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The Atmospheric General Circulation in Thermodynamical Coordinates

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ABSTRACT

7 The zonal and meridional components of the atmospheric general circulation are used to
 8 define a global thermodynamic stream function in dry static energy versus latent heat coor-
 9 dinates. Diabatic motions in the tropical circulations and fluxes driven by midlatitude eddies
 10 are found to form a single, global thermodynamic cycle. Calculations based on the ERA-
 11 Interim reanalysis dataset indicate that the cycle has a peak transport of 428 Sv ($\text{Sv} = 10^9$
 12 kg s^{-1}). The thermodynamic cycle encapsulates a globally interconnected heat and water
 13 cycle comprising ascent of moist air where latent heat is converted into dry static energy,
 14 radiative cooling where dry air loses dry static energy, and a moistening branch where air is
 15 warmed and moistened. It approximately follows a tropical moist adiabat and is bounded by
 16 the Clausius-Clapeyron relationship for near-surface air. The variability of the atmospheric
 17 general circulation is related to ENSO events using reanalysis data from recent years (1979-
 18 2009) and historical simulations from the EC-Earth coupled climate model (1850-2005). The
 19 thermodynamic cycle in both EC-Earth and ERA-Interim widens and weakens with posi-
 20 tive ENSO phases and narrows and strengthens during negative ENSO phases with a high
 21 correlation coefficient. Weakening in amplitude suggests a reduction in moist convection in
 22 the tropics, while widening suggests an increase in mean tropical near-surface moist static
 23 energy.

1. Introduction

The atmospheric general circulation forms as a response to differential solar heating and oceanic heat transport, acting to redistribute energy from warm regions to cold regions, mainly through the exchange of dry static energy (DSE) and latent heat (LH) (Trenberth and Stepaniak 2003a). The geographical distribution of these exchanges dictates, among other climatic features, global patterns of precipitation and evaporation (Muller and O’Gorman 2011). The atmospheric circulation is known to vary as the driving forces (solar heating and ocean heat fluxes) vary.

El Niño–Southern Oscillation (ENSO) is one of the dominant modes of inter-annual variability in the climate system (Lu et al. 2008). It is characterised by weakened / strengthened winds and zonal gradients of sea-surface temperature (SST) over the Pacific during positive/negative ENSO phases (Bjerknes 1969; Oort and Yienger 1996). It has been shown that ENSO events also affect tropical-mean surface temperatures (Seager et al. 2003) and the meridional overturning circulation (Lu et al. 2008). The Walker circulation is a zonally asymmetric feature of the total circulation that plays a key role in ENSO events (Bjerknes 1969) and is closely tied to the zonal SST gradient in the tropical Pacific. It redistributes DSE and LH zonally (Trenberth and Stepaniak 2003b) and is driven by convection over the western tropical Pacific and by subsidence over the east Pacific and the Atlantic (Peixoto and Oort 1992). Because of its East-West orientation, the Walker circulation is not captured by zonal-mean diagnostics of the meridional circulation. It is often viewed schematically and seldom studied using closed stream functions, since its poleward extent is not well defined. This makes most definitions of the Walker circulation strength ambiguous (attempts include Trenberth et al. (2000), Vecchi et al. (2006), Tokinaga et al. (2012) and L’Heureux et al. (2013)).

The meridional transport of dry and moist static energy (MSE, equal to the sum of DSE

and LH) can be represented by the zonal-mean meridional overturning circulations in dry and moist isentropic coordinates (Fig. 1) (Townsend and Johnson 1985; Karoly et al. 1997; Held and Schneider 1999; Pauluis et al. 2008; Döös and Nilsson 2011). DSE is approximately conserved for dry isentropic motions (adiabatic motions, where potential temperature, θ , is constant). In the same way, MSE is almost conserved for moist isentropic motions (moist adiabatic motions, where the equivalent (or “moist”) potential temperature, θ_e , is constant) (Emanuel 1994). Following Döös and Nilsson (2011), we compute the meridional overturning circulations in isobaric, dry isentropic and moist isentropic coordinates (Fig. 1, and eq. A1) using 6-hourly ERA-Interim data. Details of the calculations are presented in Appendix A. In the midlatitudes, mass fluxes by synoptic eddies are lost when zonally averaging in isobaric coordinates, but included when using dry or moist isentropic coordinates (Peixoto and Oort 1992; McIntosh and McDougall 1996; Held and Schneider 1999; Pauluis et al. 2008; Döös and Nilsson 2011; Laliberté et al. 2012; Kjellsson and Döös 2012). Furthermore, moist isentropic coordinates includes fluxes of moist air that cancel in dry isentropic coordinates (Pauluis et al. 2010). As a result, in the midlatitudes the dry and moist isentropic circulations are much stronger than the isobaric circulation, and the moist isentropic circulation is stronger than the dry isentropic, which can be seen in Fig. 1. This indicates the importance of considering both DSE and LH when studying the global atmospheric circulation.

Variations of the atmospheric general circulation can occur in both the meridional and zonal components. There are however few diagnostics that include both zonal and meridional overturning circulations. A similar problem exists in the ocean. Recent studies (Döös et al. 2012; Zika et al. 2012) have made use of purely thermodynamic (temperature–salinity) coordinates to represent the circulation in a single picture that describes global exchanges of heat and freshwater as well as water-mass transformations. The unified picture then allows for analysis of the interconnected nature and variations of the global ocean circulation (Ballarotta 2013; Zika et al. 2013). In this study we use a similar technique to explore the

redistribution of DSE and LH in the atmospheric general circulation. Previous studies have investigated the atmospheric circulation in θ - θ_e (potential and equivalent potential temperature, similar to DSE-MSE) coordinates (Pauluis et al. 2008; Laliberté et al. 2012, 2013a), but only included meridional mass fluxes. In a recent study, Pauluis and Mrowiec (2013) examined the circulation in z - θ_e coordinates using a cloud-resolving model, but this only included vertical mass fluxes. We include mass fluxes in the zonal, meridional and vertical directions to obtain a new description of the global atmospheric circulation as a cycle in purely thermodynamic coordinates (LH-DSE space). This new description gives a unified picture of the zonal and meridional circulations and allows us to estimate the strength of its zonally asymmetric component. We diagnose the variability of the global thermodynamic cycle with a focus on annual and inter-annual variations and the impacts of ENSO, and also compare the results from ERA-Interim reanalysis and the EC-Earth climate model.

2. Data & Method

We use 6-hourly surface pressure, p_s , temperature, T , specific humidity, q , and horizontal winds, u, v , on model levels from ERA-Interim reanalysis (Berrisford et al. 2009) for the period 1979-2009. We use data with 1.25° horizontal resolution and 60 model levels, of which roughly 30 are in the troposphere. Using surface pressure, the pressure at each model level interface, k , is $p_k = A_k + B_k p_s$, where A_k and B_k are constants (Simmons and Burridge 1981). The geopotential height, z [m], is calculated by integrating the hydrostatic equation vertically using temperature, specific humidity, and pressure. Vertical velocity, ω [Pa s^{-1}], is calculated using the model continuity equation following Simmons and Burridge (1981) (see Kjellsson and Döös (2012) for a similar derivation). We wish to stress that the calculated vertical velocity does not include any explicit parameterization of convection. Rather, it is a diagnostic of horizontal motions at each level. The vertical mass flux is therefore realistic in spatial coordinates. However, the LH and DSE values in a grid box may be different from

those that would have been in a convective updraft if convection was fully resolved. In such a case the mass fluxes in l, s space may be positioned differently and the integral, $\psi(l, s)$, could be different. The implications of this is discussed in Section 6 of this paper.

