Consistency and Homogeneity of Atmospheric Energy, Moisture, and Mass Budgets in ERA5

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(Manuscript received 2 September 2020, in final form 8 February 2021)

ABSTRACT: This study uses advanced numerical and diagnostic methods to evaluate the atmospheric energy budget with the fifth major global reanalysis produced by ECMWF (ERA5) in combination with observed and reconstructed top of the atmosphere (TOA) energy fluxes for the period 1985–2018. We assess the meridional as well as ocean–land energy transport and perform internal consistency checks using mass-balanced data. Furthermore, the moisture and mass budgets in ERA5 are examined and compared with previous budget evaluations using ERA-Interim as well as observation-based estimates. Results show that peak annual mean meridional atmospheric energy transports in ERA5 (4.58 ± 0.07 PW in the Northern Hemisphere) are weaker compared to ERA-Interim (4.74 ± 0.09 PW), where the higher spatial and temporal resolution of ERA5 can be excluded as a possible reason. The ocean–land energy transport in ERA5 is reliable at least from 2000 onward (~ 2.5 PW) such that the imbalance between net TOA fluxes and lateral energy fluxes over land are on the order of $\sim 1 \text{ W m}^{-2}$. Spinup and spindown effects as revealed from inconsistencies between analyses and forecasts are generally smaller and temporally less variable in ERA5 compared to ERA-Interim. Evaluation of the moisture budget shows that the ocean–land moisture transport and parameterized freshwater fluxes agree well in ERA5, while there are large inconsistencies in ERA-Interim. Overall, the quality of the budgets derived from ERA5 is demonstrably better than estimates from ERA-Interim. Still some particularly sensitive budget quantities (e.g., precipitation, evaporation, and ocean–land energy transport) show apparent inhomogeneities, especially in the late 1990s, which warrant further investigation and need to be considered in studies of interannual variability and trends.

KEYWORDS: Energy transport; Energy budget/balance; Heat budgets/fluxes; Moisture/moisture budget; Reanalysis data; Climate variability

1. Introduction

Earth's atmosphere exchanges vast amounts of energy with space, the ocean, and land surfaces. Large-scale lateral energy transports balance the arising differential heating, which exhibits a strong meridional gradient (Peixoto and Oort 1992). Moreover, atmospheric energy transports and their divergence are closely linked to spatial patterns of surface energy flux (Liu et al. 2017; Mayer et al. 2017; Trenberth and Fasullo 2017). Thus, atmospheric transports also determine patterns of net heat uptake by the oceans, where most of the excess heat arising from anthropogenic climate change is stored (von Schuckmann et al. 2020). The atmospheric moisture cycle represents the link between the energy budget and the mass budget. Evaporation transfers water vapor and thus mass and latent heat into the atmosphere, which subsequently transports moisture to regions where precipitation exceeds evaporation, where the latent heat is released and warms the atmosphere. Changes to the atmospheric moisture cycle are of direct relevance to society, and climate models predict an enhancement of climatologically dry and wet zones (Held and Soden 2006; Seager et al. 2010; Collins et al. 2013). These trends have been detected

indirectly through changes in the oceans (e.g., Boyer et al. 2005; Rhein et al. 2013; Li et al. 2020; Cheng et al. 2020), but direct detection of these changes with observational and reanalysis datasets is challenging due to existing biases (Trenberth et al. 2011) and lack of temporal stability (e.g., Robertson et al. 2016).

Atmospheric reanalyses provide a four-dimensional gridded estimate of the atmospheric state and thus in principle are perfectly suited for the quantification of atmospheric budgets and changes thereof. However, the assimilating models are imperfect, which can introduce biases where observational information is lacking. In addition, the observing system is constantly evolving, which often introduces spurious jumps in reanalyzed quantities that are relevant for budget evaluations (Chiodo and Haimberger 2010; Berrisford et al. 2011; Hersbach et al. 2020). Nevertheless, reanalysis is an iterative process, and every new generation of these products is expected to come with qualitative improvements over its predecessors (Dee et al. 2014; Buizza et al. 2018).

Here we use data from the fifth major global reanalysis produced by ECMWF (ERA5; Hersbach et al. (2020); publicly available via the Copernicus Climate Change Service climate.copernicus.eu; see <https://confluence.ecmwf.int/display/CKB/How+to+download+ERA5> for instructions) to evaluate various aspects of the atmospheric energy, moisture, and mass budgets. Results are validated with observational products wherever

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DOI: 10.1175/JCLI-D-20-0676.1

possible and extensive comparison is made with ERA5's predecessor ERA-Interim (Dee et al. 2011) to document progress. Improvements are demonstrated in many climate-related aspects but also in terms of evaluation methods, such as mass adjustment (Trenberth 1991; Mayer and Haimberger 2012) and reduction of spectral noise over topography.

This paper is structured as follows. The input data are introduced in section 2. Section 3 describes numerical and diagnostic methods applied in this study. Section 4 presents results from the evaluation of the atmospheric mass and moisture budget. The evaluation of the atmospheric energy budget using ERA5 and a comparison with ERA-Interim and previous studies is presented in section 5. Summary and an outlook follow in section 6.

2. Data

The atmospheric mass, moisture, and energy budgets are calculated for the period 1985–2018 using ECMWF's latest atmospheric reanalysis dataset ERA5, which provides global data on a reduced Gaussian grid N320 (equivalent to 0.28° spatial resolution) with 1-hourly temporal resolution and in 137 atmospheric levels up to a pressure of 0.01 hPa. ERA5 is currently available from 1979 onward and consists of analyses and short-range forecasts initialized from the analyzed fields daily at 0600 and 1800 UTC. It also provides ten low-resolution ensemble members with 3-hourly temporal and N160 (equivalent to 0.56°) spatial resolution for uncertainty estimates. All budget quantities are aggregated to monthly means, which can be compared, e.g., to ERA-Interim data as presented by Mayer et al. (2017), that were calculated with 6-hourly time resolution on a reduced N128 Gaussian grid covering the full study period 1985–2018.

Net TOA fluxes are taken from the recently published DEEP-C v4.0 dataset (Liu et al. (2020); freely available on <https://researchdata.reading.ac.uk/271/>) providing monthly averages on a full Gaussian grid F128. The DEEP-C dataset is a backward extension of the Clouds and the Earth's Radiant Energy System–Energy Balanced and Filled (CERES-EBAF) product in version 4.1 (Loeb et al. 2009, 2018) using satellite data, AMIP5 model simulations, and ERA5 TOA flux data for reconstructions prior to the CERES era. Its availability from January 1985 to June 2019 determined the analysis period for this study.

Observation-based precipitation estimates are taken from the Global Precipitation Climatology Centre Full Data Monthly V2018 (GPCC; Becker et al. 2013; Schneider et al. 2016) and Global Precipitation Climatology Project v2.3 monthly analysis (GPCP; Adler et al. 2018) products providing monthly averages on a 1° and 2.5° regular grid, respectively. GPCC data are available from 1891 to 2016 using quality-controlled precipitation data from various sources (national weather and hydrological services, and historical data) and the robust SPHEREMAP interpolation method (Willmott et al. 1985) to provide near-global land coverage (with the exception of Antarctica). Nonetheless, sampling errors are still an issue in GPCC varying in space depending on the rain gauge network density (up to 40% in data-sparse regions according to Schneider et al. 2014). GPCP combines GPCC rain gauge data with satellite measurements and provides global data for the full study period.

In this study, ocean averages are referred to the global ocean area including regions covered by sea ice. Inland waters such as

the Great Lakes and the Caspian Sea are excluded. Land averages refer to all areas that are excluded in the ocean average, unless otherwise specified.

3. Methodology

a. Moisture and mass budget

The conservation of mass and moisture is a fundamental property of planetary physical systems such as Earth's atmosphere. The degree to which this property is satisfied represents an important method to assess the internal consistency and quality of reanalyses. We formulate the vertically integrated moisture budget as

$$\nabla \cdot \frac{1}{g} \int_0^{p_s} (\mathbf{v}q) dp = -\frac{1}{g} \frac{\partial}{\partial t} \int_0^{p_s} q dp - E - P, \quad (1)$$

where g is the gravitational constant, p_s is the surface pressure, q is the specific humidity, $\mathbf{v}$ is the horizontal wind vector, and E and P are the evaporation and precipitation, respectively. Evaporation E and precipitation P are flux densities in units of kilograms per square meter per second ($\text{kg m}^{-2} \text{s}^{-1}$), which are positive if directed downward. This equation describes the balance between the vertically integrated horizontal moisture flux divergence (left side; denoted as VIWVD hereafter) and the vertically integrated moisture (i.e., total column vapor) tendency (denoted as QTEND) plus surface freshwater fluxes (i.e., $P + E$). Vertical moisture fluxes at the TOA are zero. To quantify the moisture budget in ERA5 and ERA-Interim, we prefer analyzed quantities as they are constrained by observations as much as possible. Hence, we use analyzed hourly VIWVD (denoted as VIWVD_{AN}) and QTEND fields. Evaporation and precipitation are only available from short-term forecasts and are taken from twice-daily 12-hourly forecasts started at 0600 and 1800 UTC, respectively. In section 4b we will also take advantage of the availability of twice-daily instantaneous forecast fields of wind and moisture, which allows computation of forecasted moisture flux divergence fields (denoted as VIWVD_{FC}). We note that the ERA5 archive contains also accumulated moisture flux divergence fields, but they are numerically inaccurate (Paul Berrisford, ECMWF, 2020, personal communication) and hence are not used here.

Furthermore, we write the vertical integral of the atmospheric total mass budget as

$$\nabla \cdot \frac{1}{g} \int_0^{p_s} \mathbf{v} dp = -\frac{1}{g} \frac{\partial}{\partial t} p_s - E - P, \quad (2)$$

where the expression on the left represents the vertical integral of the total mass flux divergence including dry and moist air masses (denoted as MDIV_{total}), and the right side consists of the mass tendency (denoted as MTEND), evaporation and precipitation. Analogously to the moisture budget, P and E are taken from forecasts, and MTEND and MDIV_{total} are computed using analyzed fields.

b. Total energy budget

Mayer et al. (2017) proposed a formulation of the atmospheric total energy budget that treats enthalpy contained in

horizontal *and* vertical vapor fluxes in a consistent manner. Here, we use the simplified version of their equation, where horizontal *and* vertical enthalpy fluxes associated with water (and snow) are neglected. The introduced inaccuracies (discussed extensively in Mayer et al. 2017) are negligible in the context of this study. Thus, we formulate the vertically integrated divergence of the dry static plus kinetic energy fluxes (denoted as TEDIV hereafter) as follows:

$$\begin{aligned} \text{TEDIV} &= \nabla \cdot \frac{1}{g} \int_0^{p_s} [(1-q)c_a T_a + L_v(T_a)q + \Phi + k] \mathbf{v} dp \\ &= R_{\text{TOA}} - F_S - \text{AET}, \end{aligned} \quad (3)$$

where c_a is the specific heat capacity of dry air, T_a is the temperature of air measured in kelvin, L_v is the latent heat of vaporization, Φ the geopotential, and k is the kinetic energy. It is important to note that results are truly independent of reference temperature only in the steady state (which was assumed in Mayer et al. 2017), but there are ambiguities arising from local mass variations in the nonsteady state. These ambiguities become larger with larger local mass variations, which generally is the case for shorter averaging periods (see, e.g., Liang et al. 2018). However, in the present manuscript the minimum-averaging period is 12-months (the length of the temporal averaging window for the time series). At this time scale, local differences between results using the kelvin or Celsius scale are on the order of 1.0 W m^{-2} , and the RMS difference of ocean–land transport is 0.1 W m^{-2} . Due to this acceptably small ambiguity, we opted for the use of the kelvin scale here to ensure consistency with the other datasets used for comparison.

TEDIV can be calculated directly using analyzed state quantities, in this case referred to as TEDIV_{dir}. This is the method of choice if the total energy budget equation is used to infer an estimate of net surface energy fluxes F_S (sum of turbulent plus radiative heat fluxes) by combining satellite-based net TOA energy fluxes R_{TOA} (here from DEEP-C), vertically integrated atmospheric energy tendency AET, and directly computed TEDIV_{dir} (mass-adjusted as described below; Liu et al. 2015; Trenberth and Fasullo 2017; Mayer et al. 2017). For this application, AET is calculated from *analyzed* state quantities (denoted as AET_{AN}), since these are supposed to give the most accurate estimates of AET.

