

**The annual cycle of the energy budget:**  
**Global mean and land-ocean exchanges**

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## Abstract

The mean and annual cycle of energy flowing into the climate system and its storage, release, and transport in the atmosphere, ocean, and land surface are estimated with recent observations. An emphasis is placed on establishing internally consistent quantitative estimates with discussion and assessment of uncertainty. At the top-of-atmosphere (TOA), adjusted radiances from Earth Radiation Budget Experiment (ERBE) and Clouds and the Earth's Radiant Energy System (CERES) are used, while in the atmosphere National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis and European Centre for Medium Range Weather Forecasts (ECMWF) reanalysis (ERA-40) estimates are used. The net upward surface flux ( $F_S$ ) over ocean is derived as the residual of TOA and atmospheric energy budgets, and is compared with direct calculations of ocean heat content ( $O_E$ ) and its tendency ( $\delta O_E/\delta t$ ) from several ocean temperature datasets. Over land  $F_S$  from a stand-alone simulation of the Community Land Model forced by observed fields is used. A depiction of the full energy budget based on ERBE fluxes from 1985 to 1989 and CERES fluxes from 2000 to 2004 is constructed that matches estimates of the global, global-ocean, and global-land imbalances. In addition, the annual cycle of the energy budget during both periods is examined and compared with ocean heat content changes.

The near balance between net TOA radiation ( $R_T$ ) and  $F_S$  over ocean and thus with  $O_E$ , and between  $R_T$  and atmospheric total energy divergence over land, are documented both in the mean and for the annual cycle. However, there is an annual mean transport of energy by the atmosphere from ocean to land regions of  $2.2 \pm 0.1$  PW ( $10^{15}$  watts) primarily in the northern winter when the transport exceeds 5 PW. The global albedo is dominated by a semiannual cycle

over the oceans, but combines with the large annual cycle in solar insolation to produce a peak in absorbed solar and net radiation in February, somewhat after the perihelion, and with the net radiation 4.3 PW higher than the annual mean, as it is enhanced by the annual cycle of outgoing long-wave radiation that is dominated by land regions. *In situ* estimates of the annual variation of  $O_E$  are found to be unrealistically large. Challenges in diagnosing interannual variability in the energy budget and its relationship to climate change are identified in the context of the episodic and inconsistent nature of observations.

## 1. Introduction

The primary driver of Earth's climate system is the uneven distribution of net downward radiation ( $R_T$ ) at top-of-atmosphere (TOA) owing principally to sun-Earth geometry. Upon entering the system, the incoming radiative flux is partitioned among internal heat, and potential, latent, and kinetic energy (Trenberth and Stepaniak 2004). While basic aspects of the global-mean TOA budget, such as the near balance between absorbed solar radiation ( $ASR$ ) and outgoing longwave radiation ( $OLR$ ) have long been known (Dines 1917), the quantification of the flow and storage of energy, and the conversions among different forms of energy, has evolved continually. Moreover, the full energy budget, which is comprised of multiple flux, transport, and storage terms, has been assessed only rarely.

Kiehl and Trenberth (1997) provided an assessment of the global annual mean budget at TOA and for the atmosphere based on a mix of observations and model results. Major recent advances in understanding the energy budget have been provided by satellite data and globally gridded reanalyses (e.g., Trenberth et al. 2001, Trenberth and Stepaniak 2003a,b). Trenberth et al. (2001) performed comprehensive estimates of the atmospheric energy budget based on two first generation atmospheric reanalyses and several surface flux estimates, and made crude estimates of uncertainty. Trenberth and Caron (2001) inferred the net surface fluxes as a residual of TOA satellite measurements and the atmospheric energy budget, and computed meridional ocean heat transports as a residual. Although physical constraints on the requirement for a global mean energy balance adjusted for effects of climate change were used, no error analysis of the satellite data was included. However, inferred annual mean meridional energy transports in the ocean agreed with direct estimates to within error bars of order 0.3 PW (1 petawatt is  $10^{15}$  watts)

