphere within $[\phi_S, \phi_N]$ (third term on right-hand side).

The relation (3) makes explicit and quantifies what is intuitively clear: the cross-equatorial atmospheric energy flux F_0 in the relation (1) for the EFE or ITCZ latitude depends on the oceanic energy flux G_0 across the equator. To understand how atmosphere-ocean coupling damps EFE and ITCZ shifts, the EFE and the cross-equatorial oceanic energy flux G_0 need to be quantitatively linked. The mechanical coupling of tropical atmosphere and ocean circulations through Ekman balance provides motivation for such a quantitative linkage.

3. Ekman Coupling of Atmosphere and Oceans

In the atmosphere near the surface, the mean zonal winds $\bar{u}$ experience a zonal surface stress τ_A^x (positive sign for eastward stress). The zonal surface stress retards the near-surface winds and so has opposite sign; for example, a simple model is Rayleigh drag with $\tau_A^x \propto -\bar{u}$. The zonal surface stress τ_A^x is associated with a mean meridional Ekman mass flux V_A in the near-surface layer, defined so that the Coriolis force on this ageostrophic Ekman mass flux V_A balances the zonal stress: $fV_A = -\tau_A^x$, with Coriolis parameter f. Averaged zonally over sectors wide enough that zonal pressure gradients (and hence geostrophic mass fluxes) can be neglected, the Ekman mass flux V_A accounts for almost all the total mean meridional mass flux near the surface in the atmosphere. Quantitatively, we can use the local Rossby number $\mathrm{Ro} = -\bar{\zeta}/f$, with relative vorticity ζ, as a measure of the strength of nonlinear modifications of Ekman balance [*Walker and Schneider*, 2006;

Schneider, 2006]. In the annual and zonal mean (on which this paper focuses), reanalysis data show Ro $\ll 0.2$ near the surface outside 2° of the equator [*Schneider et al.*, 2010]. Even in northern summer, when the ITCZ is farthest north and nonlinear modifications are strongest, we still have Ro $\ll 0.2$ poleward of 6°N. Given that horizontal eddy momentum fluxes in the near-surface layer are generally small [e.g., *Green*, 1970; *Schneider et al.*, 2010], it follows that the bulk of the near-surface mean meridional mass flux outside a few degrees of the equator consists of the Ekman mass flux. By implication, the bulk of the near-surface meridional mass flux in the zonal-mean Hadley cells consists of the Ekman mass flux. The Ekman mass flux is directed equatorward ($fV_A < 0$) in regions of surface easterlies ($\bar{u} < 0$ so $\tau_A^x > 0$), for example, over the tropical Atlantic and Pacific Oceans year-round [*Schneider et al.*, 2014]; it is directed poleward ($fV_A > 0$) in regions of surface westerlies ($\bar{u} > 0$ so $\tau_A^x < 0$), for example, in the region of monsoonal westerlies during summer over the Indian Ocean, where the monsoonal mean meridional mass flux is directed from the equator northward toward Asia [e.g., *Gadgil*, 2003; *Webster*, 2006; *Bordoni and Schneider*, 2008].

In the oceans' near-surface layer, the zonal wind stress drives an analogous mean meridional mass Ekman flux V_O, with $fV_O = -\tau_O^x$. By Newton's Third Law, the wind stress τ_O^x driving the ocean Ekman mass flux V_O is equal and opposite to the stress that retards the atmosphere's zonal winds overhead. *Held* [2001] pointed out that if a fraction $0 \leq \xi \leq 1$ of the total stress the atmosphere exerts on the surface in a latitude band is exerted on the oceans, then $\tau_O^x = -\xi\tau_A^x$ and $V_O = -\xi V_A$. Provided the surface stresses over land are a small fraction of the total (i.e., $\xi \approx 1$), the Ekman mass fluxes V_A and V_O in the atmosphere and oceans are approximately equal and opposite. This is so although the densities of air and water differ by a factor 10^3. The Ekman mass flux V_O near the oceans' surface is directed poleward ($fV_O > 0$) where near-surface zonal winds are easterly and the near-surface atmospheric mass flux V_A is directed equatorward ($fV_A < 0$); it is directed equatorward when the reverse holds.