TEDIV can also be estimated indirectly (denoted as TEDIV_{ind}) from the parameterized state quantities of the right side of Eq. (3) containing the net TOA flux R_{TOA} , the net surface flux F_S , and the vertical integral of the *forecast* total atmospheric energy tendency AET_{FC}. The use of AET_{FC} in the evaluation of TEDIV_{ind} is consistent with forecasts of vertical fluxes (Mayer and Haimberger, 2012). All quantities used for computation of TEDIV_{ind} represent monthly averages of 12-hourly short-term forecasts.

c. Mass adjustment

The calculation of vertically integrated flux divergences requires considerable care. Before individual budget terms are computed, each input field is transformed to a quadratic full Gaussian grid in order to reduce aliasing effects (Durran 2013) at high latitudes caused by the spectral transform method that

is applied. Still this does not imply that the right side of Eq. (2) is equal to the left side. Consistency of the mass budget is achieved by iteratively adjusting the horizontal wind field $\mathbf{v}$ according to the mass budget residual where parameterized precipitation and evaporation fluxes are approximated by the analyzed divergence and tendency of the vertically integrated water vapor content [i.e., Eq. (1) is substituted into Eq. (2)] in order to get the expression $\text{MDIV}_{\text{total}} + \text{MTEND} = \text{VIWVD}_{\text{AN}} + \text{QTEND}$ (Trenberth 1991; Fasullo and Trenberth 2008; Mayer and Haimberger 2012). That is, we put all terms on the left side such that

$$\begin{aligned} &\nabla \cdot \frac{1}{g} \int_0^{p_s} \mathbf{v} dp - \nabla \cdot \frac{1}{g} \int_0^{p_s} (\mathbf{v}q) dp + \frac{1}{g} \frac{\partial}{\partial t} p_s - \frac{1}{g} \frac{\partial}{\partial t} \int_0^{p_s} q dp \\ &= \nabla \cdot \frac{1}{g} \int_0^{p_s} [(1-q)\mathbf{v}] dp + \frac{1}{g} \frac{\partial}{\partial t} \int_0^{p_s} (1-q) dp = \text{RE}_{\text{Mass}}, \end{aligned} \quad (4)$$

where RE_{Mass} is an estimate of the error in the divergence of atmospheric dry mass flux. Thus, taking the gradient of the inverted Laplacian of RE_{Mass} yields a two-dimensional field of the associated vertically integrated horizontal erroneous mass flux. The erroneous mass flux is converted to a vertically averaged spurious wind field (through division by local atmospheric mass), which is used to barotropically adjust the horizontal wind vector $\mathbf{v}$. Since this adjustment also affects VIWVD_{AN} on the left side of the equation (which itself is affected by spurious divergent winds, albeit only very weakly), one needs to make 2–3 iterations until RE_{Mass} vanishes and mass consistency is achieved. Please note the magnitude of this correction varies in space and time, as individual terms of Eq. (4) vary. The TEDIV_{dir} and VIWVD_{AN} fields are then computed using the mass-consistent wind which greatly reduces artificial noise over high topography. Also note that the difference between mass-adjusted and unadjusted VIWVD_{AN} is negligible (see, e.g., Berrisford et al. 2011), because moisture flux divergence is mainly driven by advection rather than wind divergence. This approach has already been introduced by Trenberth (1991) and Mayer and Haimberger (2012) in a slightly modified way.

Trenberth and Fasullo (2018) proposed a different approach to the mass adjustment, with a barotropic correction in a first step and an additional three-dimensional correction in a second step using the divergent component of moisture fluxes in each model level. That is, the 3D correction step weights the adjustment according to the distribution of the atmospheric moisture. However, it is unclear whether this method further reduces artificial noise over high topography. This will be tested in future studies.

All operations in spectral space as well as spectral transforms use routines used by the Integrated Forecasting System (IFS), which are provided through an openIFS license. This ensures maximum accuracy of our computations.

d. Diagnostics of budget consistency

The degree of closure of the diagnosed energy budget is an important indicator of the credibility of the results and their subsequent applications. Previous budget evaluations have shown substantial inconsistencies, but improvements are made

with every new iteration of reanalyses (Dee et al. 2014; Buizza et al. 2018). Here, we explore inconsistencies in the energy budget based on the degree of closure and satisfaction of physical constraints. To study budget closure, we combine parameterized R_{TOA} and F_S with $\text{TEDIV}_{\text{dir}}$ derived from analyses. Consequently, either forecasted or analyzed tendencies can be used to check the degree of closure. Depending on what is used, different aspects of the budget consistency are emphasized as outlined in the following.

First, we define the *residual* to be the remainder of Eq. (3) using analyzed tendencies AET_{AN} such that

$$\text{RE}_{\text{AN}} = R_{\text{TOA}} - F_S - \text{AET}_{\text{AN}} - \text{TEDIV}_{\text{dir}}. \quad (5)$$

Assuming that AET_{AN} is small in the long-term averages, this diagnostic measures the balance between the divergence of horizontal and vertical energy fluxes. It thus emphasizes budget inconsistencies arising from spinup and spindown effects in the parameterized fluxes.

Alternatively, forecasted tendencies AET_{FC} can be used to assess *self-consistency* of the budget, such that

$$\begin{aligned} \text{RE}_{\text{FC}} &= R_{\text{TOA}} - F_S - \text{AET}_{\text{FC}} - \text{TEDIV}_{\text{dir}} \\ &= \text{TEDIV}_{\text{ind}} - \text{TEDIV}_{\text{dir}}. \end{aligned} \quad (6)$$

This equation essentially compares directly computed TEDIV using analysis fields with the indirectly estimated TEDIV using forecasts (Mayer and Haimberger 2012). In this diagnostic, inconsistencies from spinup or spindown effects are largely cancelled out by the forecasted tendency. Instead, inconsistencies arising from sampling ($\text{TEDIV}_{\text{dir}}$ uses hourly instantaneous analysis fields, while $\text{TEDIV}_{\text{ind}}$ uses accumulated forecast fields), the mean drift of the atmospheric circulation in short-term forecasts, and potential inconsistencies in the diagnostic equations are emphasized. In other words, this diagnostic shows differences between energy transports obtained from the analyzed atmospheric state and those obtained from the model physics and dynamics driving the forecasts.

We investigate the closure of the mass and moisture budget analogously, but only use analyzed tendencies (i.e., equivalent to the residual). Another approach for the quality assessment of the results is the satisfaction of physical constraints; e.g., the inferred surface energy fluxes over land should be close to zero as the rate of energy storage in landmasses is small (on the order of $\leq 0.1 \text{ W m}^{-2}$ as estimated by von Schuckmann et al. 2020).

4. Atmospheric mass budget

a. Total mass and moisture budget

In this section, we investigate the atmospheric moisture and mass budget in ERA5. 1985–2018 averages of individual moisture budget terms are shown in Fig. 1 including $P + E$ in Fig. 1a, VIWVD_{AN} in Fig. 1b, and the moisture budget residual [difference between left and right side of Eq. (2)] in Fig. 1c. The analyzed moisture tendency is small compared to the moisture fluxes and is thus not shown here. Please note that the scaling in Fig. 1c differs from that in Figs. 1a or 1b. VIWVD_{AN} and $P + E$

FIG. 1. Global maps of individual moisture budget terms (mm day^{-1}) averaged over 1985–2018. (a) Precipitation plus evaporation ($P + E$), (b) vertically integrated moisture flux divergence (VIWVD_{AN}), and (c) the residual of the moisture budget ($\text{RE}_{\text{AN},\text{moist}}$). Note the different scaling. All fields are spectrally truncated at wavenumber 179.

are in good agreement such that the residual is relatively small across the globe, indicating small spinup and spindown effects in the forecasts. The largest discrepancies can be seen over high topography, e.g., along the Andes, Himalaya, or Rocky Mountains, where spectral noise is present in the VIWVD_{AN} field. Over the tropical ocean, the residual resembles the VIWVD_{AN} pattern, i.e., VIWVD_{AN} is too strong (root-mean-square RMS = 3.07 mm day^{-1} for 30°N – 30°S over the ocean) to balance $P + E$ (RMS = 3.03 mm day^{-1}) properly, which suggests that the tropical hydrological cycle spins down during the forecasts. Please note that the moisture budget residual has an RMS = 0.23 mm day^{-1} for 30°N – 30°S over the ocean, which is 92% smaller than that of $P + E$ in the same region.

The total mass flux divergence $\text{MDIV}_{\text{total}}$ and the residual of the total mass budget are presented in Fig. 2. $\text{MDIV}_{\text{total}}$ in the left panel is relatively homogeneous over the ocean, especially

FIG. 2. Global maps of (a) the total mass flux divergence ($\text{MDIV}_{\text{total}}$) and (b) total mass budget residual ($\text{RE}_{\text{AN},\text{total}}$) averaged over 1985–2018 (mm day^{-1}). Both fields are spectrally truncated at wavenumber 179.

in the Pacific Ocean basin where no clear $P + E$ signal is visible as we would expect it for the long-term mean total (moist plus dry) mass flux divergence. However, a signature of $P + E$ can be seen in the zonal average suggesting that its spatial pattern is buried by the spurious patterns of the dry air divergence (note that this diagnostic is based on the unadjusted wind fields). Wavelike structures parallel to coastal lines stem from the spectral method that is used, and the large-scale pattern of artificial noise over high topography may stem from steep gradients of surface pressure, although the field is spectrally truncated at wavenumber 179. The total mass budget residual in Fig. 2b is the difference between $\text{MDIV}_{\text{total}}$ (plus mass tendency, which is not shown here) and $P + E$ (Fig. 1a). It exhibits the artificial noise from the $\text{MDIV}_{\text{total}}$ field, and a clear $P + E$ signal in the equatorial Pacific Ocean as well as in the zonal mean, suggesting that moisture flux divergence and dry mass divergence are inconsistent. Global means and RMS values of the mass and moisture budget terms are summarized in the appendix B.

b. Ocean–land moisture transport

Figure 3 presents land (top panel) and ocean averages (bottom panels) of individual moisture budget terms. Figures 3a and 3c focus on P and E as obtained from reanalyses and observational products (see also Hersbach et al. 2018). Figures 3b and 3d assess the consistency between $P + E$ from forecasts and VIWVD_{AN} . The degree of agreement between these two terms is a measure of spinup and spindown effects (emphasized by the residual RE_{AN}) and their temporal variations.

Over land (Fig. 3a; please note that Antarctica is excluded in the following land averages as there are no GPCC observations in that region), precipitation from ERA5 has a 1985–2018 mean of 2.44 mm day^{-1} , very similar to that from ERA-Interim, which is 2.41 mm day^{-1} . Precipitation from ERA5 exhibits a weak but abrupt change around 2000, as seen in various energy budget terms (as will be discussed below), but is temporally stable around 2.41 mm day^{-1} afterward. Precipitation from ERA-Interim is stable around its climatological mean value so that it is smaller than P from ERA5 before 2002 and slightly larger afterward. We also show observation-based precipitation estimates from GPCP and GPCC. The former has a 1985–2018 land average of

2.31 mm day^{-1} , which is significantly less precipitation than from any other source. Precipitation from GPCC has a 1985–2016 mean of 2.43 mm day^{-1} and agrees well with those from reanalysis-based precipitation estimates. After 1997, both reanalysis products lie in between the values from the observation-based products. During this period, the temporal variability of the four precipitation products shown is very similar. Prior to 1997, we find that P from ERA-Interim matches that from GPCC reasonably well, whereas that from ERA5 is clearly larger than the other considered precipitation products. Note that signals such as the precipitation reduction associated with the eruption of Mt. Pinatubo in 1991 as well as strong El Niño–Southern Oscillation (ENSO) events are well captured by all four products. Land averages of evaporation from ERA5 as well as from ERA-Interim exhibit remarkably good temporal stability over the given time span, with no clear trend or change around 2000. They do not differ greatly in terms of temporal variability, but have an offset of $<0.1 \text{ mm day}^{-1}$.

Moisture transport from the ocean to land is realized by positive (negative) VIWVD over ocean (land), which is balanced by the vertical freshwater flux $P + E$. Over land (Fig. 3b), El Niños and the Mt. Pinatubo eruption cause a clear signal in the VIWVD_{AN} and $P + E$ time series from both ERA5 and ERA-Interim. Nonetheless, there are sizeable differences between $P + E$ and VIWVD_{AN} from ERA5 at the beginning of the time series (VIWVD_{AN} is $\sim 17\%$ smaller relative to $P + E$), which get gradually smaller until around 2005. After 2005, these fields agree very well except for an offset of $\sim 0.02 \text{ mm day}^{-1}$. The 1985–2018 land averages are -0.7 mm day^{-1} for VIWVD_{AN} and 0.8 mm day^{-1} for $P + E$.