(Trenberth and Caron 2001). The atmospheric energy budget has been documented in some detail for the annual cycle (Trenberth and Stepaniak 2003a, 2004) and for El Niño–Southern Oscillation (ENSO) and interannual variability (Trenberth et al. 2002; Trenberth and Stepaniak 2003a). Challenges remain, however, and in this paper new observations, improved estimates of the current TOA and surface imbalances, and a more holistic approach allow for improvements in the budget to be made and for uncertainty estimates to be constructed. Also addressed here is quantification of the mean annual cycle and land-ocean exchanges, including a decomposition of the energy flows for the ocean and land domains, as this allows the constraints of the two systems to be exploited. Moreover, a comparison of the residuals from TOA and atmospheric budgets at the surface with independent measurements of integrated ocean heat content is presented, and thus surface fluxes are better quantified.

A schematic of the flows of energy and notation is given in Fig. 1 and acronyms are defined in Appendix II. At TOA, the dominant energy terms include solar insolation ( $SI$ ),  $ASR$ , and  $OLR$ , with  $R_T = ASR - OLR$ . Upon entering the climate system, energy imbalances at TOA can be partitioned mainly between the divergence and tendency of total atmospheric energy ( $A_E$ ) and oceanic energy ( $O_E$ ), including sea ice, and secondarily by changes in land energy storage ( $L_E$ ), and energy transport by rivers. In the atmosphere the transport of energy ( $\mathbf{F}_A$ ) and its divergence  $\nabla \cdot \mathbf{F}_A$  dominates the tendency locally. Although the atmosphere represents only a fraction of the climate system’s heat capacity (equivalent to 3.5 m of ocean, Trenberth and Stepaniak 2004), it is the dominant contributor to the mean poleward transport of energy outside of the Tropics (Trenberth and Caron 2001). While globally over long time scales,  $R_T$  is small, its annual cycle is considerable owing principally to fluctuations in Earth’s albedo and  $SI$ , as Earth orbits through

its perihelion (January 3) and aphelion (July 3). The associated annual cycle of global-mean  $R_T$  is of order  $20 \text{ W m}^{-2}$  from peak to trough or, integrated globally, about 10 PW (e.g., Kyle et al. 1993). This is large compared with the estimated total TOA imbalance arising from changing atmospheric composition, which Hansen et al. (2005) estimate recently at  $0.85 \pm 0.15 \text{ W m}^{-2}$ . On monthly timescales, the regional variability of total atmospheric energy tendency ( $\delta A_E/\delta t$ ) and divergent energy flux ( $\nabla \cdot \mathbf{F}_A$ ) associated with  $R_T$  are considerable (Trenberth and Stepaniak 2003a). Stationary and transient components of atmospheric dry static, latent, and kinetic energy transports combine and, in instances mutually compensate, to result in a seamless poleward transport of energy that belies the complexity of the physical mechanisms of which it is comprised (Trenberth and Stepaniak 2003a,b). The TOA and atmospheric budgets are closed by the net upward surface energy flux ( $F_S$ ), which consists principally of shortwave and longwave radiative fluxes and turbulent fluxes of latent and sensible heat. In the ocean,  $F_S$  in the Tropics is balanced principally by the transport of ocean energy (mainly heat) ( $\mathbf{F}_O$ ) and its divergence ( $\nabla \cdot \mathbf{F}_O$ ) while in mid-latitudes it is largely balanced locally by changes in ocean heat storage  $\delta O_E/\delta t$  (Jayne and Marotzke 2001, Trenberth et al. 2001).