We also make use of atmospheric data from historical simulations of years 1850–2005 from the EC-Earth coupled climate model (Hazeleger et al. 2011). The atmospheric component of the EC-Earth model, IFS (Integrated Forecasting System), is also used by ECMWF when generating the ERA-Interim reanalysis. After more than 1000 years of coupled spin-up integration where radiative forcing and atmospheric composition (including aerosols) are held constant at pre-industrial conditions, the model is integrated with observed or reconstructed radiative forcing and atmospheric composition (Meinshausen et al. 2011) from the years 1850–2005. The data has a 1.125° horizontal resolution, 62 model levels and is stored every 6 hours. As for the ERA-Interim data, we use u, v, T, q, p_s to calculate z and ω .

Using the six variables u, v, T, q, z, p we calculate LH, $l = L_v q$ [J kg^{-1}], DSE, $s = c_p T + gz$ [J kg^{-1}], and MSE, $h = l + s$ [J kg^{-1}], where $L_v = 2.5 \cdot 10^6$ J kg^{-1} is the latent heat of vaporization, $c_p = 1004$ $\text{J kg}^{-1} \text{K}^{-1}$ is the specific heat capacity of dry air, and $g = 9.81$ m s^{-2} is the gravitational acceleration at the Earth’s surface (Emanuel 1994; Wallace and Hobbs 2006).

We begin by binning the material derivative of DSE, $\dot{s} = ds/dt = \partial_t s + \vec{v} \cdot \nabla s$, from the eulerian space into a LH–DSE space:

$$\dot{S}(l, s) = \frac{1}{t_1 - t_0} \int_{t_0}^{t_1} \int_{\Omega} \delta[l - l'(x, y, p, t)] \delta[s - s'(x, y, p, t)] [\partial_t s'(x, y, p, t) + \vec{v} \cdot \nabla s'(x, y, p, t)] dM dt, \quad (1)$$

where $dM = g^{-1} dx dy dp$ is a mass element, $\delta(\phi)$ is the Dirac delta function ($\delta[\phi] = 1$ when $\phi = 0$ and 0 otherwise) and primed variables are in x, y, p space. The result is shown in Fig. 2a and the method to obtain it is described in appendix B. We then apply the same binning

procedure to the material derivative of LH, $\dot{l} = dl/dt = \partial_t l + \vec{v} \cdot \nabla l$:

$$\dot{L}(l, s) = \frac{1}{t_1 - t_0} \int_{t_0}^{t_1} \int_{\Omega} \delta[l - l'(x, y, p, t)] \delta[s - s'(x, y, p, t)] [\partial_t l'(x, y, p, t) + \vec{v} \cdot \nabla l'(x, y, p, t)] dM dt. \quad (2)$$

In eqs. (1) and (2), $\dot{S}$ and $\dot{L}$ are mass integrals of $\dot{s}$ and $\dot{l}$ respectively. Both have the unit $[\text{kg s}^{-1} (\text{kJ kg}^{-1})^{-1}]$. It can be shown that with these definitions $\partial_s \dot{S}(l, s) + \partial_l \dot{L}(l, s) = 0$ and therefore the binning of DSE tendencies is associated with a closed and unique stream function, i.e.

$$\psi^{\text{tot}}(l, s) = \int_0^l \dot{S}(l', s) dl'. \quad (3)$$

As global atmospheric circulations tend to have magnitudes of $\sim 10^{10} \text{ kg s}^{-1}$ (cf. Fig. 1) we will present stream functions in units of Sverdrup, where $1 \text{ Sv} = 10^9 \text{ kg s}^{-1}$.

The stream function, $\psi^{\text{tot}}(l, s)$, represents mass fluxes across a DSE surface and is due to both the movement of the surface (a local eulerian change, $\partial_t s$) and the transport of air from one DSE value to another (advective change, $\vec{v} \cdot \nabla s$). In eq. (3), there are thus two effects that are binned onto the l, s plane. The total stream function can therefore be seen as the sum of two stream functions, $\psi^{\text{tot}} = \psi^{\text{loc}} + \psi^{\text{adv}}$. The former stream function, ψ^{loc} is a stream function based only on local eulerian changes, $\partial_t s$. It represents air at a certain point in x, y, p coordinates changing its LH and/or DSE from one time to another. In particular, it will be non-divergent if all x, y, p points have the same initial and final positions in l, s coordinates. The latter stream function, ψ^{adv} , is based on advection of DSE. At any instant it represents global mass fluxes projected onto the LH–DSE plane. For ψ^{adv} non-closed stream lines arises only in the presence of sources and sinks of mass. Sources and sinks of mass are present in our data since neither reanalysis nor climate models are perfectly mass conserving (Berrisford et al. 2011). We also make the important note that for a purely adiabatic flow, $\psi^{\text{tot}} = 0$, in which case ψ^{adv} and ψ^{loc} are either both zero or equal and of opposite sign.

We show ψ^{loc} and ψ^{adv} calculated from ERA-Interim data 1979-2009 in Fig. 2b and Fig. 2c, respectively. In Fig. 2c, not all stream lines close. Unclosed streamlines mean that

there are net mass fluxes across some DSE surfaces. The maximum mass flux outside closed stream lines is small and corresponds to a maximum error in mass conservation of $< 5\%$ of the amplitude. In Fig. 2b, there are also errors in ψ^{loc} corresponding to $\sim 13\%$ of the amplitude. These errors are mostly due to the fact that a closed circulation in ψ^{loc} can only be obtained if a DSE surface returns to its initial position at the end. We find that ψ^{adv} contributes much more to the stream function ψ^{tot} than ψ^{loc} does. For this reason, we will assume that for the data we are analyzing:

$$\psi^{\text{tot}} \approx \psi^{\text{adv}}. \quad (4)$$

However we do not argue that this approximation holds generally, e.g. for different data sets.

We denote the stream function ψ^{adv} as the *hydrothermal* stream function to be consistent with terminology used for the oceanic thermohaline stream function in temperature–salinity coordinates in previous studies (Zika et al. (2012); Döös et al. (2012); a quantification of ψ^{adv} is presented in Groeskamp et al. (2013)). We will hereafter drop the superscript and denote the hydrothermal stream function as $\psi(l, s)$. The hydrothermal stream function can be expressed succinctly as

$$\psi(l, s) = \frac{1}{t_1 - t_0} \int_{t_0}^{t_1} \int_{\Omega} \mu[l - l'(x, y, p, t)] \delta[s - s'(x, y, p, t)] \vec{v} \cdot \nabla s'(x, y, p, t) dM dt, \quad (5)$$

where $\mu[l - l']$ is the Heaviside function which is the integral of the Dirac delta function, $\delta[l - l']$. The possible effects of including ψ^{loc} in the analysis will be discussed in Section 6 of the paper.