The results from ERA-Interim, however, show large discrepancies over the whole period. While VIWVD_{AN} steadily decreases until around 2005 and again increases afterward, $P + E$ is relatively stable except for the period 2005–15. Compared to ERA5, moisture transport and freshwater fluxes are significantly weaker in ERA-Interim, with a 1985–2018 land average of -0.6 mm day^{-1} for VIWVD_{AN} and 0.7 mm day^{-1} for $P + E$. Trenberth and Fasullo (2013) estimated a climatological ocean to land moisture transport of -0.7 mm day^{-1} using river discharge data from Dai et al. (2009), which is in better agreement

FIG. 3. Temporal evolution of (a),(c) precipitation and evaporation as well as (b),(d) VIWVD_{AN} and $P + E$ from various products. (top) Land averages and (bottom) ocean averages. Note that the Antarctica is excluded in land averages shown in (a). Solid lines illustrate ERA5 fields and dotted lines ERA-Interim fields, whereas precipitation is colored orange and evaporation is colored blue. Precipitation from GPCP and GPCC are dashed green and cyan, respectively; $P + E$ and VIWVD_{AN} are red and black, respectively. VIWVD_{FC} is shown as a solid gray line. Note that the signs of E and $P + E$ are inverted in this figure. Each line is smoothed using a 12-month running mean. Units are mm day^{-1} on the left figure axes and Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) on the right axes. To convert mm day^{-1} to $\text{kg m}^{-2} \text{ s}^{-1}$, divide by 86 400.

with the moisture transport derived from ERA5 than that from ERA-Interim. Furthermore, $P + E$ must balance the global river discharge plus changes in the terrestrial water storage (TWSC). Dai and Trenberth (2002) estimated the river discharge of the 912 largest rivers to be about $37.2 \times 10^3 \text{ km}^3 \text{ yr}^{-1}$. TWSC can be derived from Gravity Recovery and Climate Experiment satellite observations and is roughly $+80 \text{ km}^3 \text{ yr}^{-1}$ for 2002–09 (Llovel et al. 2010). This is equivalent to a total freshwater flux of roughly 0.67 mm day^{-1} and in good agreement with the long-term mean of $P + E$ from ERA-Interim.

Over the ocean (Fig. 3c), the spread among different precipitation products is substantially larger than over land. Mean precipitation from GPCP is significantly lower than averages from reanalyses, with a 1985–2018 mean of 2.9 mm day^{-1} and good temporal stability, while those from both ERA5 and ERA-Interim have a mean of 3.2 mm day^{-1} . Reanalysis-based precipitation exhibits several discontinuities over the ocean. Precipitation from ERA5 increases in the early 2000s but is relatively stable before and after. Precipitation from ERA-Interim decreases rapidly around 1992, with a further decline until ~ 2005 , when it starts to increase again until 2010. Evaporation products from the reanalyses agree reasonably well, but both exhibit a strong positive trend. Again, ENSO events (e.g., in 1997/98, 2009/10 and 2015/16) are well captured by the products shown, except for P from ERA-Interim, which is mainly dominated by the spurious temporal variability described above.

Nogueira (2020) showed that both ERA-Interim and ERA5 overestimate precipitation mainly over the global ocean (especially in the tropics) and over the Himalaya and Andes when using GPCP as reference. ERA5 shows some improvements over ERA-Interim owing to better representation of deep convection. However, its 1985–2018 mean is only slightly closer to GPCP values. Nonetheless, the strong trends in P from ERA5 (leading to a larger discrepancy with GPCP during 2000–18; see Fig. 3c) are not well understood (Hersbach et al. 2020) and need further investigation. On the other hand, precipitation estimates from GPCP (using rain gauge data and satellite measurements) exhibit larger biases in regions with sparse rain gauge coverage (the relative bias is 10.5% over the ocean and 7.5% over land; Adler et al. 2012) and must thus be treated with caution.

VIWVD_{AN} and $P + E$ from ERA5 agree well over the ocean area (Fig. 3d), particularly between 1992 and 2015, with a 1985–2018 mean of 0.30 mm day^{-1} for VIWVD_{AN} and $-0.29 \text{ mm day}^{-1}$ for $P + E$. VIWVD_{AN} from ERA5 has a positive trend, with a slope of $1.0 \times 10^{-3} \text{ mm day}^{-1} \text{ yr}^{-1}$ (taking autocorrelation into account, confidence intervals are 2.2×10^{-3} and $-4.6 \times 10^{-4} \text{ mm day}^{-1} \text{ yr}^{-1}$ for $\alpha = 0.05$), which means that ocean–land moisture transport in ERA5 suggests a steady increase over the given time span. This would imply a rather strong increase of $\sim 3\%$ per decade, which is at least one order of magnitude larger than the TWSC estimated by Llovel et al. (2010) and several orders of magnitude larger than river discharge trends provided by Su et al. (2018). We

conclude that this strong but insignificant trend is unrealistic, since no statistically significant increase in land precipitation is evident in GPCC or other observation-based precipitation products, and there is no evidence that evapotranspiration over land is decreasing (Hartmann et al. 2013). Oceanic $P + E$ from ERA5 steadily increases before 2000 and decreases after around 2010 making $P + E$ smaller than VIWVD_{AN} before 1992 and after 2015. In ERA-Interim, these fields are much more inconsistent during the entire period, with $P + E$ having an exceptionally large increase (by a factor of 4) between 1991 and 1999 and another sharp decrease between 2008 and 2012. VIWVD_{AN} from ERA-Interim is temporally more stable than $P + E$ but also has a weak positive trend prior to around 2005 and a negative trend afterward. These spurious trends are associated with changes in satellite measurements of atmospheric water vapor (Berrisford et al. 2011; Dee et al. 2011), and inconsistencies in the moisture budget generally stem from analysis increments that can artificially remove or add moisture (Trenberth et al. 2011).

The VIWVD field from ERA5 is also available as instantaneous forecast (VIWVD_{FC}) quantity using twice-daily 12-hourly forecasts, which is also shown in Figs. 3b and 3d. In contrast to VIWVD_{AN} , no clear trend is visible in VIWVD_{FC} indicating that the ocean–land moisture transport is temporally more stable in short-term forecasts compared to the analyses. While forecasted and analyzed VIWVD agree well in the late 1980s, differences after around 2009 are on the order of 0.1 mm day^{-1} over land and $\sim 0.04 \text{ mm day}^{-1}$ over the oceans. This results in a noticeable disparity between $P + E$ and VIWVD_{FC} over land after 2000, a period when $P + E$ agree very well with VIWVD_{AN} . The moisture tendencies from short-term forecasts (not shown), which can be interpreted as a result of the model drifting toward its own climate, indeed balance the disparity between $P + E$ and VIWVD_{FC} qualitatively well (over land, for example, stronger drift before 2000, weaker drift after 2000), but not exact, probably because VIWVD_{FC} represents instantaneous rather than accumulated fields (a note on accumulated fields follows below). Hence, we conclude that $P + E$ agrees better with VIWVD_{AN} after around 2000, especially over land, indicating small model drift and a good balance between observations and model climate. Prior to 2000, however, differences between analyzed and forecast VIWVD are smaller so that the difference to $P + E$ is similarly large for both quantities, which overall indicates the presence of drift in the short-term forecasts.

We note that the ERA5 archive also offers accumulations of forecasted moisture flux divergence. However, this field is computed with numerically less accurate methods (P. Berrisford, ECMWF, 2020, personal communication) that do not satisfy global properties such as a global zero mean (see Gutenstein et al. 2021), and thus is not suited to obtain a best estimate of ocean–land transport and hence is not used in the present study.

In summary, interpretation of temporal variations in the moisture budget in ERA5 is not straightforward. VIWVD_{AN} from ERA5 agrees well with $P + E$ after 2000 indicating a good balance of the assimilation system during that time, but it exhibits a strong and spurious trend, which is not seen in

VIWVD_{FC} . This indicates that temporally varying spinup and spindown effects in P (visible in unrealistic trends in $P + E$) tend to be compensated rather by temporally varying drift in atmospheric moisture than by forecasts of the horizontal moisture transport. At the same time, we see an effect of the varying analysis increments on the analyzed moisture transports as they suffer a rather strong and unrealistic trend.

Overall, these diagnostics indicate that the forecast quantity VIWVD_{FC} may indeed be better suited for long-term studies than VIWVD_{AN} because of its superior temporal stability. However, it should be kept in mind that we here consider averages over large areas. The conclusions may be different for more regional moisture budgets, but further investigations are beyond the scope of this study.

c. Moisture budget residuals

In this section, we investigate the internal consistency of the moisture budget. Keep in mind that we use moisture tendencies calculated from analyzed state quantities to evaluate the moisture budget. Analyzed tendencies are small compared to the flux terms such that the offset between VIWVD_{AN} and $P + E$ in Fig. 3 can be considered a good indicator for the magnitude of moisture budget residuals on a global average. Figure 4 presents zonally averaged anomalies of the moisture budget residuals. Over the land area, ERA5 residuals (Fig. 4a) are generally of the same sign with latitude, with positive anomalies (directed downward) before the discontinuity around 2000 and negative anomalies afterward. To evaluate the location of this discontinuity more precisely, we apply a 4-years moving average standard normal homogeneity test (SNHT; Alexandersson and Moberg 1997; not shown) on times series of zonally averaged residual anomalies. It shows multiple breaks between 1997 and 2003 for the region within $\pm 30^\circ$ latitude that might be caused by changes in the observing system, particularly the increased amount of SSM/I wind observations and the assimilation of ERS-1/2 data in the mid-1990 (Hersbach et al. 2020; Robertson et al. 2020). At higher latitudes, no significant breaks can be identified by the SNHT.

ERA-Interim residuals (Fig. 4b) exhibit a prominent dipole structure at low latitudes, with negative anomalies along the equator and positive anomalies at higher latitudes, suggesting temporally varying strengths of spinup and spindown effects of $P + E$ in the tropics. This structure is, however, inverted in the early 2000s and becomes weaker afterward.

Over the global ocean, ERA5 (Fig. 4c) is more stable than ERA-Interim and features large discrepancies only near the equator following the seasonal cycle of the intertropical convergence zone, which also moves regions of strongest anomalies (between 1995 and 2003 and after 2016). The ERA-Interim budget residual over the ocean area (Fig. 4d) has a smaller mean than that from ERA5, and is in general similar to that over land. It does not exhibit this prominent dipole structure, but variations of the residual are also strongest along the equator. Moreover, anomalies of the ERA-Interim moisture budget residual have a 24% (70%) larger RMS value over land (ocean) compared to those from ERA5. Hence, the small mean of the ERA-Interim residual over ocean may be a result of taking the mean of positive and negative values with similar

FIG. 4. Hovmöller plots of monthly moisture budget residual ($RE_{AN,moist}$) anomalies (mm day^{-1}) for the period 1985–2018 using (a),(c) ERA5 and (b),(d) ERA-Interim. (top) Land averages and (bottom) ocean averages.

magnitude. To conclude, $VIWVD_{AN}$ and $P + E$ agree relatively well in ERA5, while there are large discrepancies in ERA-Interim. In both products moisture budget residuals are largest along the equator.

5. Atmospheric energy budget

a. Divergence of total atmospheric energy flux

Evaluations of TEDIV are central in this paper, since it can be used for indirectly estimating the net surface energy flux F_S but also for consistency estimates (RE_{AN} and RE_{FC}) as defined in Eqs. (5) and (6), which are useful indicators for consistency problems and inhomogeneities. Here we focus on the consistency aspect. Updated indirect estimates of F_S fields as presented, e.g., in Mayer et al. (2017) and Trenberth and Fasullo (2018) will be discussed in future work.

We present 1985–2018 averages of different TEDIV products in Fig. 5. Figure 5a displays the $TEDIV_{dir}$ derived from ERA5 using advanced numerical and diagnostic methods introduced above, and Fig. 5b shows the same field but from previous budget evaluations using ERA-Interim. Additionally, we depict the indirectly estimated $TEDIV_{ind}$ from parameterized fluxes as they are stored in ERA5. Each field is spectrally truncated at wavenumber 179. Regions of positive TEDIV are indicative of divergent lateral energy fluxes, which is generally the case in the tropics, but most notably over western boundary currents (Sverdrup 1947; Stommel 1948; Seager and Simpson 2016) where the atmosphere is supplied with energy from warm water masses that are transported from the equator poleward, e.g., along the Gulf Stream and Kuroshio in the Northern Hemisphere (NH), and the Agulhas and East Australian Currents in the Southern Hemisphere (SH). Convergent lateral energy fluxes (negative TEDIV) can primarily be found in polar regions where Earth is losing energy to space. It is also noted that the gradients along sea ice

edges, western boundary currents, and coastal lines are sharper in the novel ERA5 fields (Figs. 5a and 5c; see also Mayer et al. (2019) for closeups) than in previous assessments, where stronger truncations had to be applied to remove the much stronger spectral noise (see, e.g., Trenberth and Fasullo 2018). Moreover, the fields in Fig. 5 are much smoother over high topography (e.g., along the Andes, Rocky Mountains, and Himalayas) reducing the RMS by at least $\sim 10\%$ relative to that from ERA-Interim. Note that this improvement not only comes from the enhanced methods that are used, but also from the higher spatial resolution of ERA5. That is, the truncation to T179 reduces the native spatial resolution of ERA5 by $\sim 70\%$, while it reduces that of ERA-Interim by only $\sim 30\%$, i.e., the relative truncation is stronger in ERA5 which may contribute to the stronger noise reduction in ERA5. However, there is still some residual noise over the Andes and Himalaya as these regions exhibit the strongest artificial noise. $TEDIV_{ind}$ looks similar to $TEDIV_{dir}$ derived from ERA5, both are very smooth over land compared to $TEDIV_{dir}$ from ERA-Interim. $TEDIV_{ind}$ has a 1985–2018 global mean of -2.9 W m^{-2} , which indicates an inconsistency between forecast vertical fluxes and energy tendency. Note that this discrepancy is larger than that found in ERA-Interim using the same method (see Mayer and Haimberger 2012). The reason for this is unclear and warrants further investigation.