Estimation of  $F_S$  is particularly challenging, as *in situ* estimates contain considerable systematic and random uncertainties that frequently exceed  $25 \text{ W m}^{-2}$  (Gleckler and Weare 1997; Yu et al. 1999). If even a small component of the uncertainty (e.g., 10% or  $2.5 \text{ W m}^{-2}$ ) is systematic, the associated global budget uncertainty (1.3 PW) can obscure key characteristics of the TOA and surface budgets. Estimates of  $F_S$  from reanalyses are model derived and contain errors (Trenberth et al. 2001, 2002). Perhaps most troublesome is that the errors are both random and systematic, and the data thus fail to satisfy basic global constraints.

An alternative is based on deriving  $F_S$  over ocean from the residual of TOA and reanalysis atmospheric budgets (Trenberth 1997). While the regional error of this method is of similar magnitude to that of in situ estimates, the error converges to be the largely random error in TOA estimates because the systematic TOA bias is first removed, global mean  $\nabla \cdot \mathbf{F}_A$  is zero by construction, and  $\delta A_E / \delta t$  is small generally. With climate change, a small imbalance in  $F_S$  exists and there is an order 0.5 PW net TOA heat flux (Hansen et al. 2005), which can be further separated into its land (<0.1 PW, Huang 2006) and ocean (~0.4 PW, Willis et al. 2004) components. These present day estimates compare favorably in relative importance to those of Levitus et al. (2005), who estimated contributions of the imbalance from 1955 to 1998 to the oceans (0.10 PW) and continental landmass and glaciers (0.01 PW). An open science question addressed here is whether the atmospheric reanalyses, in conjunction with available estimates of  $R_T$ , are able to attain accuracy in  $\nabla \cdot \mathbf{F}_A$ , and thus in  $F_S$ , that is suitable for resolving simultaneously the global, global-ocean, and global-land annual cycles and net imbalances from this residual method.

The purposes of the present manuscript are therefore: 1) to explore the current range of observations of the global, global-land, and global-ocean energy budgets and their annual cycle, 2) to develop estimates of uncertainty for each term, 3) to establish a comprehensive best-estimate of the present day budget. The term “best” is used in the sense of “best practice<sup>2</sup>” in recognition of the fact that successive data and analyses are likely to improve upon the results

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<sup>2</sup> Best Practice is a management idea that asserts a technique, method, or process is more effective at delivering a particular outcome than any other.

presented here. Section 2 describes the data and methods while Section 3 documents aspects of the mean energy cycle, both for the globe and for global-land and global-ocean domains. The results are presented in Section 4 followed by discussion (Section 5) and concluding remarks (Section 6). This paper deals with the global, land and ocean domain averages, but it is essential to delve into the full spatial structure to determine uncertainties and sources of error. The meridional structure and poleward transports are therefore documented in Fasullo and Trenberth (2007) and the full three dimensional aspects for the ocean are presented in Trenberth and Fasullo (2007), where a more complete error analysis is performed.

## **2. Data and methods**

In dealing with energy and its partitioning, we have found that it is important to ensure conservation by dealing with actual amounts, generally expressed in terms of petawatts rather than watts per square meter ( $\text{W m}^{-2}$ ). Moreover in communicating the relative importance of the various energy terms – as a function of latitude for example – PW is the preferred unit as this integrated quantity is more directly relevant to: 1) understanding the balances and closures that exist in the budget, and, 2) portraying realistically the relative magnitude of the multiple energy terms. In instances, area-average quantities are used for reference to earlier works. The conversion globally is  $1 \text{ W m}^{-2} = 0.510 \text{ PW}$ , over the global-ocean  $1 \text{ W m}^{-2} = 0.372 \text{ PW}$ , and over the global-land  $1 \text{ W m}^{-2} = 0.138 \text{ PW}$ .