Thus, the strength of the zonal-mean Ekman mass flux V_O in the near-surface oceans is similar to the Ekman mass flux V_A in the near-surface atmosphere, but they have opposite directions. By mass conservation, the oceanic Ekman mass flux near the surface must form the upper branch of an overturning circulation, with divergence of oceanic Ekman mass fluxes near the ITCZ, where atmospheric near-surface mass fluxes converge (Fig. 2). Because the ITCZ is generally located at a moist static energy maximum and usually near an SST maximum [e.g., *Sobel and Neelin*, 2006; *Sobel*, 2007; *Privé and Plumb*, 2007], water masses in the upper ocean that flow away from the ITCZ cool and are subducted along their way toward the Hadley circulation termini, where zonal surface winds turn westerly and the oceanic Ekman mass flux V_O also changes sign. The subducted water masses return below the surface in an approximately adiabatic flow in the thermocline, to close the circulation with upwelling near the ITCZ [*McCreary, Jr. and Lu*, 1994; *Gu and Philander*, 1997; *Klinger and Marotzke*, 2000; *Schott et al.*, 2004]. The resulting overturning cells take up energy in the upwelling region near the ITCZ and transport it toward their subtropical termini, that is, they transport energy in the same direction as the Hadley cells in the atmosphere overhead. This wind-driven overturning circulation with Ekman mass flux V_O in its upper branch accounts for most of the mass and energy fluxes in what are known as the subtropical cells in the tropical and subtropical oceans [*McCreary, Jr. and Lu*, 1994; *Klinger and Marotzke*, 2000; *Held*, 2001; *Talley*, 2003;

Schott et al., 2004]. When upwelling occurs off the equator because the ITCZ is far enough displaced into one hemisphere, as occurs in the Indian Ocean during northern summer, the shallow overturning cells are known as cross-equatorial cells [*Lee and Marotzke*, 1998; *Webster et al.*, 1999; *Loschnigg and Webster*, 2000; *Jayne and Marotzke*, 2001; *Webster*, 2006]. The energy transport associated with the shallow overturning cells dominates the oceanic energy transport in low latitudes. Oceanic energy transports by geostrophic gyres and deep overturning circulations such as the Atlantic Meridional Overturning Circulation (AMOC) also contribute in low latitudes; they dominate in the extratropics. But they are generally weaker in the deep tropics [e.g., *Jayne and Marotzke*, 2001; *Hazeleger et al.*, 2004; *Boccaletti et al.*, 2005; *Czaja and Marshall*, 2006], though they can become significant in the subtropics, especially in the Atlantic [*Talley*, 2003].

If the principal atmosphere-ocean coupling relevant for EFE and ITCZ shifts occurs through the energy transport by the Ekman-coupled atmospheric and oceanic overturning circulations, we can derive an approximation of how the coupling affects the relation between F_0 and G_0. Following *Held* [2001], assume the atmospheric energy flux in low latitudes can be written as $F \approx -S_A V_A$, where S_A is an atmospheric gross stability [*Held and Suarez*, 1978; *Neelin and Held*, 1987], and the minus sign arises because the energy flux is in the direction of the mass flux in the upper branches of the circulation cells for positive gross stability. Write the oceanic energy flux analogously as $G = S_O V_O + \hat{G} \cos^{-1} \phi + \tilde{G}$, where S_O is an oceanic gross stability; $\hat{G}$ is a constant within $[\phi_S, \phi_N]$ so that $\hat{G} \cos^{-1} \phi$ is divergence-free flux in low latitudes, representing, for example, the divergence-free component of AMOC energy transport in low latitudes; and $\tilde{G}$ is the residual oceanic energy flux, e.g., by geostrophic gyres, which is assumed to be nonzero only away from the atmospheric EFE. (The non-Ekman oceanic energy flux can always be written as $\hat{G}\cos^{-1}\phi + \tilde{G}$, simply by choosing $\hat{G}$ to be the non-Ekman energy flux at the EFE, and denoting the residual by $\tilde{G}$.) The coupling relation $V_O = -\xi V_A = \xi F/S_A$ then implies

$$G = \mu F + \hat{G}\cos^{-1}\phi + \tilde{G}. \tag{4}$$

Oceans and atmosphere both have energy flux components F and μF directed away from the EFE and ITCZ, with their ratio given by the factor $\mu = \xi(S_O/S_A)$, which varies with latitude but is $O(1)$ averaged over the tropics [*Held*, 2001]. In particular, and crucially for what follows, the atmospheric and oceanic energy flux components F and μF both vanish at the atmospheric EFE, although the total oceanic energy flux $G(\phi_E) = \hat{G}$ need not vanish there. Like the atmospheric energy flux F, the oceanic energy flux G varies approximately linearly near the equator in the zonal and annual mean (Fig. 1). Linearizing the oceanic energy flux G around the equator analogously to the linearization of the atmospheric energy flux F in section 2 gives $G(\phi) \approx G_0 + (\operatorname{div} G_0)a\phi$, and using that $G(\phi_E) = \hat{G}$ at the EFE yields