3. The Hydrothermal Cycle

We show the hydrothermal stream function, $\psi(l, s)$, calculated from ERA-Interim data 1979-2009 Fig. 2c. The result is a single anti-clockwise cycle with an amplitude of -428

Sv. We define the hydrothermal stream function maximum as the difference between the maximum and minimum value along the LH axis for each value of DSE (Fig. 2d) and note that the maximum value occurs at $\text{DSE} \approx 320 \text{ kJ kg}^{-1}$.

The outer streamlines (*e.g.* -30 Sv in Fig. 2c) of the hydrothermal stream function, $\psi(l, s)$, have three branches that share characteristics with different atmospheric processes. One branch extends from $(l = 45 \text{ kJ kg}^{-1}, s = 300 \text{ kJ kg}^{-1})$ to $(l = 0 \text{ kJ kg}^{-1}, s = 345 \text{ kJ kg}^{-1})$ and is bounded by a moist adiabat (constant MSE, $h \approx 345 \text{ kJ kg}^{-1}$), shown in Figs. 2a,b,c as a dashed line. It exhibits the behaviour of ascent with precipitation where LH is converted into DSE such that MSE is approximately conserved (Emanuel 1994). It is worth noting that this branch has a minimum of MSE at about $(l = 15 \text{ kJ kg}^{-1}, s = 320 \text{ kJ kg}^{-1})$. We speculate that convective detrainment of convecting near-surface tropical air and its subsequent mixing with drier air in the tropical midtroposphere contribute to this minimum. A similar minimum was also found by Pauluis and Mrowiec (2013). We also identify a branch where dry air with $l \sim 0 \text{ kJ kg}^{-1}$ loses DSE through radiative cooling (Peixoto and Oort 1992) and we speculate that it is associated with large-scale subsiding motions. This branch is bounded to the left by the limit of completely dry air ($l = 0 \text{ kJ kg}^{-1}$). The third branch extends from $(l = 0 \text{ kJ kg}^{-1}, s = 260 \text{ kJ kg}^{-1})$ to $(l = 45 \text{ kJ kg}^{-1}, s = 300 \text{ kJ kg}^{-1})$ and is bounded by a curve representing saturated air at the surface. This curve shows the LH and DSE of saturated surface air which can be derived from the Clausius–Clapeyron relationship as discussed in Appendix C. In this branch, cold and dry lower tropospheric air masses are warmed and moistened as they are mixed with near-surface air (a phenomenon similar to that described by Laliberté et al. (2012)). However, following a stream line at $\sim 400 \text{ Sv}$ we find that moistening of dry air also occurs at relatively high DSE values ($\sim 315 \text{ kJ kg}^{-1}$), which may not be near-surface air but rather mixing with air moistened by shallow convection. Given these processes and for the remainder of this work, we will refer to the first, second and third branches as the precipitating, cooling and moistening branches, respectively.

We now estimate the spatial structure of the hydrothermal cycle by projecting the stream function onto the time-averaged location of air masses. We select air masses between the -100 and -400 Sv stream lines of the hydrothermal cycle and divide them into sections similar to the hours on a clock (Döös et al. 2012) centered at the minimum value of the stream function, $\min_{l,s}(\psi)$. Each section is coloured such that 12 o'clock (green) is high DSE, 6 o'clock (dark blue) is low DSE, 3 o'clock (purple) is high LH, and 9 o'clock (light blue) is low LH. We then project these sections onto the time-averaged three-dimensional (longitude-latitude-pressure) LH and DSE values for ERA-Interim data (Fig. 3). The zonal mean of the projection (Fig. 3a) shows that the precipitating branch of the hydrothermal cycle occurs in the tropics, the cooling branch at high altitudes, and the moistening branch at low altitudes in midlatitudes and the subtropics. The hydrothermal cycle thus includes a meridional overturning cell in each hemisphere with ascent in the tropical regions and subsidence almost everywhere polewards of 30° . The reader should note that Fig. 3 is a projection of a thermodynamic stream function, which represents both mean and time varying processes, into a time-averaged atmospheric state that does not take temporal variations into account. In particular, the projection neglects seasonal changes and cannot represent convection associated with cyclones in the midlatitude storm tracks (Pauluis et al. 2010), which play an important role in the meridional overturning circulation. Given this important caveat, these results are in qualitative agreement with the circulations that have been described using isentropic-mean or residual-mean frameworks (Karoly et al. 1997; Held and Schneider 1999; Juckes 2001).

A slice through the midtroposphere (the 570 hPa level is shown in Fig. 3b) shows that the precipitating branch is located over the equatorial west Pacific warm pool and the Inter-Tropical Convergence Zone (ITCZ), where we generally find precipitation and heating across the column from LH release (Källberg et al. 2005). In the tropics, the cooling branch is found over the equatorial Atlantic and east Pacific, regions that are associated with cooling from

radiative cooling. The zonal asymmetries in ascending and subsiding motions are associated with the Walker circulation and give a sense of its poleward extent. Note that in midlatitudes and polar regions, the cooling branch is highly zonally symmetric.

As previously noted, the maximum mass flux of the meridional overturning circulation in moist isotropic coordinates is stronger than in that in dry isentropic coordinates (Fig. 1). We calculate the combined maximum mass flux of the meridional overturning circulation as the difference between the maximum and minimum value of the meridional overturning stream function marked as black dots in Fig. 1. The resulting mass fluxes are 191 Sv when using dry isentropic coordinates and 282 Sv when using moist isentropic coordinates, in close agreement with previous studies (Pauluis et al. 2008; Döös and Nilsson 2011). While moist isentropic coordinates takes both DSE and LH variations into account, the maximum mass flux is still less than the hydrothermal cycle maxima of 428 Sv. Based on these results we conclude that the meridional overturning component can at most account for 282 Sv (66%) of the total hydrothermal stream function suggesting that zonally oriented flow features account for at least the remaining 146 Sv (34%). The projection onto spatial coordinates (Fig. 3b) shows stationary zonal asymmetries mainly in the tropics with ascent over the west tropical Pacific and subsidence to the east, both associated with the Walker circulation. We therefore suggest that the Walker circulation is a dominant feature of the zonally asymmetric component of the hydrothermal cycle.

4. Annual and Inter-annual Variability

The hydrothermal stream function projects the general circulation of the atmosphere onto a plane in thermodynamical (LH–DSE) coordinates. The resulting hydrothermal cycle thus includes the Hadley and Walker circulations as well as midlatitude eddies, and presents an approximation of their combined mass fluxes across DSE surfaces. We study the tem-

poral variability of the hydrothermal stream function by calculating it for every month for the period 1979-2009 using ERA-Interim reanalysis data. We define the strength of the hydrothermal cycle as the average of the hydrothermal stream function maximum between 310 kJ kg⁻¹ and 325 kJ kg⁻¹ and denote this as $\Delta\psi$. The values 310 kJ kg⁻¹ and 325 kJ kg⁻¹ were chosen since the amplitude of the hydrothermal stream function is somewhat constant over those values and because the maximum mass flux occurs somewhere in that range.

While $\Delta\psi$ is a measure of the maximum mass flux in the hydrothermal stream function, we also present a measure of the range in LH over which mass fluxes occur. We define the width of the hydrothermal cycle as the difference between the maximum and minimum LH value within the -50 Sv stream line at DSE $s = 310$ kJ kg⁻¹. The width is then a measure of the difference in LH between the precipitating branch and the cooling branch, and we denote it as ΔLH . The precipitating branch approximately follows a moist adiabat, and thus ΔLH is in fact a measure of changes in near-surface MSE. Furthermore, as DSE and LH of near-surface air are connected by the Clausius-Clapeyron relationship (Fig. 2c), ΔLH can be seen as a measure of changes in near-surface DSE and thus near-surface temperature. For ascending motions over the ocean we can therefore expect a correlation between ΔLH and SST. We also note that ΔLH is fairly insensitive to the choice of DSE value at which it is evaluated.