### *a. Energy budgets*

The atmospheric energy budget equation, when vertically integrated (Trenberth 1997), can be written as



$$\frac{\partial A_E}{\partial t} + \nabla \cdot \mathbf{F}_A = ASR - OLR + F_s \quad (1)$$

$$\text{where } \mathbf{F}_A = \frac{1}{g} \int_0^{p_s} \mathbf{v} (c_p T + \ddot{O} + k + Lq) dp \text{ and } A_E = \frac{1}{g} \int_0^{p_s} (c_p T + \ddot{O}_s + k + Lq) dp.$$

$\mathbf{F}_A$  is the total vertically integrated atmospheric energy transport. The fields  $k$ ,  $T$ ,  $\mathbf{v}$ ,  $q$ ,  $p$ , and  $z$  denote the atmospheric kinetic energy, temperature, wind, specific humidity, pressure, and geopotential height, respectively, and  $\Phi = gz$  is the geopotential. Standardized values  $a$ ,  $g$ ,  $c_p$ , and  $L$ , correspond to Earth's radius, the acceleration due to gravity, the specific heat of air at constant pressure, and the latent heat of vaporization. The subscript  $s$  refers to the surface. Note that the atmospheric energy includes a surface geopotential term and the quantity in brackets differs from the total energy that is transported owing to the pressure-work term in the thermodynamic equation. Consequently, the terms in  $\mathbf{F}_A$  consist of the transport of the sensible heat, potential energy, kinetic energy and latent energy, and the first two can be combined as the dry static energy which, combined with the last term,  $Lq$ , give the moist static energy.  $R_T = ASR - OLR$ , and  $F_s$  have been previously defined. By definition the divergent term in (1) drops out for global mean budgets while for the full four-dimensional fields, both the meridional and zonal energy transports must be considered.

As previously discussed,  $F_s$  over ocean is diagnosed from the residual of the atmospheric budget such that

$$F_s = \partial A_E / \partial t + \nabla \cdot \mathbf{F}_A - R_T. \quad (2)$$

We introduce the notation that a superscript corresponds to the domain: g (global), l (land) and o (ocean). As the annual mean of  $F_S^l$  is approximately zero (e.g., Huang 2006), the net transport of energy from ocean to land can be estimated from

$$\overline{F_A(ocean \rightarrow land)} = \overline{R_T^l} + \overline{\frac{\partial A_E^l}{\partial t}} \quad (3)$$

where the overbar corresponds to annual averaging. Additionally, integrated ocean energy ( $O_E$ ) is computed to include only the heat content, as changes in oceanic kinetic energy and geopotential are small, such that,

$$O_E = \int T(z) C_w dz, \quad (4)$$

where  $z$  is here the depth of the ocean,  $T$  is the ocean temperature and  $C_w$  is the specific heat of sea water. The ocean datasets used to diagnose  $T$  include the World Ocean Atlas 2005 (WOA, Locarini et al. 2006), the ocean analysis of the Japanese Meteorological Agency Version 6.2 (JMA, Ishii et al. 2006), and the recently corrected Global Ocean Data Assimilation System (GODAS, Behringer and Xue 2004, Behringer 2006).

An independent estimate of global ocean  $F_S$  can be obtained as a residual from

$$F_S^o = \partial A_E^g / \partial t - R_T^g - F_S^l \quad (5)$$

and for the global ocean

$$\partial O_E / \partial t = -F_S^o. \quad (6)$$

The challenges posed in diagnosing the terms in (1) include achieving adequate sampling of the sizable spatial and temporal gradients associated with radiative fluxes,  $\nabla \cdot \mathbf{F}_A$ , and  $F_S$ , and

diagnosing atmospheric fields accurately, particularly with regard to tropical moisture, geopotential height, and extratropical temperature, whose transports are dominant contributors to the total energy divergence. Satellite measurements provide the “best-estimate” of TOA terms in (3) while the most comprehensive estimate of global atmospheric temperature and moisture, fields are available from reanalyses of the National Centers for Environmental Prediction / NCAR reanalysis (NRA) and the second generation European Centre for Medium Range Weather Forecasts (ECMWF) Reanalysis (ERA-40). Contemporaneous discrepancies between the reanalyses provide an important initial estimate of uncertainty in the atmospheric energy storage and flow.  $F_S^I$  averaged over long time periods is small, but its annual cycle, which is associated primarily with energy storage, is non-negligible (Dai and Trenberth 2002). As direct observations of  $F_S^I$  are few, model simulations from the CLM forced by observed fields are used (Qian et al. 2006), as described below.