$$G_0 \approx \hat{G} - O_0 a\phi_E, \tag{5}$$

because $\operatorname{div} G_0 = O_0$. (As for the atmospheric flux, one could have linearized the oceanic flux G about the EFE instead of the equator, leading to an expression like (5) but with O_0 replaced by $O(\phi_E)$ to the same level of approximation.) The relation (5) for G_0 implies that the farther the EFE ϕ_E is in the northern hemisphere, the weaker is the northward oceanic energy flux G_0 across the equator for given $\hat{G}$ and oceanic energy flux divergence O_0 (slope of the flux G as a function of latitude). Compared with the analogous relation for the atmospheric energy flux,

$$F_0 = -\mathrm{NEI}_0 a \phi_E, \tag{6}$$

it is clear that for an EFE shift $\delta\phi_E$, the ratio of the changes in cross-equatorial oceanic energy fluxes δG_0 to atmospheric energy fluxes δF_0 is $\delta G_0/\delta F_0 \approx O_0/\mathrm{NEI}_0$, provided O_0, NEI_0, and $\hat{G}$ remain approximately fixed under the perturbation considered. Given that $O_0 \approx 48\mathrm{Wm}^{-2}$ and $\mathrm{NEI}_0 \approx 23\mathrm{Wm}^{-2}$ in the annual and zonal mean (Fig. 1), one expects the cross-equatorial oceanic energy flux to respond about $48/23 \approx 2$ times more strongly to EFE shifts than its atmospheric counterpart, for perturbations that leave O_0, NEI_0, and $\hat{G}$ approximately unchanged. Because it is the sum $F_0 + G_0$ of cross-equatorial atmospheric and oceanic energy fluxes that, according to (3), balances hemispherically asymmetric perturbations in the energy balance, a given off-equatorial perturbation can be expected to be balanced primarily by changes in oceanic energy fluxes and only secondarily by changes in atmospheric energy fluxes—which is precisely what is seen in recent simulation studies [*Kay et al.*, 2016; *Hawcroft et al.*, 2016; *Tomas et al.*, 2016; *Green and Marshall*, 2017]. Perturbations in the energy balance that additionally modulate, e.g., the cross-equatorial oceanic energy flux component $\hat{G}$ (e.g., AMOC perturbations, or perturbations in geostrophic gyres) will modify these results, in ways that can be quantitatively captured by quantifying the modulations of $\hat{G}$ in the preceding considerations.

Crucial for the connection between the EFE and the cross-equatorial atmospheric and oceanic energy fluxes is that at the EFE, both the atmospheric energy flux F and the Ekman-coupled oceanic energy flux component μF vanish (Fig 2). This connection exerts a powerful damping feedback on the response of the atmosphere to perturbations that can shift the EFE and ITCZ—a damping feedback that can be expected to act on the turnover timescales of the shallow oceanic overturning circulation, which are of order years [*Gu and Philander*, 1997; *Harper*, 2000].

4. Coupling Feedbacks on ITCZ Shifts

To arrive at a final first-order relation for how the EFE relates to the energy balance when atmosphere and oceans are coupled, substitute the relation (3) for the cross-equatorial atmospheric energy flux into the expression (2) for the EFE and use the relation (5) for the cross-equatorial oceanic energy flux. If we choose the latitudes ϕ_S and ϕ_N as the subtropical termini of the atmospheric Hadley circulation and of the oceanic shallow overturning cells that are in Ekman balance with the zonal wind stress, the oceanic energy flux component μF vanishes at ϕ_S and ϕ_N, and the atmospheric energy flux F there is reduced to an eddy energy flux. Solving for the EFE leads to

$$\phi_E \approx -\frac{1}{a}\,\frac{\left\{(F + \tilde{G})\cos\phi\right\}_S^N - \left\{\int_0^\phi (S - L)\, a\cos\phi\, d\phi\right\}_S^N}{\gamma \mathrm{NEI}_0}, \tag{7}$$