The benefit of 1-hourly temporal resolution for budget evaluations from ERA5 is highlighted in Fig. 6. The left panel shows monthly mean $TEDIV_{dir}$ derived from ERA5 for January 2010 based on 1-hourly evaluation. The right panel depicts the same, but is evaluated with a reduced temporal resolution of 6 h. While the large-scale pattern remains unaffected by the temporal resolution, regional differences attain values up to $\sim 300 \text{ W m}^{-2}$, especially in midlatitudes where fast moving cyclones cannot be sampled properly. Considering annual averages these differences are still on the order of $\sim 90 \text{ W m}^{-2}$ (not shown). Note that both fields are presented at

FIG. 5. Global maps of TEDIV (W m^{-2}) averaged over the period 1985–2018. It shows the TEDIV derived from (a) ERA5, (b) ERA-Interim, and (c) parameterized fluxes from ERA5. Each field is spectrally truncated at wavenumber 179.

full spectral resolution T639 such that the artificial noise over high topography is not filtered out, which is already reduced (due to the iterative mass correction) compared to the noise in mass-unadjusted TEDIV_{dir} fields (not shown). A further reduction of this noise will be addressed in future work.

FIG. 6. Monthly averaged TEDIV_{dir} field (January 2010; W m^{-2}) at full spectral resolution T639 using ERA5 with (a) 1-hourly and (b) 6-hourly temporal resolution. Please note the different scaling compared to Fig. 5.

In Fig. 7, Hovmöller plots of TEDIV_{dir} anomalies from ERA5 (left panels) and ERA-Interim (right panels) are presented. The upper panels show anomalies for the global land area and the lower panels show anomalies for the global ocean. Over land, anomalies of TEDIV_{dir} derived from ERA5 are dominated by an abrupt change of sign in the late 1990s, with primarily positive anomalies before and negative anomalies afterward. The ERA-Interim analog in Fig. 7b does not exhibit this abrupt change, but exhibits a meridional dipole structure with positive anomalies at the equator and negative anomalies along subtropical latitudes after around 2005, and vice versa before. The noise at high latitudes mainly stems from regions where the sparse number of grid points contributing to the zonal mean are dominated by the artificial noise over high topography.

Over the ocean, TEDIV_{dir} anomalies from ERA5 are similar to those from ERA-Interim. However, there are some differences. The largest disparity can be seen between 2002 and 2010 along 10°N, where ERA-Interim is mostly negative and ERA5 has no uniform pattern. There are some minor differences in the amplitude of individual peaks at the equator representing the ENSO, e.g., the positive anomaly at the beginning of 1998 and the subsequent negative anomaly are more pronounced in ERA5. Also, TEDIV_{dir} from ERA-Interim is temporally less stable in the tropical region (30°N–30°S) after about 2010 ($\text{RMS} = 5.8 \text{ W m}^{-2}$ in this region whereas that from ERA5 is $\text{RMS} = 4.9 \text{ W m}^{-2}$ for the 2010–18 period) as a decreasing number of satellite data were assimilated in ERA-Interim during this period [see, e.g., Mayer et al. (2018)]. TEDIV_{dir} anomalies derived from ERA5 are slightly more negative over the tropical ocean before 2000 and positive or close to zero afterward. The TEDIV_{dir} ocean average for 30°N–30°S is 34.1 W m^{-2} for 1985–1999 and 35.8 W m^{-2} for 2000–18, i.e., the changes in the TEDIV over the global ocean mainly stem from changes in the tropics. This is coincident with a sudden increase of ERA5's 10-m wind speed and evaporation in the eastern tropical Pacific during the late 1990s documented by Robertson et al. (2020), which they attributed to changes in the observing system. In summary, TEDIV_{dir} anomalies over the global ocean are very similar in ERA5 and ERA-Interim. Over land, ERA5 exhibits a persistent and spatially relatively uniform transition from positive to negative anomalies in the late 1990s, whereas ERA-Interim has some spatial discrepancies as well.

FIG. 7. Hovmöller plots of $\text{TEDIV}_{\text{dir}}$ anomalies (W m^{-2}) (top) over land and (bottom) over the ocean derived from (a),(c) ERA5 and (b),(d) ERA-Interim.

b. Meridional energy transport

In this section, we estimate the atmospheric meridional energy transport (AMET) for the period 2010–18 using various TEDIV fields derived from ERA5 as well as ERA-Interim (Fig. 8). Meridional transports are obtained by inverting the Laplacian applied to TEDIV and computing the meridional derivative of the resulting potential function.

The maximum AMET derived from 1-hourly evaluated $\text{TEDIV}_{\text{dir}}$ using ERA5 is 4.58 PW (peak value standard deviation $\sigma_{\text{peak}} = 0.07$ PW and peak latitude $\sigma_{\text{lat}} = 0.28^\circ$) in the NH and -5.19 PW ($\sigma_{\text{peak}} = 0.07$ PW, $\sigma_{\text{lat}} = 0.54^\circ$) in the SH both around 40° latitude, whereas that from ERA-Interim is considerably stronger with 4.74 PW ($\sigma_{\text{peak}} = 0.09$ PW, $\sigma_{\text{lat}} = 0.17^\circ$) and -5.28 PW ($\sigma_{\text{peak}} = 0.10$ PW, $\sigma_{\text{lat}} = 0.41^\circ$). To rule out the differing spatial and temporal resolution of these products as potential source for this discrepancy, we also derive the AMET from ERA5 products with reduced resolution. That is, we also evaluate $\text{TEDIV}_{\text{dir}}$ 1) with 6-hourly temporal but same spatial resolution, and 2) using one ERA5 ensemble member with 6-hourly temporal and reduced spatial (N160) resolution. The former has a maximum transport of 4.55 PW in the northern and -5.13 PW in the Southern Hemisphere, whereas the latter is at 4.53 PW and -5.11 PW. All three ERA5 products agree well within ~ 0.1 PW, as Fig. 8 shows. Thus, the discrepancy between ERA5 and ERA-Interim does not result from differences in spatial or temporal resolution. Moreover, this result suggests that the AMET is quite robust to changes of the spatial resolution of the reanalysis and that 6-hourly temporal resolution is sufficient to capture relevant eddy heat fluxes, at least when considering zonal integrals of the transports.

Additionally, we derive the AMET from the indirectly estimated $\text{TEDIV}_{\text{ind}}$, which is slightly weaker than that from the aforementioned ERA5 products. It has a maximum of 4.46 PW in the northern and -5.28 PW in the Southern Hemisphere. Note that $\text{TEDIV}_{\text{ind}}$ is uniformly adjusted to zero global mean

by subtracting its 2010–18 average of -3.06 W m^{-2} before the AMET is computed.

We also show the total (atmosphere plus ocean) meridional energy transport derived from CERES and parameterized net TOA fluxes from ERA5 and ERA-Interim, which are also uniformly adjusted to zero global mean (2010–18 averages are 1.07, 0.64, and -1.93 W m^{-2} , respectively). The total energy transport derived from CERES has a peak annual mean of 5.90 PW in the NH and -5.77 PW in the SH. The total energy transports derived from parameterized net TOA fluxes are substantially smaller, with maxima of 5.33 and -5.34 PW in ERA5 and 5.11 and -5.27 PW in ERA-Interim. This is a 10 (14) % weaker poleward transport of energy in ERA5 (ERA-Interim) in the NH when using CERES as reference. In the Southern Hemisphere, differences are in general smaller (7% weaker in ERA5 and 9% weaker in ERA-Interim). This indicates that net TOA flux biases in ERA5 are qualitatively similar to those in ERA-Interim, with too little energy absorption in the tropics and too little energy loss in high latitudes (consistent with Mayer and Haimberger 2012), but they are reduced in ERA5, leading to a more realistic total meridional energy transports in this product.

The cross-equatorial total energy transport derived from CERES-EBAF net TOA fluxes is northward (+15 TW). Combining this with our estimates of cross-equatorial AMET that range between -380 TW (low-resolution ERA5 ensemble member) and -540 TW ($\text{TEDIV}_{\text{ind}}$ from ERA5) yields an oceanic cross-equatorial northward energy transport of about 480 ± 80 TW (derived from the average of our AMET estimates), which is in good agreement with estimates provided by Liu et al. (2020).

In contrast to the CERES-based estimate of 15 TW, the total cross-equatorial energy flux is southward when using reanalysis data (-151 TW in ERA5 and -361 TW in ERA-Interim), which implies a hemispheric asymmetry in the radiation bias of the

FIG. 8. (a) Meridional energy transport for the period 2010–18. The total transport is calculated from CERES (solid red) and parameterized R_{TOA} fluxes from ERA5 (dotted red) and ERA-Interim (dashed red line). The atmospheric transport (black and gray lines) is derived from TEDIV using: ERA5 with 1-hourly (solid black) and 6-hourly (dotted black) temporal resolution, low-resolution ERA5 ensemble member (dash-dotted black), parameterized fluxes from ERA5 (dashed black), and ERA-Interim (solid gray). Positive values represent northward transport of energy. Units are petawatts ($1 PW = 10^{15} W$). (b) Zonal averages of absorbed shortwave (ASR; green lines; $W m^{-2}$) and outgoing longwave radiation (OLR; blue lines; $W m^{-2}$) from CERES-EBAF (solid lines), ERA5 (dotted lines), and ERA-Interim (dashed lines).

reanalyses. To understand this better, we show in Fig. 8b zonal averages of the absorbed shortwave (ASR) and outgoing longwave radiation (OLR) at TOA from ERA5, ERA-Interim, and CERES. ASR and OLR from ERA5 are at almost all latitudes closer to the CERES radiation (RMS difference is $7.8 W m^{-2}$ for ASR and $3.2 W m^{-2}$ for OLR) than those from ERA-Interim (RMS differences are 10.7 and $6.1 W m^{-2}$, respectively), i.e., ERA5 exhibits a stronger (weaker) ASR (OLR) in the tropics. At higher latitudes, both ASR and OLR are weaker in ERA5 than in ERA-Interim. This results in generally stronger and more realistic (when using CERES as reference) R_{TOA} ($R_{TOA} = ASR - OLR$) fluxes in ERA5, particularly at low latitudes and in the SH. An outstanding feature of Fig. 8b is the positive ASR bias in both ERA-Interim and ERA5 over the Southern Ocean, which represents a long-standing bias common to many atmospheric models (see, e.g., Hyder et al. 2018), and likely contributes to the negative bias in cross-equatorial energy flux discussed above.

In addition to the stronger net TOA input in the tropics, ERA5 also exhibits an increase in parameterized net downward surface energy fluxes in the tropics (mainly arising from stronger shortwave absorption and weaker evaporation; not shown) compared to ERA-Interim, which is more pronounced than the changes in R_{TOA} . This results in a weaker convergence

of vertical energy fluxes ($R_{TOA} - F_S$) in ERA5 in the tropics and the SH, and thus in a weaker atmospheric meridional energy transport in ERA5.

c. Ocean–land energy transport

Here, we assess the temporal evolution of the ocean–land energy transport and other relevant energy budget terms. The transport from ocean to land is realized by positive TEDIV over the ocean (or convergence over land). Energy transports associated with river discharge are of order $\sim 1.2 \times 10^6 m^3 s^{-1}$ (Dai and Trenberth 2002; Dai et al. 2009), which is equivalent to an enthalpy flux of $\sim 0.3 W m^{-2}$ for an assumed temperature difference of 10 K between the river and ocean, and are thus neglected here. Figure 9 shows time series of parameterized as well as reconstructed net TOA fluxes, $TEDIV_{dir}$ derived from ERA5, and F_S inferred from the latter two. In addition, we show the inferred F_S using R_{TOA} from CERES-EBAF v4.1 for the period 2001–18 as well as the mass-unadjusted TEDIV, as it is stored in ERA5, and $TEDIV_{ind}$. The left column presents absolute fluxes using a 12-month running mean and the right column displays corresponding anomalies using a 13-month Gaussian filter (Trenberth et al. 2007; Solomon et al. 2007), whereas land averages are at the top and ocean averages at the bottom.

Parameterized and reconstructed net TOA fluxes are in good agreement from a global perspective. Parameterized TOA fluxes have a global mean of $0.51 W m^{-2}$ and those from observations and reconstructions $0.45 W m^{-2}$ for the full period. Both products capture ENSO events with stronger (weaker) energy input during La Niña (El Niños). ERA5 uses realistic aerosol forcing (Hersbach et al. 2020) which is reflected in the negative R_{TOA} anomalies associated with the Mt. Pinatubo eruption in 1991 and was not represented in ERA-Interim (e.g., Dee et al. 2011). However, the impact on parameterized TOA fluxes appears to be longer lasting compared to DEEP-C. It is also noteworthy that the El Niño event in 2009/10, which was significant in terms of moisture (see Fig. 3) but only moderate in terms of SST anomalies, exhibits a strong TOA signal in both products. We also applied the standard normal homogeneity test to the DEEP-C TOA fluxes, which did not reveal clear jumps neither over ocean nor land.