#### *b. Satellite retrievals*

Satellite retrievals from the ERBE (Barkstrom and Hall 1982) and the CERES (Wielicki et al. 1996) datasets are used. ERBE estimates are based on observations from three satellites (ERBS, NOAA-9, and NOAA-10) for February 1985 to April 1989 and the details of the data and its adjustments are given in Appendix I. While agreeing with available estimates of the global radiative imbalance, our ERBE adjusted product also compensates for the discontinuity in inferred ocean-to-land energy transport from (3) that results from the loss of NOAA-9 and matches the OLR trend from 60°S to 60°N as reported by the ERBS satellite during the ERBE period.

The CERES instruments used here (FM1 and FM2) are flown aboard the Terra satellite, which has a morning equatorial crossing time and was launched in December 1999 with data extending to May 2004. We also make some use of CERES measurements on Aqua (from instruments FM3 and FM4), which had an afternoon equatorial crossing time, but the data are limited as it was launched in May 2002. CERES benefits from improved angular distribution models and scene identification relative to ERBE, owing in part to its enhanced scanner resolution and has achieved unprecedented instrument stability and calibration accuracy (Loeb et al. 2000, 2007). The CERES TOASRB FM1 Terra-MODIS Edition 2D Rev 1 product used here is reported on a  $1^\circ$  grid and is hereafter referred to as TOASRB. For more direct comparison to ERBE fluxes, “ERBE-like” fields CERES are also used and are based on upscaled radiances and ERBE algorithms applied to single satellite retrievals. In interpreting the differences between ERBE and CERES fields it is important to recognize the different time periods, retrieval methodologies, and retrieval instruments used. Among the methodological differences, the use of less accurate angular distribution models and crude cloud and surface scene identification in the ERBE retrieval process are fundamental (Ye and Coakley 1996). Key among structural differences is the coarser footprint and improved multi-satellite sampling of ERBE retrievals. Hence, it is not possible to isolate natural variability from direct ERBE/CERES comparisons, even through the use of CERES “ERBE-like” products. Instances of missing data in the CERES datasets are treated in the same manner as for ERBE in Trenberth (1997) with a cosine fit to the annual cycle in albedo used to infer reflected solar radiation for regions of missing data. Typically this accounts for most of the variance but it is undesirable to add a second harmonic as it raises the possibility that the albedo will exceed unity and become unphysical.

A comprehensive error analysis of the CERES mean budget (Wielicki et al. 2006), summarized in Table 1, addresses the difference between current estimates of the TOA imbalance (Hansen et al. 2005; Willis et al. 2004; Huang 2006) and the global mean imbalance of  $6.4 \text{ W m}^{-2}$  reported by TOASRB. While each error source is likely a very small percentage of the retrieved flux, the combined uncertainties can be significant. Indeed it is evident that if these errors stand up to further scrutiny, then they are not random but must all have the sign to enable them to add up to the observed physically unrealistic imbalance. Key among the error sources ( $>0.5 \text{ W m}^{-2}$ ) are Total Solar Irradiance and absolute calibration. Multiple secondary sources of error ( $<0.5 \text{ W m}^{-2}$ ) relate to sampling issues and assumptions of satellite altitude and near-terminator fluxes.