where the coefficient

$$\gamma = 1 + \frac{O_0}{\mathrm{NEI}_0} \tag{8}$$

measures the damping of EFE shifts by Ekman-coupling of atmosphere and oceans. The relation (7) with $\gamma = 1$ and $\tilde{G} = 0$ corresponds to (2) in the absence of oceanic energy fluxes. Because in the annual and zonal mean, both $O_0 > 0$ and $\mathrm{NEI}_0 > 0$ (Fig. 1), the damping coefficient γ is generally greater than one, so that EFE or ITCZ shifts for fixed O_0, NEI_0, and $\tilde{G}$ (e.g., fixed oceanic energy transport associated with geostrophic gyres) are damped in the coupled system relative to an uncoupled atmosphere. In the present climate with $O_0 \approx 48\mathrm{Wm}^{-2}$ and $\mathrm{NEI}_0 \approx 23\mathrm{Wm}^{-2}$ in the annual and zonal mean (Fig. 1), we have $\gamma \approx 3$. That is, ITCZ shifts in the coupled system in response to perturbations, e.g., in high latitudes, can be expected to be about a factor 3 weaker than in an uncoupled atmosphere provided O_0, NEI_0, and $\tilde{G}$ (e.g., ocean energy transport by geostrophic gyres) remain approximately unchanged. This is consistent with what is seen in recent simulation studies [*Kay et al.*, 2016; *Hawcroft et al.*, 2016; *Tomas et al.*, 2016; *Green and Marshall*, 2017].

A few comments are in order:

- The cross-equatorial oceanic energy flux component $\hat{G}$ does not appear explicitly in relation (7). It was assumed that $\hat{G}$ is constant so that the flux $\hat{G}\cos^{-1}\phi$ remains divergence-free throughout the latitude span $[\phi_S, \phi_N]$. Thus, its contribution to the first term in the numerator of (7) and its direct contribution to G_0 cancel each other. However, such a flux (e.g., the divergence-free component of tropical AMOC energy transport) can still affect the EFE and ITCZ. It can warm one hemisphere relative to another and lead to an asymmetry in atmospheric energy fluxes at ϕ_S and ϕ_N, which can be related to SST gradients through flux closures [*Bischoff and Schneider*, 2014]. Such an asymmetric atmospheric energy flux can displace the EFE and ITCZ via the first term in the numerator of (7). For example, the northward AMOC energy transport gives rise to $G_0 > 0$ in Earth's climate (Fig 1) and displaces the EFE and ITCZ north of the equator in the mean because it engenders a southward atmospheric eddy energy export $\{F\cos\phi\}_S^N < 0$ out of the tropics [e.g., *Marshall et al.*, 2014; *Frierson et al.*, 2013]. For a given atmospheric eddy energy export out of the tropics, such displacements are likewise weaker by a factor γ in the Ekman-coupled system than in an uncoupled atmosphere.

- If atmospheric and oceanic energy fluxes F and G had been linearized about the EFE instead of the equator, $\mathrm{NEI}(\phi_E)$ and $O(\phi_E)$ would appear in relations (7) and (8) in place of NEI_0 and O_0. Such modified relations would be equally valid as (7) and (8) to first order. This shows that there is some arbitrariness about where NEI and O are evaluated. The assumption is that they do not vary strongly between the equator and the EFE, so that a first-order approximation is useful, but this may not always hold accurately.

- For the relation (7) to capture the dependencies of the EFE on the energetic quantities on the right-hand side explicitly, NEI_0 and O_0 should be at most weakly dependent on ϕ_E to first order. Both can be expected to change as tropical mean temperatures change [e.g., *Adam et al.*, 2017]. However, for perturbations that do not change the mean temperature, O_0 and O_0/NEI_0 will primarily vary with oceanic factors such as the width of ocean upwelling [*Held*, 2001; *Levine and Schneider*, 2011], which can plausibly be taken to be independent of ϕ_E to first order. The simulations of *Green and Marshall* [2017] suggest that this is approximately true: they found modulations of NEI_0 in response to extratropical energetic perturbations to be negligible relative to the modulations of the energy fluxes themselves. In other words, it appears justified to take the divergences $\text{NEI}_0 = \operatorname{div} F_0$ and $O_0 = \operatorname{div} G_0$ in the linearization of the fluxes F and G as independent of ϕ_E to first order. However, even when its right-hand side does depend on ϕ_E, the relation (7) remains valid. It then only is an implicit relation for ϕ_E, as is the uncoupled relation (2), which likewise remains valid but whose right-hand side depends implicitly on ϕ_E.