We plot the resulting time series of $\Delta\psi$ as both seasonal (DJF, MAM, JJA, SON) and annual averages in Fig. 4. We find that the amplitude of the hydrothermal stream function undergoes large seasonal variations, but also that there is a clear inter-annual variability. By calculating a climatological annual cycle from the monthly data, we find that both $\Delta\psi$ and ΔLH vary semi-annually with peaks in the boreal winter and austral winter (Fig. 5). The strength of the meridional overturning stream function in dry isentropic coordinates previously calculated (Fig. 1b) has a semi-annual variation of similar amplitude (not shown) since

the meridional overturning circulation is stronger in the winter hemispheres. We therefore suggest that a large part of the seasonal variations of the hydrothermal stream function is due to the variations in the strength of the meridional overturning. We further speculate that the semi-annual variability in ΔLH is due to the fact that it is connected to the surface air temperature in regions of moist convection which are mostly in the tropics where surface air temperatures are known to have semi-annual peaks.

Increased sea-surface temperatures (SSTs) in the east equatorial Pacific that define positive ENSO phases are associated with a weakening of the zonal surface pressure gradient over the Pacific Ocean giving weaker zonal winds (Bjerknes 1969). We determine the monthly ENSO phase for the period 1979-2009 using the NINO3.4 index (Barnston et al. 1997) from the ERSST.v3b data set (Smith et al. 2008). Following Trenberth (1997), the NINO3.4 index is defined as the SST averaged over the region $5^{\circ}\text{S} - 5^{\circ}\text{N}$, $170^{\circ}\text{E} - 120^{\circ}\text{E}$. We then define seasonal positive or negative ENSO phases when the seasonally averaged NINO3.4 index is $\geq 0.4\text{K}$ or $\leq -0.4\text{K}$ respectively. In the same way, we denote a year as a positive or negative ENSO year when the annual mean NINO3.4 is $\geq 0.4\text{K}$ or $\leq -0.4\text{K}$.

We now correlate the seasonal and annual time series of $\Delta\psi$ and ΔLH with the NINO3.4 index. We de-trend the data by fitting a linear curve $y = at + b$, where y is either $\Delta\psi$ or ΔLH and t is time, and subtracting this curve from all time series. In ERA-Interim, $\Delta\psi$ has a linear trend of $0.46 \text{ Sv year}^{-1}$, while ΔLH has no significant trend. We note that a positive trend in the Walker circulation strength was also indicated by L’Heureux et al. (2013) and Tokinaga et al. (2012) using reanalysis of recent years. The annual de-trended time series are shown in Fig. 6. In the annual data we find correlations of $R = -0.42$ and $R = 0.63$ for $\Delta\psi$ and ΔLH respectively, both significant above the 95% level ($p = 0.02$ and $p < 0.01$ respectively). Note that $\Delta\psi$ has been multiplied by -1 in Fig. 6 to show the anti-correlation to NINO3.4 more clearly. From the correlation coefficients we find that a linear fit to the

NINO3.4 time series explains $R^2 = (-0.42)^2 = 18\%$ and $R^2 = 0.66^2 = 44\%$ of the variance
 in $\Delta\psi$ and ΔLH in Fig. 6. For the seasonal data (not shown), the correlations are somewhat
 lower, $R = -0.36$ and $R = 0.41$ for $\Delta\psi$ and ΔLH respectively. From Fig. 6, we see that
 the correlation with $\Delta\psi$ and ΔLH to NINO3.4 is higher during strong El Niño and La Niña
 events, and that it is lower during weaker events or neutral ENSO phases. If all neutral
 ENSO phases are removed from the data (not shown), the correlations for the annual time
 series improve to $R = -0.51$ and $R = 0.76$. Furthermore, we note that the correlations are
 especially low in the period 1991-1993 which could be due to changes in the radiative bal-
 ance and atmospheric circulation due to the eruption of Mount Pinatubo and the decrease
 in global-mean surface air temperature it caused (Bender et al. 2010). We also note a low
 correlation in for $\Delta\psi$ around 1982 at the time of the eruption of El Chichón. As volcanic
 eruptions have been shown to sometimes mask the effects of ENSO events (Angell 1988) it
 is conceivable that they can also affect the hydrothermal cycle, albeit differently depending
 on the magnitude and location.

The negative correlation for $\Delta\psi$ and positive correlation for ΔLH with NINO3.4 imply
 that the hydrothermal cycle weakens and widens during El Niño and strengthens and narrows
 during La Niña years. We show this by calculating composites of the hydrothermal stream
 functions during El Niño events and La Niña events in Fig. 7. The widening during El Niño
 implies that the precipitating branch of the cycle shifts towards a moist adiabat of higher
 MSE than during neutral ENSO phases. This is consistent with previous findings that El
 Niño events are associated with a overall increase in tropical surface air temperatures (Sea-
 ger et al. 2003), which results in increased tropical DSE and, by Clausius-Clapeyron scaling,
 increased LH. Because the precipitating branch approximately follows moist adiabats (MSE
 conserved) an increase in surface MSE results in an increase in MSE throughout the branch.
 Our analysis also shows that the hydrothermal cycle weakens during El Niño events (Fig. 6).
 This implies less mass flux across DSE surfaces, and we suggest that this is mainly due to

the weakened Walker circulation. Previous studies using spatial coordinates have found that El Niño events are also associated with stronger Hadley circulation (Oort and Yienger 1996; Lu et al. 2008). We also find correlations (not shown) indicating that this is true in our analysis. Since the hydrothermal cycle weakens, it suggests that the strengthening of the Hadley circulation is not enough to compensate for the weakening of the Walker circulation, leading to less mass fluxes across DSE surfaces globally, but distributed over a wider range of LH.

5. EC-Earth Historical Simulation

We now calculate the hydrothermal stream function from a historical simulation with the EC-Earth model and show the result in Fig. 8. Comparing to the results from ERA-Interim reanalysis data (Fig. 2c), we find that the precipitating branch in EC-Earth has lower DSE and LH (and thus MSE) than ERA-Interim. As MSE is approximately conserved in this branch and LH is connected to DSE by the Clausius-Clapeyron relation, this indicates that EC-Earth has lower surface air temperatures in moist convective regions, consistent with findings by Hazeleger et al. (2011). Less DSE and LH would imply less DSE and LH fluxes in the precipitating branch of the hydrothermal cycle, but EC-Earth compensates this by having more ascent of moist air and thus a stronger hydrothermal cycle. The time-averaged hydrothermal stream function in EC-Earth has an amplitude of -514 Sv. As the EC-Earth historical simulation uses observed or reconstructed radiative forcing and atmospheric composition, the incoming and outgoing radiation at the top of the atmosphere should agree between ERA-Interim and EC-Earth. The upward flux of MSE should therefore be similar, suggesting that a hydrothermal cycle with low MSE in the precipitating branch will have a high amplitude. This argument is also consistent with the thermodynamic reasoning by Held and Soden (2006), where higher surface air temperatures lead to less convective mass flux. Furthermore, we find that the meridional overturning in EC-Earth (not shown) is stronger

than in ERA-Interim and can account for 303 Sv (59%) of the amplitude of the hydrothermal stream function, suggesting that the zonal flows account for 211 Sv (41%).