To achieve an estimate of TOA fluxes that matches current estimates of global imbalances (Hansen et al. 2005; Willis et al. 2004; Huang 2006), we therefore make a number of adjustments to the CERES TOASRB retrievals. An upper bound on the longwave adjustment per Table 1 is  $1.5 \text{ W m}^{-2}$  and, as the bulk of the uncertainty relates to absolute calibration, OLR is therefore increased uniformly by this amount in constructing a “best-estimate”. To address the remaining error, we scale albedo rather than ASR directly as per Trenberth (1997). The net imbalance in  $R_T$  is reduced to an acceptable but imposed  $0.5 \text{ PW}$  (about  $1 \text{ W m}^{-2}$ ) (a scaling of  $+0.303\%$  for FM1 and  $+0.308\%$  for FM2). As no explicit tuning is done in the CERES retrieval process itself, the adjustments made here are thus the only empirically-derived corrections applied to the CERES fields. Future CERES products may well improve these estimates.

In addition to the uncertainties listed in Table 1, the estimation of climatological mean fluxes from observations spanning a limited number of years is subject to errors associated with

interannual variability, which has been estimated at 0.25 PW ( $0.5 \text{ W m}^{-2}$ ) for TOA fluxes (Ardanuy et al. 1992). Associated uncertainty estimates, equal to two sample standard deviations of interannual variability ( $2\sigma_I$ ), are reported. Moreover, the impacts of diurnal aliasing, which has already been provided by Wielicki et al. (2006) for climatological fluxes (Table 1), has yet to be fully documented for annual-cycle aspects of the energy budget. To quantify these impacts, mean fluxes from the ERBE-like Terra (FM1+FM2) retrievals are contrasted with those of the Aqua (FM3+FM4) satellite. Their difference is reported as  $\Delta_D$  in Tables 4 through 6. As differences in mean fluxes between the sensors relate primarily to satellite calibration, and thus preclude a meaningful evaluation of the impact of diurnal sampling on mean fluxes, only fields relating to variability across the annual cycle are evaluated for  $\Delta_D$ , as for such calculations calibration biases are largely subtracted out by removing the annual mean.

### *c. Reanalysis datasets*

While some previous studies (Yu et al. 1999) rely on reanalysis fields that are purely model derived, here we use only fields strongly influenced by observations, such as surface pressure, winds, atmospheric temperature and humidity (Kalnay et al. 1996). We use the NRA (Kistler et al. 2005) and the second generation ERA-40 reanalyses (Uppala et al. 2005). Trenberth et al. (2001) described how all of the energy terms for the atmosphere, their transports and divergences are computed, and compared results for NRA and the first generation ECMWF reanalysis ERA-15. These have now also been computed for ERA-40 and updated for NRA [all data are available from <http://www.cgd.ucar.edu/cas/catalog>]. Precipitation minus evaporation is computed from the atmospheric moisture budget rather than by differencing the model derived fields of precipitation and evaporation (Trenberth and Guillemot 1998; Trenberth et al 2001).

Nonetheless, there are known problems with both NRA and ERA-40 reanalyses (e.g., Trenberth et al. 2005; Uppala et al. 2005) and indeed it is a goal of the present study to further explore the limits of the reanalyses in assessing the energy budget.

#### *d. Land model integrations*

The seasonal variation of  $F_S^l$  is taken from a stand-alone integration of the CLM (Bonan et al. 2002; Qian et al. 2006). The CLM represents the surface with five primary subgrid land cover types, 16 plant functional types, and 10 layers for soil temperature and water, with explicit treatment of liquid water and ice. Representation of the seasonal cycle by the CLM shows significant improvements over previous generation models (Bonan et al. 2002). In the stand-alone integration used here, the CLM is forced with observed fields infilled with high frequency information from the NRA where observations are unavailable (Qian et al. 2006). The CLM reports fields on a T42 grid, with 128 longitude and 64 latitude points, on a monthly basis from 1948 through 2004.