- Perturbations are also conceivable that couple EFE and ITCZ shifts to the non-Ekman oceanic energy flux component $\tilde{G}$, for example, by shifts in zonal wind patterns that lead to perturbations in zonal wind stress curls and oceanic subtropical gyres. Even for such perturbations, the relation (7) with damping coefficient γ continues to hold as long as the linearization of the fluxes F and G remains adequate. But the resulting atmosphere-ocean coupling effects are only implicitly captured in (7) if, for example, the oceanic energy export out of the tropics, $\left\{\tilde{G}\cos\phi\right\}_S^N$, changes. To make such coupling effects explicit, a theory for the dependence of $\tilde{G}$ on the EFE (to first order) would need to be added. Similarly, nonlinear modifications of Ekman balance in the atmosphere may be significant near the equator and ITCZ [*Waliser and Somerville*, 1994; *Schneider and Bordoni*, 2008; *Gonzalez et al.*, 2016] and can affect ocean upwelling and hence $\tilde{G}$. Capturing such additional effects explicitly in a more general counterpart to (7) is left for future work.

- The value of $\mathrm{NEI}_0 = S_0 - L_0 - O_0$ is not precisely known because it is a small residual of large terms with measurement uncertainties of several Wm^{-2}: $S_0 \approx 323\mathrm{Wm}^{-2}$, $L_0 \approx 251\mathrm{Wm}^{-2}$, and $O_0 \approx 48\mathrm{Wm}^{-2}$ in the annual and zonal mean [*Fasullo and Trenberth*, 2008; *Loeb et al.*, 2009]. Additionally, both NEI_0 and O_0 vary near the equator, so their values vary by several Wm^{-2} depending on whether one evaluates them on the equator or averages within a few degrees of the equator. Additionally, both the equatorial net energy input NEI_0 and the ocean energy uptake O_0 differ by up to $\pm 10\mathrm{Wm}^{-2}$ across current climate models [*Adam et al.*, 2016c, 2017]. As a result, the damping coefficient $\gamma = 1 + O_0/\mathrm{NEI}_0$ is not known precisely from observations and varies from climate model to climate model. For example, across the comprehensive climate models participating in the Coupled Model Intercomparison Project Phase 5 (CMIP5), with O_0 and NEI_0 averaged between $\pm 5°$, the damping coefficient has a mean value of $\gamma = 2.8$ and standard deviation 0.5. This spread, in addition to factors that may modulate $\tilde{G}$ differently in different perturbation experiments, may give rise to quantitative differences in the degree of damping of ITCZ shifts seen in different simulation studies [e.g., *Kay et al.*, 2016; *Hawcroft et al.*, 2016; *Tomas et al.*, 2016; *Green and Marshall*, 2017].

The relation (7) for the EFE can be used in the same way in which its uncoupled counterpart (2) has been used to conceptualize ITCZ shifts $\delta\phi_I \approx \delta\phi_E$. For example, consider a brightening of clouds over the Southern Ocean, such as that simulated by *Kay et al.* [2016] and *Hawcroft et al.* [2016]. Assume that this perturbation does not affect substantially geostrophic ocean gyres and other ocean circulations that contribute to $\tilde{G}$ at the subtropical latitudes ϕ_N and ϕ_S. If the atmospheric shortwave response (e.g., cloud changes) in the tropics and the longwave response (e.g., temperature and cloud changes) are negligible or symmetric about the equator, changes in the hemispheric asymmetry of $S - L$ in the tropics can be neglected,

$$\delta \left\{ \int_0^y (S - L)\, dy \right\}_S^N \approx 0. \tag{9}$$

This is a reasonable assumption because angular momentum constraints force free-tropospheric temperatures to remain nearly symmetric about the equator even in strongly hemispherically asymmetric circulations [*Lindzen and Hou*, 1988; *Schneider and Bordoni*, 2008; *Schneider et al.*, 2014]. Cloud feedbacks on ITCZ shifts may break the symmetry, but because shortwave and longwave radiative effects of deep clouds nearly cancel each other [*Harrison et al.*, 1990; *Loeb et al.*, 2009], the net effect of feedbacks from the deep ITCZ clouds may be weak, though this is uncertain. Feedbacks from lower clouds may affect the degree of hemispheric symmetry of $S - L$ [*Kang et al.*, 2008]. However, the overall sign of cloud feedbacks on ITCZ shifts is uncertain [*Voigt et al.*, 2014]. Neglecting any such changes for now (their effects can be added when they become better known) leads to

$$\delta\phi_E \approx -\frac{1}{a}\,\frac{\{\delta F\cos\phi\}_S^N}{\gamma \mathrm{NEI}_0} - \frac{\delta(\gamma \mathrm{NEI}_0)}{\gamma \mathrm{NEI}_0}\phi_E. \tag{10}$$