While both net TOA flux products are stable over the full period and show no significant trend, we find a gradual change between 1996 and 2003 in all TEDIV time series shown (Fig. 9a). $TEDIV_{dir}$ derived from ERA5 has a 1985–2018 mean of $-15.7 W m^{-2}$ over land, with positive anomalies of around $\sim 1.5 W m^{-2}$ before the discontinuity and $-1.0 W m^{-2}$ afterward. $TEDIV_{ind}$ is very similar to $TEDIV_{dir}$ but has an offset of about $2.0 W m^{-2}$, i.e., ocean to land energy transport is stronger when derived from $TEDIV_{ind}$. The mass-unadjusted TEDIV exhibits an even stronger discontinuity resulting in a stronger (weaker) ocean–land transport after (before) 2000 compared to our mass-adjusted $TEDIV_{dir}$. Decomposition of the energy transports (not shown) reveals that the latent heat term accounts for the majority (about $-20 W m^{-2}$) of TEDIV when averaged over the land. Enthalpy and geopotential transports are of opposite sign and approximately balance each other (the kinetic energy term is an order of magnitude

FIG. 9. Time series of individual atmospheric energy budget terms for the period 1985–2018. (a),(b) Land averages and (c),(d) ocean averages. Absolute values are depicted in (a) and (c) using a 12-month running mean, while anomalies are shown in (b) and (d) using a 13-month Gaussian filter. Shown are R_{TOA} from DEEP-C (solid red lines), $\text{TEDIV}_{\text{dir}}$ evaluated from ERA5 (solid black), indirectly estimated F_S (solid orange), and parameterized net TOA fluxes from ERA5 (dotted red). Additionally, we show in (a) and (c) the indirectly estimated F_S derived from CERES (instead of DEEP-C) as yellow dashed line, and in (a) and (b) the archived $\text{TEDIV}_{\text{dir}}$ from ERA5 as black dotted line and indirectly estimated $\text{TEDIV}_{\text{ind}}$ derived from parameterized fluxes as black dashed line. Missing values during the first and last year are hidden. Units are W m^{-2} on the left axes and PW (10^{15} W) on the right axes.

smaller). In the non-mass-adjusted TEDIV , however, we find a disparity between these two terms of about 6 W m^{-2} before and 2 W m^{-2} after 2000 (over land) explaining the gradual change in TEDIV , i.e., the temporal discontinuities of opposite sign in the two terms do not cancel out. In our mass-adjusted $\text{TEDIV}_{\text{dir}}$ this imbalance still exists but is reduced by $\sim 50\%$.

This gradual change might stem from the changing observing system. The ERA5 assimilation system was initially tested in the period after 2000, when generally more observational data (such as GPS-RO and AMSU) have been available than before. We also note that ocean–land energy transports in ERA-Interim (not shown; see Trenberth and Fasullo (2013) for a figure) are temporally more stable, with a temporal standard deviation of $\sim 0.3 \text{ W m}^{-2}$ compared to $\sim 0.5 \text{ W m}^{-2}$ in ERA5. Reasons for the differences between ERA-Interim and ERA5 may be the fact that some observational data that are assimilated by ERA5 have been reprocessed compared to the versions used in ERA-Interim (e.g., SSM/I microwave imager observations; see Fig. 5 in Hersbach et al. 2020), and the use of different SST products before 2009. Before that time, typical SST differences between the two products (measured as temporal standard deviations of monthly mean differences) range between $\sim 0.2 \text{ K}$ in the tropics and $\sim 0.5 \text{ K}$ in the Southern Ocean, western boundary currents, and in the high north (not shown). It is likely that these differences in SST contribute to differences in surface fluxes and also have an effect on assimilated state quantities away from the surface and consequently analyzed transports.

Consequently, this gradual change in $\text{TEDIV}_{\text{dir}}$ is also present in our indirectly estimated F_S product (derived from $\text{TEDIV}_{\text{dir}}$ and R_{TOA} from DEEP-C) such that its land average is -4.3 W m^{-2} prior to 2000 and -1.4 W m^{-2} afterward, which is in reasonable agreement with observation-based estimates of $\leq 0.1 \text{ W m}^{-2}$ from von Schuckmann et al. (2020). Aside from the large discontinuity, F_S is relatively stable around those values having a standard deviation of 0.7 W m^{-2} before and 0.4 W m^{-2} after the year 2000. The inferred F_S using R_{TOA} from CERES-EBAF v4.1 exhibits the same temporal variability as F_S derived from DEEP-C, but has a 2000–2018 land average of only -1.2 W m^{-2} (see Fig. 9a) owing to the 0.2 W m^{-2} weaker net TOA flux in CERES-EBAF over land, which is a result of a different global anchoring value in the two products (Liu et al. 2020). Consequently, its ocean average is about 0.2 W m^{-2} larger when using CERES-EBAF (long-term mean is 1.8 W m^{-2} ; see Fig. 9c) instead of DEEP-C (see further below). This is a fairly small difference and can be attributed to observational uncertainties [e.g., von Schuckmann et al. (2020)] and different averaging periods for the choice of anchoring values for DEEP-C and CERES-EBAF TOA fluxes.

El Niño events and volcanic eruptions do not only have an impact on the surface and net TOA fluxes, but also on the ocean–land transport. Figures 9b and 9d highlights the strengthened ocean–land transport during these events, e.g., in 1992, 1998, 2010, and 2016, with the largest anomalies during the strong 1997/98 El Niño.

FIG. 10. Global maps of 1985–2018 averages (W m^{-2}) of the (top) self-consistency (RE_{FC}) and (bottom) budget residuals (RE_{AN}) using (a),(c) ERA5 and (b),(d) ERA-Interim.

Over the ocean (Fig. 9c), variability of F_S is coherent with the R_{TOA} variability so that it is also dominated by the signals of El Niños and volcanic eruptions. The discontinuity in $\text{TEDIV}_{\text{dir}}$ (see also Fig. 7 and corresponding discussion) causes a slight decrease of F_S after 2000 resulting in a 2000–18 mean of 1.6 W m^{-2} (energy flux into the ocean). Prior to 2000, F_S has an ocean mean of 1.9 W m^{-2} , and for 1985–2018 it is 1.7 W m^{-2} . This is in remarkably good agreement with observation-based estimates of net surface energy fluxes, e.g., von Schuckmann et al. (2020) estimated an average ocean heat uptake of 0.9 W m^{-2} for the 0–2000-m ocean depth layers during the period 1993–2018, and L’Ecuyer et al. (2015) derived an oceanic heat flux of 0.45 W m^{-2} (for the period 2000–09) from various independent flux datasets. For comparison, indirectly estimated F_S derived from ERA-Interim and DEEP-C (not shown) has a 1985–2018 ocean mean of 2.1 W m^{-2} . Please keep in mind that the calibration of DEEP-C R_{TOA} fluxes incorporates, among others, information about long-term OHC trends. This is similar to CERES-EBAF, but DEEP-C is adjusted to an updated ocean heat uptake estimate based on Argo data (see Liu et al. 2020). Nevertheless, the good agreement between F_S over ocean and global ocean OHC trends suggests a realistic balance between TOA fluxes and TEDIV over ocean.

We also computed the ocean to land energy transport for 2000–06 using the new ERA5.1 dataset (not shown; Simmons et al. 2020), where a cold bias in the stratospheric temperature is corrected, but this has virtually no effect on the global energy transport. Differences in the ocean to land energy transport are on the order of $\sim 0.1 \text{ W m}^{-2}$.

As for the moisture transport, there are forecasts available for the TEDIV (note that these are not mass-adjusted; not shown) exhibiting the same characteristics as the TEDIV fields

(based on analyses) presented above, i.e., a gradual change with the same magnitude around the year 2000. We also see this trend in the AET_{FC} and parameterized F_S over the ocean (not shown) indicating an altered balance of the forecasted energy budget likely caused by changes in the observing system.

In summary, ocean to land energy transport in ERA5 exhibits a discontinuity of $\sim 3 \text{ W m}^{-2}$ (0.45 PW) in the late 1990s so that inferred F_S after 2000 agrees within 1.5 W m^{-2} with recent estimates of ocean and land heat uptake, but disagreement is significantly larger before 2000 (all long-term averages are summarized in appendix B).

d. Internal consistency of the atmospheric energy budget

We now assess the internal consistency of the ERA5 and ERA-Interim energy budget using the residual (computed with analyzed tendencies) and self-consistency diagnostics (computed with forecasted tendencies) as described in section 3. Again, we want to emphasize that these assessments of internal consistency are independent of observation-based vertical energy fluxes, i.e., we use parameterized R_{TOA} and F_S fluxes from reanalyses. Figure 10 shows global 1985–2018 averages of the self-consistency in the upper and residuals in the lower panels. Results from ERA5 are shown in the left and those from ERA-Interim in the right column.

The self-consistency check, i.e., the agreement between $\text{TEDIV}_{\text{dir}}$ and $\text{TEDIV}_{\text{ind}}$, in ERA5 (Fig. 10a) shows rather uniform values over both ocean and land. Largest inconsistencies can be found over high topography, e.g., the Andes and Himalaya, where spectral noise is present in $\text{TEDIV}_{\text{dir}}$ (cf. Figs. 5a,c), and in the Southern Ocean. The latter is possibly related to our neglect of snow and ice in our

FIG. 11. Zonally averaged self-consistency (RE_{FC}) and residual (RE_{AN}) anomalies ($W m^{-2}$) of the total atmospheric energy budget for the period 1985–2018. Panels are arranged as in Fig. 10.

budget formulation. The global mean is $-2.9 W m^{-2}$ for the full period and $-3.0 W m^{-2}$ for 2000–18. Thus, there is an imbalance between the vertical energy fluxes and AET_{FC} (the global average of $TEDIV_{dir}$ is 0), which suggests that some energetic effects are missing in our diagnostics. The self-consistency of results from ERA-Interim (Fig. 10b) exhibits comparatively strong artificial noise over land originating from the directly computed $TEDIV_{dir}$ (see Fig. 5b), but is as smooth over ocean as in ERA5. Its global mean is -0.9 and $-0.8 W m^{-2}$ for the aforementioned periods, which is significantly lower than in ERA5. This is a surprising result given the evaluations are carried out analogously for both ERA5 and ERA-Interim.

The energy budget residuals from both ERA5 and ERA-Interim (Figs. 10c,d; emphasizing the balance between analysis-based horizontal and forecast-based vertical energy fluxes) are spatially more heterogeneous compared to their self-consistency, with generally larger RMS values (cf. RMS = $11.1 W m^{-2}$ for ERA5 and RMS = $27.0 W m^{-2}$ for ERA-Interim with self-consistency RMS = $8.5 W m^{-2}$ for ERA5 and RMS = $25.2 W m^{-2}$ for ERA-Interim). The self-consistency metric for ERA5 (Fig. 10c) shows relatively high values where precipitation dominates freshwater fluxes (positive values in Fig. 1a), especially in the ITCZ. This appears related to the negative ASR and positive OLR biases (see Fig. 8b), resulting in a negative bias of net TOA fluxes in the tropics. Other inconsistencies may stem from different sources and need further investigation. Nonetheless, compared to other energy budget residual estimates in the literature, a value of $11.1 W m^{-2}$ for a multiannual RMS at T179 spectral resolution (equivalent to 1°) is remarkably small. It is substantially improved compared to estimates from ERA-Interim [$27.0 W m^{-2}$ at T179, $9.4 W m^{-2}$ at T63 (Mayer et al. 2017), and $13.0 W m^{-2}$ at T63 (Hantel and Haimberger 2016)]. ERA-Interim energy budget residuals (Fig. 10d) are dominated by the artificial noise over land, but are

homogeneously negative over the ocean. The global 1985–2018 average of the ERA5 (ERA-Interim) budget residual is $-5.3 W m^{-2}$ ($-7.8 W m^{-2}$), which is significantly more negative than the global mean of the self-consistency diagnostic (-2.9 and $-0.9 W m^{-2}$, respectively). This (i) demonstrates the improved balance between analyses and short-term forecasts in ERA5 and (ii) confirms that forecasted tendencies (as used in the self-consistency diagnostic) largely cancel spinup and spindown effects of the vertical fluxes.

In Fig. 11, we show the temporal evolution of the zonally averaged self-consistency as well as budget residual as anomalies, which is arranged in the same way as Fig. 10. We find that self-consistency in ERA5 (Fig. 11a) is temporally homogeneous, with an RMS = $1.5 W m^{-2}$ and no clear patterns and good temporal stability. In ERA-Interim (Fig. 11b), however, regions and times of positive and negative anomalies are present, which change sign on decadal time scales, e.g., negative anomalies around $30^\circ S$ prior to the year 1997 and positive anomalies afterward, resulting in an RMS = $2.3 W m^{-2}$.