The NINO3.4 index for 1979-2009 indicates that only 17 of the 31 years of ERA-Interim data can be associated with ENSO events. To test our statistic on a longer time series we repeat our analysis using the 155-year historical simulation with EC-Earth. Sterl et al. (2012) found that EC-Earth can reproduce the ENSO phenomenon well, although the ENSO phases are somewhat weaker and less frequent than observations. The period 1850-2005 results in 60 years that can be associated with an ENSO event. In the EC-Earth historical simulation, $\Delta\psi$ has a trend of -0.05 Sv year $^{-1}$ and ΔLH has a trend of 0.02 kJkg $^{-1}$ year $^{-1}$, both statistically significant ($p < 0.01$) and indicating a weakening and widening of the cycle consistent with the thermodynamic scaling arguments by Held and Soden (2006) and Vecchi et al. (2006).

We also calculate the standard deviations of de-trended annual-mean NINO3.4, $\Delta\psi$ and ΔLH using both ERA-Interim and EC-Earth. We find that standard deviations of all three metrics are lower in EC-Earth than in ERA-Interim, indicating that ENSO events are weaker in EC-Earth. Moreover, we identify 60 ENSO years of the total 155 years in the EC-Earth historical simulation, while there are 17 ENSO years in 31 years of ERA-Interim data. This indicates that ENSO events are also less frequent in EC-Earth than recent observations.

As done for the ERA-Interim data, we calculate a mean annual cycle for $\Delta\psi$ and ΔLH in EC-Earth. The results, shown in Fig. 9, are similar to the results for ERA-Interim with a semi-annual variation with both $\Delta\psi$ and ΔLH peaking in January and July. The amplitudes of the annual cycles of $\Delta\psi$ and ΔLH are slightly higher in EC-Earth than ERA-Interim, which is likely due to the meridional overturning (not shown) being slightly stronger, as mentioned earlier.

386

387 We show the de-trended, annual-mean NINO3.4 index together with $-\Delta\psi$ and ΔLH
388 from the EC-Earth data in Fig. 10. Note that, just as in Fig. 6, $\Delta\psi$ has been multiplied by
389 -1 . We find that $\Delta\psi$ and ΔLH are correlated with the NINO3.4 index by $R = -0.81$ and
390 $R = 0.45$ respectively, both significant at the 99% level (both $p < 0.01$). Hence, just as in the
391 30 years of ERA-Interim data, the hydrothermal cycle widens and weakens during El Niño
392 events and narrows and intensifies during La Niña events. Composites of the hydrothermal
393 stream function for El Niño and La Niña events in EC-Earth (not shown) are similar to those
394 shown for ERA-Interim in Fig. 7. Compared to the correlations obtained in ERA-Interim
395 the correlation is higher for $\Delta\psi$ but lower for ΔLH . Hazeleger et al. (2011) showed that the
396 SST variations in EC-Earth associated with ENSO events are more confined to the equa-
397 tor than in observations, which we speculate could partially explain the lower correlation
398 between ΔLH and NINO3.4. Both data sets include observed volcanic gases and aerosols
399 therefore include the effects of eruptions. This is particularly discernible in the ERA-Interim
400 time series (Fig. 6) where both $\Delta\psi$ and ΔLH decrease in the year 1992 when the NINO3.4
401 index increases. Such events have a larger impact on the correlations using ERA-Interim
402 data since that data set is much shorter than the EC-Earth simulation.

403

404 6. Discussion

405 We have shown (Figs. 2c, 8) that global mass fluxes form a single hydrothermal cycle
406 in latent heat (LH) versus dry static energy (DSE) coordinates. Results from ERA-Interim
407 data 1979-2009 show that the cycle has an amplitude of 428 Sv. This thermodynamic cycle
408 describes three processes: heating and moistening of near-surface air masses following the
409 Clausius–Clapeyron relationship, heating and drying of air masses along moist adiabats in
410 the tropics, and radiative cooling of air masses at the top of the atmosphere and at high

latitudes. A projection onto spatial coordinates (Fig. 3) shows that the hydrothermal stream function captures both the Walker circulation and the meridional overturning circulation. The hydrothermal stream function thus includes both the zonal and meridional overturning circulations due to their exchanges of LH and DSE and does not differentiate between them due to their orientation in spatial coordinates. Less than 70% of the total mass flux in the hydrothermal stream function can be explained by the meridional circulation in moist isentropic coordinates (Pauluis et al. 2008; Döös and Nilsson 2011). Because they play a central role in the global heat budget, we suggested that the unexplained 30% of the total mass fluxes is associated with zonally oriented flows such as the Walker circulation. From our analysis based on ERA-Interim data, this would lead to about 148 Sv attributable to the Walker circulation, which is in line with an estimate of more than 100 Sv from other studies (Trenberth et al. 2000).

The hydrothermal cycle gives a unified description of the heat fluxes within the Hadley and Walker circulations and within synoptic eddies in midlatitudes. By calculating it for both ERA-Interim reanalysis and historical simulations from the EC-Earth climate model, we find that the hydrothermal cycle is narrower and stronger in EC-Earth (Fig. 8). We attribute this to the colder global-mean surface air temperature in EC-Earth leading to less surface DSE and LH (and thus MSE) which is compensated by more ascending motion such that the upward MSE flux is similar to that in ERA-Interim. We speculate that examining several climate models would show that those with colder surface air temperatures than observations would have a stronger and narrower hydrothermal cycle than those with warmer temperatures. The dependence on the surface temperatures can be tested by calculating the hydrothermal stream function for different ENSO phases. We have shown in both ERA-Interim reanalysis and EC-Earth historical simulations that the hydrothermal cycle widens and weakens with increased SSTs and vice versa. On inter-annual time scales, variations in NINO3.4 index from the ERSST.v3b data set (Smith et al. 2008) explain 18% and 44% of

the variations in strength and width of the hydrothermal stream function respectively. From 155-year historical simulations with the EC-Earth climate model the numbers are 66% and 20%. This shows that the results obtained with ERA-Interim are verified in longer simulated data sets. It also gives some evidence that EC-Earth is able to simulate both ENSO and its effects on the atmospheric general circulation, although ENSO events are somewhat weaker and less frequent than in observations.

We have defined the hydrothermal stream function using globally integrated advection of DSE, $\vec{v} \cdot \nabla s$, and shown that it is related to advection of LH. The total mass flux across DSE surfaces is given by the globally integrated tendency $d_t s = \partial_t s + \vec{v} \cdot \nabla s$. Calculating a stream function, ψ^{tot} , based on this tendency yields results very similar to those obtained for the hydrothermal stream function, ψ . Repeating the analysis in this paper with ψ^{tot} instead of ψ , the correlations of $\Delta\psi$ and ΔLH to ENSO become slightly smaller in both ERA-Interim and EC-Earth, but are still robust and statistically significant. The composites (Fig. 7) do not change. Hence, the hydrothermal stream function is a good approximation to the total mass fluxes across DSE surfaces in the data.