If the second term is negligible, as it is, for example, in the simulations of *Green and Marshall* [2017], the primary effect of cloud brightening over the Southern Ocean is an increased pole-equator temperature contrast in the southern hemisphere. A southward eddy energy export perturbation $\{\delta F\cos\phi\}_S^N < 0$ out of the tropics follows in response [*Kay et al.*, 2016; *Hawcroft et al.*, 2016], implying a northward EFE shift $\delta\phi_E > 0$. However, the shift is a factor $\gamma \approx 3$ weaker in the Ekman-coupled system than in an uncoupled atmosphere because any small shift $\delta\phi_E$ in the coupled system leads primarily to an oceanic energy flux perturbation $\delta G_0 \approx -O_0 a\delta\phi_E$, and only secondarily to an atmospheric energy flux perturbation $\delta F_0 \approx -\mathrm{NEI}_0\, a\delta\phi_E$, because $O_0/\mathrm{NEI}_0 \approx 2$. (It is again assumed that NEI_0 and O_0 perturbations can be neglected. In that case, the flux perturbations correspond to $\delta\phi_E$ translations in latitude of the energy fluxes F and G near the equator.) However, it is the perturbation in the cross-equatorial atmospheric energy flux F_0 that determines the magnitude of the EFE and ITCZ shifts, which hence is muted in the coupled system. These dynamics are seen to play out in the simulations of *Kay et al.* [2016] and *Hawcroft et al.* [2016], with slightly different partitionings between oceanic and atmospheric energy flux responses, probably because of model differences in O_0 and NEI_0.

5. Discussion and Conclusions

In summary, the components of the atmospheric energy flux and of the oceanic energy flux that are in Ekman balance near surface are mechanically coupled to one another [*Klinger and Marotzke*, 2000; *Held*, 2001], and this mechanical coupling damps EFE and ITCZ shifts, as previously discussed in qualitative terms and illustrated in simulations

[*Schneider et al.*, 2014; *Green and Marshall*, 2017]. A result of this coupling is that although the low-latitude zeros of the total atmospheric and oceanic energy fluxes do not necessarily coincide, the low-latitude zeros of the Ekman components of the energy fluxes do coincide (Fig. 1), so both shift when the atmospheric EFE shifts. The adjustments in the oceans occur on timescales of years, over which the shallow ocean overturning circulations respond [*Gu and Philander*, 1997; *Harper*, 2000; *Green and Marshall*, 2017]. Using this result and linearizing atmospheric and oceanic energy fluxes around the equator led to a quantitative relation (7) that shows how the atmospheric EFE depends on energetic quantities such as the energy export out of the tropics, or the ocean energy uptake and the net energy input to the atmosphere near the equator. Unlike its previously derived counterpart (2), this relation makes explicit the dependence of the cross-equatorial oceanic energy flux on the EFE through Ekman coupling, and it shows how this dependence damps EFE and ITCZ shifts in response to perturbations. In essence, hemispherically asymmetric perturbations in the energy balance are accommodated by the sum of cross-equatorial atmospheric and oceanic energy fluxes. Smaller EFE and ITCZ shifts are needed to accomplish the cross-equatorial energy fluxes when both atmosphere and oceans can adjust than when the adjustment is confined to the atmosphere alone. The sensitivity of the cross-equatorial atmospheric energy flux to EFE shifts is given by the equatorial net energy input NEI_0, and that of the cross-equatorial oceanic energy flux by the equatorial ocean energy uptake O_0. EFE and ITCZ shifts in the coupled system are damped by a factor $\gamma = 1 + O_0/\mathrm{NEI}_0$, which is approximately 3 in the current climate in the annual and zonal mean. This is consistent with the damping of ITCZ shifts seen in recent simulations studies.

These results can be generalized in straightforward ways, analogously to how their atmosphere-only counterparts have been generalized previously. For example, one can explicitly take time-dependence and energy storage in the atmosphere and oceans into account, or zonal energy fluxes into or out of a longitude sector [cf. *Adam et al.*, 2016a, b; *Boos and Korty*, 2016]. Or one may consider higher-order generalizations that take non-linear dependence of energy fluxes on latitude near the equator into account [*Bischoff and Schneider*, 2016; *Adam et al.*, 2016b]. While such generalizations are straightforward, some limitations of the results will remain, most of which being generic to any account of the EFE and ITCZ that focuses on the energy balance alone:

- Identifying the ITCZ with the EFE, or ITCZ shifts with EFE shifts, is not always accurate [*Chiang and Friedman*, 2012; *Donohoe et al.*, 2013; *Adam et al.*, 2016b].