The temporal evolution of the ERA5 budget residual (Fig. 11c) is clearly dominated by the discontinuity in the late 1990s, with RMS = $3.8 W m^{-2}$. Except for the polar regions, it has uniformly negative anomalies before 2000 and positive anomalies afterward. The SNHT shows significant jumps in the tropical region ($\pm 15^\circ$) in 1997/98 and at $30^\circ S$ in 1995, which are at least two years earlier than the discontinuities in ERA5's moisture budget residuals (see Fig. 4a; note that both land and ocean anomalies of the energy budget residual exhibit the same discontinuity as the global anomalies in Fig. 11c). In ERA-Interim (Fig. 11d), we again see the spatially more heterogeneous pattern with alternating signs on decadal time scales, but with larger anomalies compared to its self-consistency (RMS = $4.1 W m^{-2}$ compared to $2.3 W m^{-2}$).

In summary, inconsistencies caused by the model spinup and spindown effect are small in ERA5 and buried by the

discontinuity around the year 2000 (see Fig. 11c). Consequently, anomalies based on the period 2000–2018 (not shown) are even smaller, with an RMS = 2.3 W m^{-2} (compared to 3.6 W m^{-2} for ERA-Interim). In ERA-Interim, spinup effects are causing large temporal and spatial inconsistencies, indicating a degraded regional balance between forecasted and analyzed energy fluxes (see Fig. 11d). On a global average, these inconsistencies are well compensated by the forecasted tendency such that the self-consistency global mean is smaller (i.e., better consistency between model forecasts and analyses) in ERA-Interim than in ERA5. Nonetheless, the regional discrepancies between forecasted and analyzed energy transports (recall $\text{RE}_{\text{FC}} = \text{TEDIV}_{\text{ind}} - \text{TEDIV}_{\text{dir}}$) obviously vary temporally and spatially in ERA-Interim, leading to temporal and spatial variations in the self-consistency metric for this reanalysis (see Fig. 11b). In ERA5, however, these inconsistencies as well as the discontinuity around 2000 are effectively cancelled out by the forecasted tendency resulting in a very uniform but more negative self-consistency (see Fig. 11a).

6. Discussion

We investigate the atmospheric energy budget in ECMWF's latest reanalysis dataset ERA5. We use a mass-consistent formulation of the energy budget in combination with advanced numerical and diagnostic methods to compute the divergence of total atmospheric energy fluxes with unprecedented accuracy. These fields are used to 1) estimate the atmospheric meridional energy transport, 2) compute the ocean–land energy transport, 3) indirectly estimate surface energy fluxes, and 4) assess the consistency of the energy budget. Furthermore, we evaluate the mass and moisture budget in ERA5 and compare individual terms with those from previous evaluations using ERA-Interim. We also compare parameterized precipitation fluxes from ERA5 and ERA-Interim with observation-based precipitation estimates. The main outcomes of this study are the following:

- Freshwater fluxes from ERA5 are generally well balanced by the analyzed moisture flux divergence, with a moderate spindown effect in the tropics. Moreover, global ocean to land moisture transports agree well with the continental freshwater flux. Also, the terms of the moisture budget in ERA5 are temporally more stable than those from ERA-Interim. However, globally averaged moisture transports from analyses exhibit an unrealistic positive trend over the considered period. This spurious trend is not seen in moisture transports derived from short-term forecasts, which makes those better suited for long-term studies. However, it remains unclear whether this conclusion also holds for more regional moisture budgets.
- Precipitation, evaporation, and ocean to land energy transport exhibit an unrealistically strong increase in the equator region during the period 1995–2005, particularly in 1997/98, which likely stems from changes in the observing system.
- TEDIV from ERA5 is much smoother than in earlier evaluations, thanks to ERA5's enhanced spatial and temporal resolution. Results can thus be used at much higher resolution than before.
- Peak meridional energy transport derived from TEDIV is 4.6 PW in the NH and -5.2 PW in the SH in ERA5. This is at least $\sim 0.1 \text{ PW}$ weaker than in ERA-Interim. Investigations show that this cannot be explained by ERA5's higher temporal and spatial resolution, but by stronger convergence of vertical energy fluxes in the tropical region compared to ERA-Interim.
- Ocean to land energy transports in ERA5 are stable for the period 2000–18 such that major climate signals like effects of ENSO events can be robustly quantified, and indirectly inferred surface energy fluxes over land (1.4 W m^{-2}) agree well with observation-based estimates of land heat storage to within $\sim 1.3 \text{ W m}^{-2}$. The generally good temporal stability of the obtained energy transports is partly achieved through the applied mass adjustment, i.e., archived energy transports exhibit stronger spurious trends and thus are less well suited for climate studies.
- Inconsistencies caused by the model spinup and spindown (referred to as residuals) are substantially smaller in ERA5 compared to ERA-Interim. Furthermore, energy as well as moisture budget residuals are temporally and spatially more homogeneous in ERA5 than in ERA-Interim.

The very likely spurious increase of forecast precipitation and evaporation and also the shift in the energy budget residual in the late 1990s warrants further discussion. ERA5 uses the 41r2 cycle version of the Integrated Forecasting System operational in 2016, and the ERA5 assimilation system was initially tested in the period after 2000, when e.g., GPS-RO and AMSU data have been available. This means that it uses the model and data assimilation at that time, which was tuned to work best with the observing system operational at that time (Simmons et al. 2020). This is also reflected in Fig. 8c in Hersbach et al. (2020), which clearly shows that the energetic consistency of ERA5 (in terms of the balance between global mean net flux at TOA and the surface) is best in the late 2000s and early 2010s. Hence, it is not surprising that inferred surface energy fluxes over land (see Fig. 9) are more realistic after 2000 than before.

Furthermore, almost all satellite observations that were assimilated by ERA-Interim have also been used by ERA5, but many of them are reprocessed, such as SSM/I microwave observations [see Fig. 5 in Hersbach et al. (2020)]. This could also have an effect on the transports as the spurious trend in ERA5's energy budget around 2000 is larger compared to that in ERA-Interim (see, e.g., Trenberth and Fasullo 2013). In addition, we also note that ERA5 uses a different SST product than ERA-Interim before 2009, which likely contribute to the found differences in surface fluxes and subsequently atmospheric transports.

Nonetheless, our novel $\text{TEDIV}_{\text{dir}}$ fields show unprecedented accuracy, with sharp gradients along coastal lines, ice edges, and western boundary currents such that regionally more detailed evaluations are possible. Earlier studies using older reanalysis datasets usually truncated results at T63 or below to obtain reasonably smooth fields (see, e.g., Trenberth and Fasullo 2008; Mayer et al. 2016; Trenberth and Fasullo

TABLE A1. List of acronyms for Algorithm A1. Parameter ID values as used by the ECMWF. The data are either used on a reduced Gaussian grid (RGG) or as spherical harmonics (SH). Asterisks mark input variables.

Variable	Description	Parameter ID	Representation
PHIS*	Surface geopotential	129	SH
VORT*	Vorticity	138	SH
DIV*	Divergence	155	SH
Q^*	Specific humidity	133	RGG
T^*	Temperature in kelvin	130	SH
LNPS*	Logarithm of surface pressure	152	SH
VIQ*	Total column water vapor	137	RGG
UV	Meridional and zonal wind component	131/132	—
MASSDIV	Vertical integral of divergence of mass flux	162081	—
MASSTEND	Vertical integral of mass tendency	162092	—
QTEND	Vertical integral of moisture tendency	—	—
VIWVD _{AN}	Vertical integral of divergence of moisture flux	162084	—
DIVERR	Error divergence, residual of the total mass budget	—	—
UVERR	Two-dimensional wind field adjustment	—	—
EKIN	Kinetic energy	—	—
EPOT	Geopotential energy	—	—
ELAT	Latent heat	—	—
EENT	Enthalpy of dry air	—	—
TEFLUX	Vertical integral of total atmospheric energy flux	—	—
TEDIV _{dir}	Divergence of TEFLUX	162086	—

2017), but this obviously prevents a realistic representation of sharp gradients (see Fig. 5 in Mayer et al. (2019) for a comparison of different truncations). Even at full spatial resolution (see, e.g., Fig. 6a), our fields are very smooth over ocean ($\text{RMS} = 62.3 \text{ W m}^{-2}$), while, e.g., those archived from ERA5 show extensive patterns of artificial noise ($\text{RMS} = 159.0 \text{ W m}^{-2}$; not shown). At 1° resolution (truncation at wavenumber 179; see Fig. 5a) and below, our TEDIV_{dir} field unfolds its full potential, where almost no artificial noise is left over near high topography. Some remainders are visible only along the Andes, while other regions show no artificial noise at all, e.g., in North America, Greenland, and Siberia.

We also want to briefly highlight the indirectly estimated surface energy fluxes F_S using TEDIV_{dir} and reconstructed R_{TOA} fields from DEEP-C. The F_S over ocean (1.6 W m^{-2} for 2000–18) is in much better agreement with the long-term mean ocean heat uptake of $<1.0 \text{ W m}^{-2}$ (Wild et al. 2012; L'Ecuyer et al. 2015; von Schuckmann et al. 2020) than widely used

satellite-derived surface flux products (uncertainties of $\sim 20 \text{ W m}^{-2}$; Rhein et al. 2013; Mayer et al. 2017). Over land, we find good agreement with a recent observation-based estimate of land heat uptake from von Schuckmann et al. (2020). We also note that the F_S global mean is unbiased by construction (global average of TEDIV_{dir} is 0) and thus well-suited for climate studies and model evaluations. Future work will include a comprehensive evaluation of F_S and additional in-depth studies.

We used the RMS value of the self-consistency and residual diagnostic as metric for the accuracy of our budget evaluations. The RMS values of both self-consistency and residual can be reduced by $\sim 60\%$ using ERA5 instead of ERA-I. We recommend using the RMS metric in other budget evaluations to objectively measure the quality of results.

In conclusion, the performance of ERA5 is clearly improved compared to its predecessor, but some fields contain unrealistically strong trends and should be used with caution. While the reason for the strong gradual change in the late

TABLE A2. Constants as used in Algorithm A1.

Constant	Value	Name
c_p	1004.7	Specific heat of dry air at constant pressure ($\text{J kg}^{-1} \text{K}^{-1}$)
c_{pv}	1846.1	Specific heat of water vapor at constant pressure ($\text{J kg}^{-1} \text{K}^{-1}$)
c_w	4218.0	Specific heat of liquid water at 273.15 K ($\text{J kg}^{-1} \text{K}^{-1}$)
R_{dry}	287.0	Gas constant of dry air ($\text{J kg}^{-1} \text{K}^{-1}$)
L_v	2.5008×10^6	Latent heat of vaporization (J kg^{-1})
g	9.806 65	Gravitational acceleration (m s^{-2})

TABLE B1. Mean and RMS values of individual moisture and mass budget terms (mm day^{-1}) for the periods 1985–2018 and 2000–18. All fields are truncated at wavenumber 179 and derived from ERA5, unless otherwise specified.