A caveat of the analysis in this paper is the absence of resolved convection in ERA-Interim and EC-Earth, as moist convection is an important process for the conversion of LH to DSE. Following Simmons and Burridge (1981), the vertical velocity, ω , is calculated from the horizontal mass fluxes and changes in surface pressure. As the vertical mass fluxes must balance the divergence of the horizontal mass fluxes, which in turn are calculated partly using assimilated observations, we argue that the vertical mass flux amount in spatial coordinates is realistic. Convection occurs mostly on sub-grid scales so we expect both ERA-Interim and EC-Earth to underestimate the correlations between vertical velocity, ω , and DSE and LH. As a consequence, it is thus possible that some moist convection occurs at somewhat higher LH and DSE than the precipitating branch in Figs. 2c and 8. In a recent study Pauluis and

Mrowiec (2013) show convection occurring at θ_e values $\sim 5\text{K}$ higher than the mean θ_e profile. This could have the effect that in our analysis some ascent occurs at the same LH and DSE as nearby descent, which would make us underestimate the mass flux to some extent.

Future applications of the analysis presented here should include a study of the hydrothermal cycle response to anthropogenic forcing in climate models. Results from an intercomparison of climate models (Meehl et al. 2007) suggest that the pattern of global warming is similar to that of an El Niño event, in which case we would expect a widening and weakening of the hydrothermal cycle. The hydrothermal cycle would reveal global changes in mass fluxes as well as exchanges of MSE, DSE and LH, and thereby changes in precipitation and evaporation.

In concurrent work (Laliberté et al. 2013b), we have applied a similar methodology to study how irreversible entropy production associated with moist processes constrains the maximum work output of the atmospheric heat engine in a climate change scenario. The results indicate that a wider hydrothermal cycle size is associated with a reduction in maximum work output for the global atmospheric heat engine.

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APPENDIX A

Meridional overturning

We calculate meridional overturning stream functions from the ERA-Interim (Fig. 1) and EC-Earth (not shown) data sets, using pressure, p , dry static energy, s , and moist static energy, h , as vertical coordinates, following Döös and Nilsson (2011).

$$\begin{aligned} \psi(y, \chi) = & \frac{1}{t_1 - t_0} \int_{t_0}^{t_1} \oint_x \int_0^{p_s} \\ & \mu[\chi'(x, y, p, t) - \chi] \frac{v(x, y, p, t)}{g} dp dx dt, \end{aligned} \quad (\text{A1})$$

where χ stands for vertical coordinates p , s , or h . μ is a unit step function defined as

$$\begin{aligned} \mu[\chi'(x, y, p, t) - \chi] &= 1 \quad \text{if} \quad \chi'(x, y, p, t) - \chi \geq 0 \\ \mu[\chi'(x, y, p, t) - \chi] &= 0 \quad \text{if} \quad \chi'(x, y, p, t) - \chi < 0 \end{aligned}$$

As in the hydrothermal stream function, t_0 is 1 January 1979 and t_1 is 31 December 2009. We find that amplitudes and shapes of the stream functions were consistent with other studies (Pauluis et al. 2008; Döös and Nilsson 2011). We estimate the total mass flux by the meridional circulation in moist isentropic coordinates as $\max(\psi(y, h)) - \min(\psi(y, h))$.

APPENDIX B

Discretisation

We discretize eq. 5 as follows and illustrate the setup in Fig. 11. We denote the zonal, meridional and vertical mass fluxes as U, V, W [kg s^{-1}] respectively. The zonal mass flux

through the interface between grid box (i, j, k) and $(i + 1, j, k)$ is $U_{i,j,k}$. If $s_{i+1,j,k} > s_{i,j,k}$, then we find the smallest positive integer n such that $s_{i,j,k} + n\Delta s > s_{i+1,j,k} > s_{i,j,k}$. This indicates a positive mass flux through the DSE isosurfaces s and $s + \Delta s$ at latent heat $l_{i,j,k}$. We then record a mass flux of size $U_{i,j,k}$ at coordinates $(l, s), (l, s + \Delta s), \dots (l, s + n\Delta s)$. If $s_{i+1,j,k} < s_{i,j,k}$, then we find the smallest positive integer n such that $s_{i,j,k} - n\Delta s < s_{i+1,j,k} < s_{i,j,k}$. We then record a mass flux of size $-U_{i,j,k}$ at coordinates $(l, s), (l, s - \Delta s), \dots (l, s - n\Delta s)$. Contributions from meridional and vertical mass fluxes are added in a similar fashion. We then obtain the hydrothermal stream function by integrating the net mass fluxes F in hydrothermal coordinates.

$$\psi_{m,r} = \frac{1}{N} \sum_n \sum_m F_{m,r,n}, \quad (\text{B1})$$

where m, r are discrete coordinates in hydrothermal space such that $l_m = l_0 + m\Delta l$ and $s_r = s_0 + r\Delta s$. The index n is the time step and N the total number of time steps over which the stream function is integrated.

The local eulerian part of eq. 1, ψ^{loc} , is discretized in a similar way. However, instead of using mass fluxes from one grid box to an adjacent one, we use the mass of a grid box and the local rate of change in DSE from time step n to $n + 1$. Hence, if $s_{i,j,k,n+1} > s_{i,j,k,n}$, then we find the smallest positive integer a such that $s_{i,j,k} + a\Delta s > s_{i,j,k,n+1} > s_{i,j,k,n}$. The mass flux from one time step to another is $dM/\Delta t$, i.e. the mass of the grid box and the length of the time step. As done for the advective part above, we then record a mass flux from $s_{i,j,k,n}$ to $s_{i,j,k,n+1}$, and integrate the mass fluxes.

APPENDIX C

Clausius-Clapeyron

537

538 In Figs. 2a,b,c and 8, we show the saturated moist static energy (MSE) for surface air as
539 a dashed line. We calculated it as follows. For a given dry static energy (DSE), s , we obtain
540 the maximum temperature, T , by adiabatic descent to $z = 0$, in which case $T = s/c_p$. For
541 a given temperature, we find the saturation vapor pressure, e_s , by integrating the Clausius-
542 Clapeyron equation (Wallace and Hobbs 2006). e_s is related to the saturation mixing ratio,
543 q_s , by the approximation $q_s \approx 0.622e_s/1013\text{hPa}$ if we assume the ambient pressure is taken
544 at the surface pressure $p_s = 1013\text{ hPa}$. Thus, for saturated air close to the surface, LH is a
545 function of DSE. With the approximations of Wallace and Hobbs (2006), we thus obtain

$$l = L_v \frac{0.622}{1013\text{hPa}} e_0 \times e^{5420 \times \left(\frac{1}{273\text{K}} - \frac{c_p}{s} \right)}, \quad (\text{C1})$$

546 where $e_0 = e_s(273\text{K}) = 6.11\text{hPa}$. The surface saturated MSE is then, $h = l + s$, shown as a
547 dashed line in Fig. 2a,b,c.