- Quantities such as the equatorial net energy input NEI_0 and equatorial ocean energy uptake O_0 appear in the results from the energy balance, but they themselves depend on the atmospheric and oceanic circulations. Thus, this theoretical account is not closed. Closure requires theories for the gross stabilities of atmosphere and oceans, such as those offered by *Held* [2001], and a theory for the Hadley circulation, which remains an outstanding challenge [*Schneider et al.*, 2010].

Such limitations notwithstanding, the expressions derived here capture quantitatively the damping of EFE and ITCZ shifts owing to Ekman coupling of atmosphere and oceans. Thus, they represent one step toward a future, more complete theory of the ITCZ and tropical atmosphere-ocean coupling.

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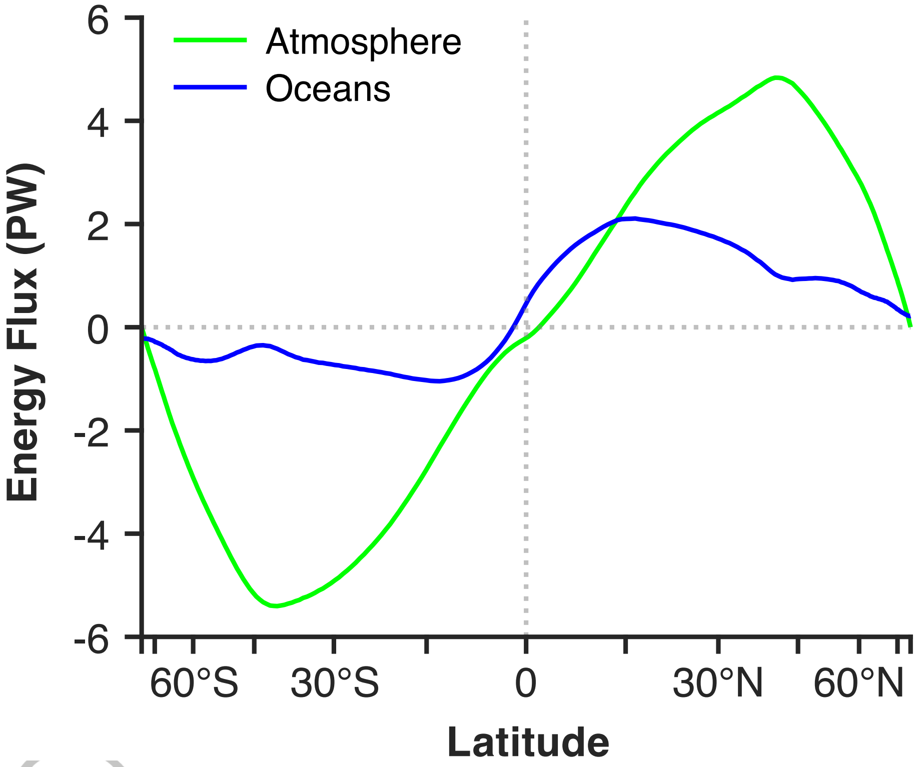

Figure 1. Atmospheric energy flux F (green) and oceanic energy flux G (blue) in the annual and zonal mean for 2001-2014. On the latitude axis, fixed increments are proportional to $\sin \phi$, so that equal increments correspond to equal area increments. The atmospheric energy flux is from ERA Interim reanalysis [*Dee et al.*, 2011], with corrections to ensure mass balance following *Fasullo and Trenberth* [2008]. The oceanic energy flux is calculated by subtracting the atmospheric energy flux from the radiative imbalance at the top of the atmosphere derived within NASA's Clouds and the Earth's Radiant Energy System (CERES) project [*Loeb et al.*, 2009]. Near the equator (averaged between $\pm 2.5°$ latitude), the divergences of the fluxes are $\mathrm{NEI}_0 = \mathrm{div}\, F_0 \approx 23\mathrm{Wm}^{-2}$ and $O_0 = \mathrm{div}\, G_0 \approx 48\mathrm{Wm}^{-2}$.