Term	1985–2018		2000–18	
	Mean	RMS	Mean	RMS
$P_{\text{GPCP ocean}}$	2.91	3.45	2.92	3.48
$P_{\text{GPCP land}}$	2.15	2.87	2.16	2.88
$P_{\text{GPCP land (90°N–60°S)}}$	2.31	3.01	2.32	3.02
$P_{\text{GPCP global}}$	2.69	3.29	2.70	3.32
$P_{\text{GPCC land (90°N–60°S)}}$	2.43	3.34	2.45	3.38
P_{ocean}	3.18	3.82	3.25	3.95
P_{land}	2.26	3.22	2.23	3.20
$P_{\text{land (90°N–60°S)}}$	2.44	3.37	2.41	3.35
P_{global}	2.92	3.66	2.95	3.75
$P_{\text{ERA-I ocean}}$	3.21	3.92	3.21	3.96
$P_{\text{ERA-I land}}$	2.23	3.21	2.23	3.29
$P_{\text{ERA-I land (90°N–60°S)}}$	2.41	3.36	2.41	3.45
$P_{\text{ERA-I global}}$	2.92	3.73	2.93	3.78
E_{ocean}	−3.47	3.82	−3.55	3.91
E_{land}	−1.45	1.84	−1.44	1.83
$E_{\text{land (90°N–60°S)}}$	−1.59	1.93	−1.58	1.93
E_{global}	−2.88	3.37	−2.94	3.44
$E_{\text{ERA-I ocean}}$	−3.45	3.82	−3.49	3.88
$E_{\text{ERA-I land}}$	−1.52	1.94	−1.52	1.94
$E_{\text{ERA-I land (90°N–60°S)}}$	−1.67	2.04	−1.67	2.04
$E_{\text{ERA-I global}}$	−2.89	3.39	−2.92	3.43
$(P + E)_{\text{ocean}}$	−0.29	2.40	−0.30	2.52
$(P + E)_{\text{land}}$	0.82	1.80	0.79	1.79
$(P + E)_{\text{global}}$	0.03	2.24	0.02	2.33
$(P + E)_{\text{ERA-I ocean}}$	−0.24	2.41	−0.28	2.48
$(P + E)_{\text{ERA-I land}}$	0.71	1.72	0.71	1.85
$(P + E)_{\text{ERA-I global}}$	0.03	2.23	0.08	2.32
$\text{VIWVD}_{\text{AN ocean}}$	0.30	2.43	0.31	2.53
$\text{VIWVD}_{\text{AN land}}$	−0.74	1.71	−0.76	1.73
$\text{VIWVD}_{\text{AN global}}$	0.00	2.24	0.00	2.33
$\text{VIWVD}_{\text{AN ERA-I ocean}}$	0.25	2.45	0.26	2.52
$\text{VIWVD}_{\text{AN ERA-I land}}$	−0.60	1.83	−0.64	1.96
$\text{VIWVD}_{\text{AN ERA-I global}}$	0.00	2.29	0.00	2.37
$\text{MDIV}_{\text{total ocean}}$	−0.20	2.46	−0.12	2.61
$\text{MDIV}_{\text{total land}}$	0.49	5.53	0.29	5.63
$\text{MDIV}_{\text{total global}}$	0.00	3.63	0.00	3.74
$\text{MDIV}_{\text{dry ocean}}$	−0.50	3.17	−0.43	3.23
$\text{MDIV}_{\text{dry land}}$	1.23	5.76	1.05	5.79
$\text{MDIV}_{\text{dry global}}$	0.00	4.09	0.00	4.14
$\text{RE}_{\text{AN,total ocean}}$	−0.49	3.16	−0.42	3.22
$\text{RE}_{\text{AN,total land}}$	1.31	5.87	1.08	5.88
$\text{RE}_{\text{AN,total global}}$	0.03	4.13	0.02	4.17
$\text{RE}_{\text{AN,moist ocean}}$	0.02	0.20	0.01	0.22
$\text{RE}_{\text{AN,moist land}}$	0.07	0.52	0.03	0.53
$\text{RE}_{\text{AN,moist global}}$	0.03	0.32	0.02	0.34
$\text{RE}_{\text{AN,moist ERA-I ocean}}$	0.00	0.35	−0.02	0.37
$\text{RE}_{\text{AN,moist ERA-I land}}$	0.11	0.88	0.08	0.89
$\text{RE}_{\text{AN,moist ERA-I global}}$	0.03	0.56	0.01	0.57

1990s needs further investigation, the presented fields of $\text{TEDIV}_{\text{dir}}$ are stable from 2000 onward and thus useful for quantitative assessments of large-scale climate variability and trends. In general, climate studies will benefit from the improved temporal stability and higher useful resolution of ERA5 energy

TABLE B2. Mean and RMS values of individual energy budget terms (W m^{-2}) for the periods 1985–2018 and 2000–18. All fields are truncated at wavenumber 179 and derived from ERA5, unless otherwise specified.

Term	1985–2018		2000–18	
	Mean	RMS	Mean	RMS
$\text{TEDIV}_{\text{dir ocean}}$	6.41	51.71	6.81	52.58
$\text{TEDIV}_{\text{dir land}}$	−15.73	56.24	−16.69	55.92
$\text{TEDIV}_{\text{dir global}}$	0.00	53.06	0.00	53.57
$\text{TEDIV}_{\text{dir ERA-I ocean}}$	6.00	53.62	5.96	54.51
$\text{TEDIV}_{\text{dir ERA-I land}}$	−14.69	67.02	−14.60	68.18
$\text{TEDIV}_{\text{dir ERA-I global}}$	−0.00	57.83	−0.00	58.80
$\text{TEDIV}_{\text{ind ocean}}$	3.35	51.27	3.69	52.09
$\text{TEDIV}_{\text{ind land}}$	−18.32	54.64	−19.53	54.38
$\text{TEDIV}_{\text{ind global}}$	−2.93	52.27	−3.03	52.76
$\text{TEDIV}_{\text{ind ERA-I ocean}}$	4.76	53.30	4.87	54.16
$\text{TEDIV}_{\text{ind ERA-I land}}$	−14.68	52.13	−14.60	52.58
$\text{TEDIV}_{\text{ind ERA-I global}}$	−0.87	52.96	−0.77	53.69
$\text{TEDIV}_{\text{dir archive ocean}}$	6.56	52.74	7.27	53.58
$\text{TEDIV}_{\text{dir archive land}}$	−16.08	57.14	−17.83	56.64
$\text{TEDIV}_{\text{dir archive global}}$	0.00	54.05	0.00	54.48
$R_{\text{TOA DEEP-C ocean}}$	8.14	56.40	8.39	56.49
$R_{\text{TOA DEEP-C land}}$	−18.41	55.36	−18.07	55.30
$R_{\text{TOA DEEP-C global}}$	0.45	56.10	0.72	56.08
$R_{\text{TOA param. ocean}}$	8.32	52.23	8.47	52.22
$R_{\text{TOA param. land}}$	−18.64	53.47	−18.36	53.37
$R_{\text{TOA param. global}}$	0.51	52.59	0.70	52.56
Inferred F_{S} ocean	1.71	37.65	1.56	37.98
Inferred F_{S} land	−2.70	15.74	−1.43	15.78
Inferred F_{S} global	0.43	32.84	0.69	33.11
$\text{RE}_{\text{FC ocean}}$	−3.06	5.76	−3.11	6.08
$\text{RE}_{\text{FC land}}$	−2.59	13.00	−2.84	13.09
$\text{RE}_{\text{FC global}}$	−2.93	8.52	−3.03	8.71
$\text{RE}_{\text{FC ERA-I ocean}}$	−1.24	14.91	−1.09	15.55
$\text{RE}_{\text{FC ERA-I land}}$	0.01	40.56	0.00	41.59
$\text{RE}_{\text{FC ERA-I global}}$	−0.88	25.19	−0.77	25.94
$\text{RE}_{\text{AN ocean}}$	−5.88	9.04	−3.69	7.95
$\text{RE}_{\text{AN land}}$	−3.71	14.93	−2.37	14.83
$\text{RE}_{\text{AN global}}$	−5.25	11.07	−3.31	10.42
$\text{RE}_{\text{AN ERA-I ocean}}$	−8.33	17.29	−7.24	17.29
$\text{RE}_{\text{AN ERA-I land}}$	−6.65	42.31	−6.49	43.42
$\text{RE}_{\text{AN ERA-I global}}$	−7.84	27.03	−7.02	27.54

budgets. The latter will allow for regionalized studies at higher accuracy. Another area of future work will be a backward-extension of the presented evaluations, which will become possible after the release of ERA5 data back to 1950.

Acknowledgments. JM and MM were financially supported by Austrian Science Funds (FWF) Project P33177. Computations were partly performed with routines provided through openIFS (<https://confluence.ecmwf.int/display/OIFS/OpenIFS+Documentation>). The use of openIFS was permitted by ECMWF. The authors thank Franklin R. Robertson and two anonymous reviewers for their insightful comments that helped to improve the manuscript.

Data availability statement. Data are available from the authors on request and will be accessible via the Copernicus Climate Data Store in the future.

APPENDIX A

Pseudo Code

Algorithm A1 presents a pseudo code for calculating $TEDIV_{dir}$, and Tables A1 and A2 contain a list of corresponding acronyms and constants.

Algorithm A1: Pseudo code to compute $TEDIV_{dir}$.

```

input: Hourly ERA5 data from analysis, spectral model level data in native resolution
output: Monthly means of vertically integrated total energy flux divergence ( $TEDIV_{dir}$ )
:
ReadData (PHIS);
for Month  $\leftarrow$  1 to 12 do
  StepsPerMonth  $\leftarrow$  24 * DaysPerMonth;
  for Step  $\leftarrow$  1 to StepsPerMonth do
    ReadData (VORT);
    ReadData (DIV);
    ReadData (Q);
    ReadData (T);
    ReadData (LNPS);
    ReadData (VIQ);
    Transform all fields to a quadratic Gaussian grid F480;
    UV  $\leftarrow$  Compute horizontal wind components from VORT and DIV;
    MASSTEND  $\leftarrow$  Compute vert. integrated mass tendency from exp (LNPS);
    QTEND  $\leftarrow$  (VIQ(Step+1) - VIQ(Step-1)) / (2 $\Delta t$ );
    niter  $\leftarrow$  Define number of correction steps to iteratively adjust the wind field;
    for CorrectionStep  $\leftarrow$  1 to niter do
      MASSDIV  $\leftarrow$  Compute vert. integrated mass flux divergence from UV;
      VIWVDAN AN  $\leftarrow$  Compute vert. int. moisture flux divergence from UN and Q;
      DIVERR  $\leftarrow$  MASSDIV + MASSTEND - QTEND - VIWVDAN AN;
      UVERR  $\leftarrow$  Compute two-dimensional wind field adjustment by 1.) inverting the Laplacian of DIVERR, 2.)
      take the gradient, and 3.) divide by exp (LNPS) / g
      foreach atmospheric level in UV do UV  $\leftarrow$  UV - VERR;
    end
     $T_c \leftarrow T - 273.15$ ;
    EKIN  $\leftarrow$  (UV2) / 2;
    EPOT  $\leftarrow$  Compute geopotential according to IFS documentation Part III using PHIS, Q, T and exp (LNPS);
    ELAT  $\leftarrow$  [  $L_v + (c_{pv} - c_w) * T_c$  ] * Q;
    EENT  $\leftarrow$   $c_p * T_c * (1 - Q)$ ;
    TEFLUX  $\leftarrow$  Compute vertical integral of; [ (ELAT + EKIN + EPOT + EENT) * UV ] ;
     $TEDIV_{dir}$  (Step)  $\leftarrow$  Compute divergence of TEFLUX;
  end
  Average over all  $TEDIV_{dir}$  (Step) and store as monthly mean;
end

```

APPENDIX B

Additional Tables

Tables B1 and B2 provide mean and RMS values of individual energy, moisture, and mass budget terms for the periods 1985–2018 and 2000–18.

REFERENCES

- Adler, R. F., G. Gu, and G. J. Huffman, 2012: Estimating climatological bias errors for the Global Precipitation Climatology Project (GPCP). *J. Appl. Meteor. Climatol.*, **51**, 84–99, <https://doi.org/10.1175/JAMC-D-11-052.1>.
- , and Coauthors, 2018: The Global Precipitation Climatology Project (GPCP) monthly analysis (new version 2.3) and a review of 2017 global precipitation. *Atmosphere*, **9**, 138, <https://doi.org/10.3390/atmos9040138>.
- Alexandersson, H., and A. Moberg, 1997: Homogenization of Swedish temperature data. Part I: Homogeneity test for linear trends. *Int. J. Climatol.*, **17**, 25–34, [https://doi.org/10.1002/\(SICI\)1097-0088\(199701\)17:1<25::AID-JOC103>3.0.CO;2-J](https://doi.org/10.1002/(SICI)1097-0088(199701)17:1<25::AID-JOC103>3.0.CO;2-J).
- Becker, A., P. Finger, A. Meyer-Christoffer, B. Rudolf, K. Schamm, U. Schneider, and M. Ziese, 2013: A description of the global