REFERENCES

- 550 Angell, J. K., 1988: Impact of El Nino on the Delineation of Tropospheric Cooling Due to
551 Volcanic Eruptions. *J. Geophys. Res.*, **93 (D4)**, 3697–3704.
- 552 Ballarotta, M., 2013: The thermohaline circulation during the Last Glacial Maximum and
553 in the Present-Day climate. Ph.D. thesis, Stockholm University.
- 554 Barnston, A. G., M. Chelliah, and S. B. Goldenberg, 1997: Documentation of a highly
555 ENSO-related region in the equatorial Pacific. *Atmos.-Ocean*, **35 (3)**, 367–383.
- 556 Bender, F. A.-M., A. M. Ekman, and H. Rodhe, 2010: Response to the eruption of Mount
557 Pinatubo in relation to climate sensitivity in the CMIP3 models. *Clim. Dyn.*, **35 (5)**,
558 875–886.
- 559 Berrisford, P., D. Dee, K. Fielding, M. Fuentes, P. Kållberg, S. Kobayashi, and S. Uppala,
560 2009: The ERA-Interim archive. *ERA report series*, **1 (1)**, 1–16.
- 561 Berrisford, P., P. Kållberg, S. Kobayashi, D. Dee, S. Uppala, A. J. Simmons, P. Poli, and
562 H. Sato, 2011: Atmospheric conservation properties in era-interim. *Q. J. R. Meteorol.*
563 *Soc.*, **137**, 1381–1399.
- 564 Bjerknes, J., 1969: Atmospheric Teleconnections from the Equatorial Pacific. *Mon. Wea.*
565 *Rev.*, **97 (3)**, 163–172.
- 566 Döös, K. and J. Nilsson, 2011: Analysis of the Meridional Energy Transport by Atmospheric
567 Overturning Circulations. *J. Atmos. Sci.*, **68**, 1806–1820.
- 568 Döös, K., J. Nilsson, J. Nycander, L. Brodeau, and M. Ballarotta, 2012: The World Ocean
569 Thermohaline Circulation. *J. Phys. Oceanogr.*, **42**, 1445–1460.

- Emanuel, K. A., 1994: *Atmospheric Convection*. Oxford University Press.
- Groeskamp, S., J. D. Zika, T. J. McDougall, B. M. Sloyan, and F. B. Laliberté, 2013: Analysis of ocean circulation in thermohaline-coordinate: Introducing the stationary thermohaline streamfunction. *In Prep*.
- Hazeleger, W., et al., 2011: EC-Earth V2.2: description and validation of a new seamless earth system prediction model. *Climate Dynamics*, 1–19, URL <http://dx.doi.org/10.1007/s00382-011-1228-5>, 10.1007/s00382-011-1228-5.
- Held, I. M. and T. Schneider, 1999: The Surface Branch of the Zonally Averaged Mass Transport Circulation in the Troposphere. *J. Atmos. Sci.*, **56**, 1688–1697.
- Held, I. M. and B. J. Soden, 2006: Robust Responses of the Hydrological Cycle to Global Warming. *J. Clim.*, **19**, 5686–5699.
- Juckes, M., 2001: A generalization of the Transformed Eulerian-Mean meridional circulation. *Q. J. R. Meteorol. Soc.*, **127**, 147–160.
- Kållberg, P., P. Berrisford, B. Hoskins, A. J. Simmons, S. Uppala, S. Lamy-Thépaut, and R. Hine, 2005: ERA-40 atlas. *ERA-40 Project Report Series*, **19**, 191.
- Karoly, D. J., P. C. McIntosh, P. Berrisford, T. J. McDougall, and A. C. Hirst, 1997: Similarities of the Deacon Cell in the Southern Ocean and the Ferrel Cells in the atmosphere. *Q. J. R. Meteorol. Soc.*, **123**, 519–526.
- Kjellsson, J. and K. Döös, 2012: Lagrangian decomposition of the Hadley and Ferrel Cells. *Geophys. Res. Lett.*, **39** (L15807).
- Laliberté, F., T. Shaw, and O. Pauluis, 2012: Moist Recirculation and Water Vapor Transport on Dry Isentropes. *J. Atmos. Sci.*, **69**, 875–890.
- Laliberté, F. B., T. Shaw, and O. M. Pauluis, 2013a: A theory for the lower-tropospheric structure of the moist isentropic circulation. *J. Atmos. Sci.*, **70**, 843–854.

- 594 Laliberté, F. B., J. D. Zika, L. Mudryk, P. Kushner, J. Kjellsson, and K. Döös, 2013b: Re-
 595 duced maximum work output of the moist atmospheric heat engine in a changing climate.
 596 *Manuscript.*
- 597 L’Heureux, M. L., S. Lee, and B. Lyon, 2013: Recent multidecadal strengthening of
 598 the Walker circulation across the tropical Pacific. *Accepted for Nat. Climate Change*,
 599 <http://dx.doi.org/10.1038/nclimate1840>.
- 600 Lu, J., G. Chen, and D. M. W. Frierson, 2008: Response of the Zonal Mean Atmospheric
 601 Circulation to El Niño versus Global Warming. *J. Clim.*, **21**, 5835–5851.
- 602 McIntosh, P. C. and T. J. McDougall, 1996: Isopycnal Averaging and the Residual Mean
 603 Circulation. *J. Phys. Oceanogr.*, **26**, 1655–1660.
- 604 Meehl, G., et al., 2007: Global Climate Projections. *Climate Change 2007: The Physical*
 605 *Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the*
 606 *Intergovernmental Panel on Climate Change*, S. Solomon, D. Qin, M. Manning, Z. Chen,
 607 M. Marquis, K. Averyt, M. Tignor, and H. Miller, Eds., Cambridge University Press,
 608 Cambridge, United Kingdom and New York, NY, USA.
- 609 Meinshausen, M., et al., 2011: The RCP greenhouse gas concentrations and their extensions
 610 from 1765 to 2300. *Clim. Chang.*, **109**, 213–241.
- 611 Muller, C. J. and P. O’Gorman, 2011: An energetic perspective on the regional response of
 612 precipitation to climate change. *Nat. Climate Change*, **1**, 266–271.
- 613 Oort, A. H. and J. J. Yienger, 1996: Observed Interannual Variability in the Hadley Circu-
 614 lation and Its Connection to ENSO. *J. Clim.*, **9**, 2751–2767.
- 615 Pauluis, O., A. Czaja, and R. Korty, 2008: The Global Atmospheric Circulation on Moist
 616 Isentropes. *Science*, **321**, 1075–1078.

- Pauluis, O., A. Czaja, and R. Korty, 2010: The Global Atmospheric Circulation in Moist Isentropic Coordinates. *J. Clim.*, **23**, 3077–3093.
- Pauluis, O. M. and A. A. Mrowiec, 2013: Isentropic analysis of convective motions. *Submitted to J. Atmos. Sci.*
- Peixoto, J. P. and A. H. Oort, 1992: *Physics of Climate*. Springer-Verlag.
- Seager, R., N. Harnik, Y. Kushnir, W. Robinson, and J. Miller, 2003: Mechanisms of Hemispherically Symmetric Climate Variability. *J. Clim.*, **16**, 2960–2978.
- Simmons, A. J. and D. M. Burridge, 1981: An Energy and Angular-Momentum Conserving Vertical Finite-Difference Scheme and Hybrid Vertical Coordinate. *Mon. Wea. Rev.*, **109**, 758–766.
- Smith, T. M., R. W. Reynolds, T. C. Peterson, and J. Lawrimore, 2008: Improvements to noaa’s historical merged land–ocean surface temperature analysis (1880–2006). *J. Clim.*, **21**, 2283–2296.
- Sterl, A., et al., 2012: A look at the ocean in the EC-Earth climate model. *Clim. Dyn.*, **39**, 2631–2657.
- Tokinaga, H., S.-P. Xie, C. Deser, Y. Kosaka, and Y. M. Okumura, 2012: Slowdown of the Walker circulation driven by tropical Indo-Pacific warming. *Nature*, **491**, 439–444.
- Townsend, R. D. and D. R. Johnson, 1985: A Diagnostic Study of the Isentropic Zonally Averaged Mass Circulation during the First GARP Global Experiment. *J. Atmos. Sci.*, **42 (15)**, 1565–1579.
- Trenberth, K. E., 1997: The Definition of El Niño. *Bull. Amer. Meteor. Soc.*, **78**, 2771–2777.
- Trenberth, K. E. and D. P. Stepaniak, 2003a: Covariability of Components of Poleward Atmospheric Energy Transports on Seasonal and Interannual Timescales. *J. Clim.*, **16**, 3691–3705.