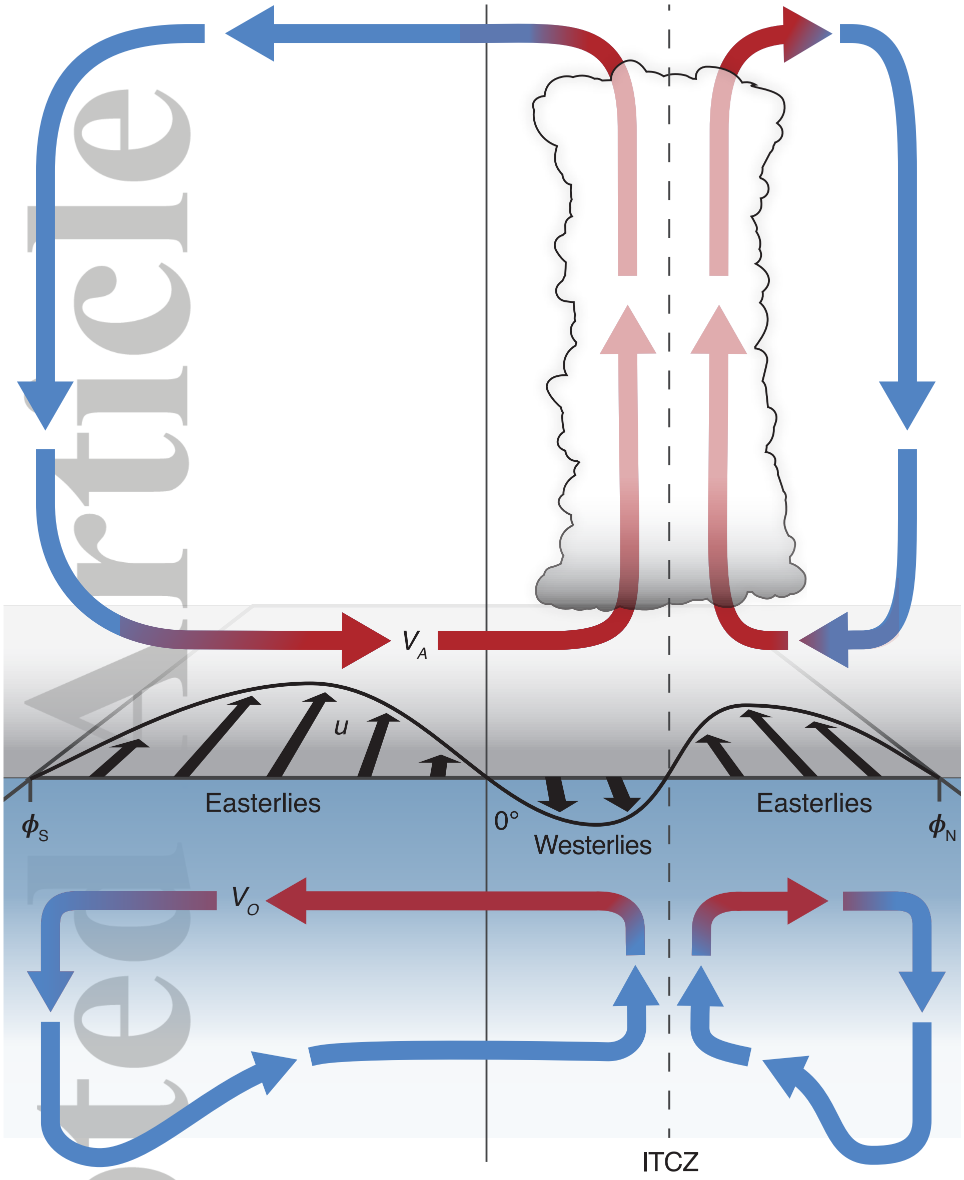

Figure 2. Schematic of Ekman-coupled atmosphere and ocean circulations in low latitudes. The Coriolis force on the meridional Ekman mass flux V_A in the near-surface atmosphere balances the surface stress on the zonal winds $\bar{u}$, implying equatorward mass flux V_A in regions of easterlies, and poleward mass flux V_A in regions of westerlies. In the oceans, the zonal surface wind stress drives a meridional Ekman mass flux V_O, which, by Newton's Third Law, has the opposite direction and is of similar magnitude as the atmospheric mass flux V_A. The water masses transported meridionally in the near-surface ocean cool and are subducted along their way toward the subtropics, returning to the upwelling region near the ITCZ in the thermocline. The resulting shallow overturning circulation in the oceans and the Hadley circulation in the atmosphere both transport energy in the direction of their upward branches, i.e., generally away from the ITCZ. Their coupled energy transport damps ITCZ shifts in response to energetic perturbations relative to the response in an uncoupled atmosphere because both atmospheric and Ekman-coupled oceanic energy transports respond similarly to perturbations. Note that other ocean circulation components such as geostrophic gyres or more complex nonlinear equatorial dynamics may also modulate ITCZ shifts but are not explicitly considered here. (After *Schneider et al.* [2014].)