- land-surface precipitation data products of the Global Precipitation Climatology Centre with sample applications including centennial (trend) analysis from 1901–present. *Earth Syst. Sci. Data*, **5**, 71–99, <https://doi.org/10.5194/essd-5-71-2013>.
- Berrisford, P., P. Kållberg, S. Kobayashi, D. Dee, S. Uppala, A. J. Simmons, P. Poli, and H. Sato, 2011: Atmospheric conservation properties in ERA-Interim. *Quart. J. Roy. Meteor. Soc.*, **137**, 1381–1399, <https://doi.org/10.1002/qj.864>.
- Boyer, T. P., S. Levitus, J. I. Antonov, R. A. Locarnini, and H. E. Garcia, 2005: Linear trends in salinity for the World Ocean, 1955–1998. *Geophys. Res. Lett.*, **32**, L01604, <https://doi.org/10.1029/2004GL021791>.
- Buizza, R., and Coauthors, 2018: Advancing global and regional reanalyses. *Bull. Amer. Meteor. Soc.*, **99**, ES139–ES144, <https://doi.org/10.1175/BAMS-D-17-0312.1>.
- Cheng, L., and Coauthors, 2020: Improved estimates of changes in upper ocean salinity and the hydrological cycle. *J. Climate*, **33**, 10 357–10 381, <https://doi.org/10.1175/JCLI-D-20-0366.1>.
- Chiodo, G., and L. Haimberger, 2010: Interannual changes in mass consistent energy budgets from ERA-Interim and satellite data. *J. Geophys. Res.*, **115**, D02112, <https://doi.org/10.1029/2009JD012049>.
- Collins, M., and Coauthors, 2013: Long-term climate change: Projections, commitments and irreversibility. *Climate Change 2013: The Physical Science Basis*, T. F. Stocker et al., Eds., Cambridge University Press, 1029–1136.
- Dai, A., and K. E. Trenberth, 2002: Estimates of freshwater discharge from continents: Latitudinal and seasonal variations. *J. Hydrometeorol.*, **3**, 660–687, [https://doi.org/10.1175/1525-7541\(2002\)003<0660:EOFDfC>2.0.CO;2](https://doi.org/10.1175/1525-7541(2002)003<0660:EOFDfC>2.0.CO;2).
- , T. Qian, K. E. Trenberth, and J. D. Milliman, 2009: Changes in continental freshwater discharge from 1948 to 2004. *J. Climate*, **22**, 2773–2792, <https://doi.org/10.1175/2008JCLI2592.1>.
- Dee, D. P., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, <https://doi.org/10.1002/qj.828>.
- , M. Balmaseda, G. Balsamo, R. Engelen, A. J. Simmons, and J.-N. Thépaut, 2014: Toward a consistent reanalysis of the climate system. *Bull. Amer. Meteor. Soc.*, **95**, 1235–1248, <https://doi.org/10.1175/BAMS-D-13-00043.1>.
- Durran, D., 2013: *Numerical Methods for Wave Equations in Geophysical Fluid Dynamics*. Springer, 466 pp.
- Fasullo, J. T., and K. E. Trenberth, 2008: The annual cycle of the energy budget. Part I: Global mean and land–ocean exchanges. *J. Climate*, **21**, 2297–2312, <https://doi.org/10.1175/2007JCLI1935.1>.
- Guttenstein, M., K. Fennig, M. Schröder, T. Trent, S. Bakan, J. B. Roberts, and F. R. Robertson, 2021: Intercomparison of freshwater fluxes over ocean and investigations into water budget closure. *Hydrol. Earth Syst. Sci.*, **25**, 121–146, <https://doi.org/10.5194/HESS-25-121-2021>.
- Hantel, M., and L. Haimberger, 2016: *Grundkurs Klima*. Springer, 418 pp.
- Hartmann, D., and Coauthors, 2013: Observations: Atmosphere and surface. *Climate Change 2013: The Physical Science Basis*, T. F. Stocker et al., Eds., Cambridge University Press, 159–254, <https://doi.org/10.1017/CBO9781107415324.008>.
- Held, I. M., and B. J. Soden, 2006: Robust responses of the hydrological cycle to global warming. *J. Climate*, **19**, 5686–5699, <https://doi.org/10.1175/JCLI3990.1>.
- Hersbach, H., and Coauthors, 2018: Operational global reanalysis: Progress, future directions and synergies with NWP. ERA Rep. Series 27, 65 pp., <https://doi.org/10.21957/tkic6g3wm>.
- , and Coauthors, 2020: The ERA5 global reanalysis. *Quart. J. Roy. Meteor. Soc.*, **146**, 1999–2049, <https://doi.org/10.1002/qj.3803>.
- Hyder, P., and Coauthors, 2018: Critical Southern Ocean climate model biases traced to atmospheric model cloud errors. *Nat. Commun.*, **9**, 3625, <https://doi.org/10.1038/s41467-018-05634-2>.
- L'Ecuyer, T. S., and Coauthors, 2015: The observed state of the energy budget in the early twenty-first century. *J. Climate*, **28**, 8319–8346, <https://doi.org/10.1175/JCLI-D-14-00556.1>.
- Li, G., L. Cheng, J. Zhu, K. Trenberth, M. Mann, and J. Abraham, 2020: Increasing ocean stratification over the past half-century. *Nat. Climate Change*, **10**, 1116–1123, <https://doi.org/10.1038/s41558-020-00918-2>.
- Liang, M., A. Czaja, R. Graversen, and R. Tailleux, 2018: Poleward energy transport: Is the standard definition physically relevant at all time scales? *Climate Dyn.*, **50**, 1785–1797, <https://doi.org/10.1007/s00382-017-3722-x>.
- Liu, C., and Coauthors, 2015: Combining satellite observations and reanalysis energy transports to estimate global net surface energy fluxes 1985–2012. *J. Geophys. Res. Atmos.*, **120**, 9374–9389, <https://doi.org/10.1002/2015JD023264>.
- , and Coauthors, 2017: Evaluation of satellite and reanalysis-based global net surface energy flux and uncertainty estimates. *J. Geophys. Res. Atmos.*, **122**, 6250–6272, <https://doi.org/10.1002/2017JD026616>.
- , and Coauthors, 2020: Variability in the global energy budget and transports 1985–2017. *Climate Dyn.*, **55**, 3381–3396, <https://doi.org/10.1007/s00382-020-05451-8>.
- Llovel, W., M. Becker, A. Cazenave, J.-F. Crétaux, and G. Ramillien, 2010: Global land water storage change from grace over 2002–2009; inference on sea level. *C. R. Geosci.*, **342**, 179–188, <https://doi.org/10.1016/j.crte.2009.12.004>.
- Loeb, N. G., B. A. Wielicki, D. R. Doelling, G. L. Smith, D. F. Keyes, S. Kato, N. Manalo-Smith, and T. Wong, 2009: Toward optimal closure of the earth's top-of-atmosphere radiation budget. *J. Climate*, **22**, 748–766, <https://doi.org/10.1175/2008JCLI2637.1>.
- , and Coauthors, 2018: Clouds and the Earth's Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) Top-of-Atmosphere (TOA) edition-4.0 data product. *J. Climate*, **31**, 895–918, <https://doi.org/10.1175/JCLI-D-17-0208.1>.
- Mayer, M., and L. Haimberger, 2012: Poleward atmospheric energy transports and their variability as evaluated from ECMWF reanalysis data. *J. Climate*, **25**, 734–752, <https://doi.org/10.1175/JCLI-D-11-00202.1>.
- , —, M. Pietschnig, and A. Storto, 2016: Facets of Arctic energy accumulation based on observations and reanalyses 2000–2015. *Geophys. Res. Lett.*, **43**, 10 420–10 429, <https://doi.org/10.1002/2016GL070557>.
- , —, J. M. Edwards, and P. Hyder, 2017: Toward consistent diagnostics of the coupled atmosphere and ocean energy budgets. *J. Climate*, **30**, 9225–9246, <https://doi.org/10.1175/JCLI-D-17-0137.1>.
- , M. Alonso Balmaseda, and L. Haimberger, 2018: Unprecedented 2015/2016 Indo-Pacific heat transfer speeds up tropical Pacific heat recharge. *Geophys. Res. Lett.*, **45**, 3274–3284, <https://doi.org/10.1002/2018GL077106>.
- , S. Tietsche, L. Haimberger, T. Tsubouchi, J. Mayer, and H. Zuo, 2019: An improved estimate of the coupled Arctic energy budget. *J. Climate*, **32**, 7915–7934, <https://doi.org/10.1175/JCLI-D-19-0233.1>.
- Nogueira, M., 2020: Inter-comparison of ERA-5, ERA-interim and GPCP rainfall over the last 40 years: Process-based analysis of systematic and random differences. *J. Hydrol.*, **583**, 124632, <https://doi.org/10.1016/j.jhydrol.2020.124632>.

- Peixoto, J. P., and A. H. Oort, 1992: *Physics of Climate*. Springer, 520 pp.
- Rhein, M., and Coauthors, 2013: Observations: Ocean. *Climate Change 2013: The Physical Science Basis*, T. F. Stocker et al., Eds., Cambridge University Press, 255–315, <https://doi.org/10.1017/CBO9781107415324.010>.
- Robertson, F. R., M. G. Bosilovich, and J. B. Roberts, 2016: Reconciling land–ocean moisture transport variability in reanalyses with $P - ET$ in observationally driven land surface models. *J. Climate*, **29**, 8625–8646, <https://doi.org/10.1175/JCLI-D-16-0379.1>.
- , and Coauthors, 2020: Uncertainties in ocean latent heat flux variations over recent decades in satellite-based estimates and reduced observation reanalyses. *J. Climate*, **33**, 8415–8437, <https://doi.org/10.1175/JCLI-D-19-0954.1>.
- Schneider, U., A. Becker, P. Finger, A. Meyer-Christoffer, M. Ziese, and B. Rudolf, 2014: GPCC's new land surface precipitation climatology based on quality-controlled in situ data and its role in quantifying the global water cycle. *Theor. Appl. Climatol.*, **115**, 15–40, <https://doi.org/10.1007/s00704-013-0860-x>.
- , —, —, —, and —, 2016: GPCC full data monthly product version 2018 at 1.0°: Monthly land-surface precipitation from rain-gauges built on GTS-based and historical data. Deutscher Wetterdienst, https://doi.org/10.5676/DWD_GPCC/FD_M_V2018_100.
- Seager, R., and I. R. Simpson, 2016: Western boundary currents and climate change. *J. Geophys. Res. Oceans*, **121**, 7212–7214, <https://doi.org/10.1002/2016JC012156>.
- , N. Naik, and G. A. Vecchi, 2010: Thermodynamic and dynamic mechanisms for large-scale changes in the hydrological cycle in response to global warming. *J. Climate*, **23**, 4651–4668, <https://doi.org/10.1175/2010JCLI3655.1>.
- Simmons, A., and Coauthors, 2020: Global stratospheric temperature bias and other stratospheric aspects of ERA5 and ERA5.1. ECMWF Tech. Memo. 859, 40 pp., <https://doi.org/10.21957/rcxqfmg0>.
- Solomon, S., and Coauthors, 2007: Technical summary. *Climate Change 2007: The Physical Science Basis*, S. Solomon et al., Eds., Cambridge University Press, 19–91.
- Stommel, H., 1948: The westward intensification of wind-driven ocean currents. *Eos, Trans. Amer. Geophys. Union*, **29**, 202–206, <https://doi.org/10.1029/TR029i002p00202>.
- Su, L., C. Miao, D. Kong, Q. Duan, X. Lei, Q. Hou, and H. Li, 2018: Long-term trends in global river flow and the causal relationships between river flow and ocean signals. *J. Hydrol.*, **563**, 818–833, <https://doi.org/10.1016/j.jhydrol.2018.06.058>.
- Sverdrup, H. U., 1947: Wind-driven currents in a baroclinic ocean; with application to the equatorial currents of the Eastern Pacific. *Proc. Natl. Acad. Sci. USA*, **33**, 318–326, <https://doi.org/10.1073/pnas.33.11.318>.
- Trenberth, K. E., 1991: Climate diagnostics from global analyses: Conservation of mass in ECMWF analyses. *J. Climate*, **4**, 707–722, [https://doi.org/10.1175/1520-0442\(1991\)004<0707:CDFGAC>2.0.CO;2](https://doi.org/10.1175/1520-0442(1991)004<0707:CDFGAC>2.0.CO;2).
- , and J. Fasullo, 2008: An observational estimate of inferred ocean energy divergence. *J. Phys. Oceanogr.*, **38**, 984–999, <https://doi.org/10.1175/2007JPO3833.1>.
- , and J. T. Fasullo, 2013: Regional energy and water cycles: Transports from ocean to land. *J. Climate*, **26**, 7837–7851, <https://doi.org/10.1175/JCLI-D-13-00008.1>.
- , and —, 2017: Atlantic meridional heat transports computed from balancing Earth's energy locally. *Geophys. Res. Lett.*, **44**, 1919–1927, <https://doi.org/10.1002/2016GL072475>.
- , and —, 2018: Applications of an updated atmospheric energetics formulation. *J. Climate*, **31**, 6263–6279, <https://doi.org/10.1175/JCLI-D-17-0838.1>.
- , and Coauthors, 2007: Observations: Surface and atmospheric climate change. *Climate Change 2007: The Physical Science Basis*, S. Solomon et al., Eds., Cambridge University Press, 235–336.
- , J. T. Fasullo, and J. Mackaro, 2011: Atmospheric moisture transports from ocean to land and global energy flows in reanalyses. *J. Climate*, **24**, 4907–4924, <https://doi.org/10.1175/2011JCLI4171.1>.
- von Schuckmann, K., and Coauthors, 2020: Heat stored in the Earth system: Where does the energy go? The GCOS Earth heat inventory team. *Earth Syst. Sci. Data*, **12**, 2031–2041, <https://doi.org/10.5194/essd-2019-255>.
- Wild, M., D. Folini, C. Schär, N. Loeb, E. Dutton, and G. König-Langlo, 2012: The global energy balance from a surface perspective. *Climate Dyn.*, **40**, 3107–3134, <https://doi.org/10.1007/S00382-012-1569-8>.
- Willmott, C. J., C. M. Rowe, and W. D. Philpot, 1985: Small-scale climate maps: A sensitivity analysis of some common assumptions associated with grid-point interpolation and contouring. *Amer. Cartogr.*, **12**, 5–16, <https://doi.org/10.1559/152304085783914686>.