- 641 Trenberth, K. E. and D. P. Stepaniak, 2003b: Seamless Poleward Atmospheric Energy
642 Transports and Implications for the Hadley Circulation. *J. Clim.*, **16**, 3706–3722.
- 643 Trenberth, K. E., D. P. Stepaniak, and J. M. Caron, 2000: The Global Monsoon as Seen
644 through the Divergent Atmospheric Circulation. *J. Clim.*, **13**, 3969–3993.
- 645 Vecchi, G. A., B. J. Soden, A. T. Wittenberg, I. M. Held, A. Leetmaa, and M. J. Harrison,
646 2006: Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing.
647 *Nature*, **441** (4), 73–76.
- 648 Wallace, J. M. and P. V. Hobbs, 2006: *Atmospheric science: An introductory survey*. 2d ed.,
649 International Geophysics Series, v. 92, Elsevier.
- 650 Zika, J. D., M. H. England, and W. P. Sijp, 2012: The Ocean Circulation in Thermohaline
651 Coordinates. *J. Phys. Oceanogr.*, **42**, 708–724.
- 652 Zika, J. D., W. P. Sijp, and M. H. England, 2013: Vertical heat transport by ocean circulation
653 and the role of mechanical and haline forcing. *J. Phys. Oceanogr.*, In Press.

List of Figures

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33

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690	6	a) Annual-mean strength ($\Delta\psi$, blue) of the hydrothermal stream function	
691		and the NINO3.4 anomaly (red) for 1979-2009. $\Delta\psi$ has been multiplied by	
692		-1 as it is anti-correlated to NINO3.4. b) Annual-mean width (ΔLH , blue)	
693		of the -50Sv stream line at $s = 310 \text{ kJkg}^{-1}$ in the hydrothermal stream	
694		function and the NINO3.4 anomaly as in the top plot. All time series are de-	
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696		of explained variance. In a) R is the correlation between $-\Delta\psi$ and NINO3.4.	37
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11 Schematic illustration of the discretisation of the hydrothermal stream func-
tion, $\psi(l, s)$ (eqs. 5 and B1). a) represents the target grid in LH-DSE coordi-
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EC-Earth data. The zonal mass flux $U_{i,j}$ (blue arrow) crosses two isolines of
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FIG. 1. Zonal-mean meridional overturning stream functions in isobaric (a), dry isentropic (b), and moist isentropic (c) coordinates in ERA-Interim reanalysis 1979-2009. Positive values represent clockwise circulation and negative values represent anti-clockwise circulation. Units in $\text{Sv} = 10^9 \text{ kg s}^{-1}$. Zonal-mean surface values of pressure (a), DSE (b), and MSE (c) are shown as black dashed lines. The maximum and minimum of the stream functions are shown as black dots.

FIG. 2. a) Global mass fluxes along DSE gradients in ERA-Interim reanalysis 1979-2009 binned onto l, s space. Units are $\text{Sv} (\text{kJ kg}^{-1})^{-1}$, where $\text{Sv} = 10^9 \text{ kg s}^{-1}$. Horizontal axis is latent heat (LH) and vertical axis is dry static energy (DSE), both in units kJ kg^{-1} . b) The part of the stream function that is only based on local tendencies, ψ^{loc} . c) Global hydrothermal stream function, $\psi(l, s)$ in Sv calculated using the mass fluxes in a). d) The hydrothermal stream function maximum is calculated as $\max(\psi) - \min(\psi)$ along the horizontal LH axis for each DSE value. The peak amplitude is $\psi = -428 \text{ Sv}$. As in Fig. 1, negative values represent anti-clockwise circulation. Superimposed is an isoline of 345 kJ kg^{-1} moist static energy (MSE, dashed). Also shown is the curve representing LH and DSE values for saturated surface air (double dashed), as given by the Clausius-Clapeyron relationship. See eq. (C1) for the derivation of the curve.

FIG. 3. The hydrothermal stream function between -100 and -400 Sv subdivided into sections (inlay in Fig. a), and projected onto spatial coordinates. Air masses outside the stream layer are in grey shading. a) The zonal mean depicts a meridional overturning circulation where air ascends in the tropics (red, yellow), moves poleward while slowly subsiding (blue), and then returns to the tropical boundary layer (purple). b) The 570 hPa level shows ascending motion (yellow, red) above the warm pool in the western Pacific and subsiding (green, blue) air in the eastern Pacific and Atlantic as part of the Walker circulation.

FIG. 4. The amplitude of the hydrothermal stream function between $s = 310 \text{ kJkg}^{-1}$ and $s = 325 \text{ kJkg}^{-1}$, $\Delta\psi$, for 1979-2009. Black line shows seasonal averages (DJF, MAM, JJA, SON) and red line shows annual averages.

FIG. 5. The average annual cycle of the strength ($\Delta\psi$, blue) and width (ΔLH , red) of the hydrothermal stream function in ERA-Interim calculated from de-trended data.

FIG. 6. a) Annual-mean strength ($\Delta\psi$, blue) of the hydrothermal stream function and the NINO3.4 anomaly (red) for 1979-2009. $\Delta\psi$ has been multiplied by -1 as it is anti-correlated to NINO3.4. b) Annual-mean width (ΔLH , blue) of the -50Sv stream line at $s = 310 \text{ kJkg}^{-1}$ in the hydrothermal stream function and the NINO3.4 anomaly as in the top plot. All time series are de-trended. Numbers (R) are correlation coefficients such that R^2 is the fraction of explained variance. In a) R is the correlation between $-\Delta\psi$ and NINO3.4.

FIG. 7. Composites of the hydrothermal stream function in ERA-Interim for months of El Niño (red) and La Niña (blue).

FIG. 8. Hydrothermal stream function as in Fig. 2a but for EC-Earth historical run 1850-2005.

FIG. 9. As Fig. 5 but for EC-Earth historical simulation 1850–2005.

FIG. 10. As in Fig. 6 but for EC-Earth 1850-2005.

FIG. 11. Schematic illustration of the discretisation of the hydrothermal stream function, $\psi(l, s)$ (eqs. 5 and B1). a) represents the target grid in LH-DSE coordinates and b) represents one level on the spatial grid of the ERA-Interim and EC-Earth data. The zonal mass flux $U_{i,j}$ (blue arrow) crosses two isolines of DSE in b) and is therefore represented in that range in LH-DSE grid (a).