#### *e. Ocean surface fluxes and storage*

The accuracy in calculating  $F_S^o$  using (2) is of order  $20 \text{ W m}^{-2}$  over 1000 km scales (Trenberth et al. 2001). Uncertainty in  $F_S$  is governed by the uncertainties in  $R_T$ ,  $\nabla \cdot \mathbf{F}_A$ , and  $\delta A_E / \delta t$ , which on large scales are much less than those associated with explicit modeling of surface radiative and turbulent fluxes. Because  $\nabla \cdot \mathbf{F}_A$  is zero globally by definition, a cancellation of errors must occur over larger scales.

Estimates of ocean temperature are taken from three datasets: the WOA, JMA and GODAS, which were current as of Jan. 2007. The WOA fields are based on climatological monthly mean

estimates of observed data after about 1960, the majority of which occur in the 1980s and 1990s. The estimates are thus not concurrent with either the ERBE or CERES periods and reported differences with  $F_S$  must be considered in the context of  $\sigma_I$ , which can be estimated from the JMA and GODAS fields. The JMA fields report monthly mean values from 1945 to 2005 and thus allow for the estimation of  $O_E$  and  $\sigma_I$ . In contrast to WOA and JMA, GODAS estimates are derived from assimilated observed ocean temperature and salinity with estimates of surface forcing (wind stress, heat flux, E-P) taken from NRA2, SST relaxed to weekly NCEP analysis, and salinity relaxed to the WOA monthly climatology. The forcings are applied to the Modular Ocean Model v3 (MOM3) spanning 1980 through 2005.

To calculate  $O_E$ ,  $\delta O_E/\delta t$ , and  $\nabla \cdot \mathbf{F}_O$  in (2) to (4) from these data, the density and specific heat of ocean water are approximated as  $1026.5 \text{ kg m}^{-3}$  and  $3990 \text{ J kg}^{-1} \text{ K}^{-1}$ , respectively. A recent WOA analysis (Antonov et al. 2004) of heat content assumes the density of seawater  $\rho$  is  $1020 \text{ kg m}^{-3}$  and the specific heat  $C_w$  is  $4187 \text{ J kg}^{-1} \text{ K}^{-1}$  whereas for the typical salinity of the ocean of  $\sim 35$  PSU,  $\rho$  is  $\sim 1028$  to  $1025 \text{ kg m}^{-3}$  and  $C_w$  is  $3985$  to  $3995 \text{ J kg}^{-1} \text{ K}^{-1}$  for temperatures from 2 to  $20^\circ\text{C}$ . The product  $\rho C_w$  is more nearly constant than either of the two components but the Antonov et al. (2004) value may be 4.4% too high, leading to an overestimate of the  $O_E$  and its annual cycle. We have performed our own integration using the 0, 10, 20, 30, 50, 75, 100, 125, 150, 200, and 250 m standard WOA levels. The depth of each layer is assumed to extend between the midpoints of the indicated depths. In the case of the surface, the layer is assumed to begin at 0 m, and at 250 m depth, the layer is assumed to terminate at 275 m – half way to the next WOA layer at 300 m. The levels above 300 m reported by JMA are identical to those of WOA while the GODAS vertical resolution exceeds that of WOA, thus providing better



resolution of gradients with depth and the thermocline. In computing  $O_E$ , we further edit out the obvious spurious values south of 20°S by accepting only the monthly departures from the annual mean that are within two standard deviations of the zonal mean – and in this way we take advantage of the lack of land and presumed homogeneity over the southern oceans. Globally, the data are filtered temporally by retaining only the first three harmonics of the annual cycle. Oceanic heat tendencies are then computed by reassembling the Fourier series and differencing heat content between the start and end of each month. As the GODAS fields are not global, but rather span 74°S to 65°N, JMA estimates have been used to infill the GODAS fields in the polar regions. The contribution of variability in such filled regions to global  $O_E$  is found to be largely insignificant, owing to both the limited area and seasonal variability of such regions. Our  $O_E$  fields agree better with the Levitus and Antonov (1997) 12-month harmonic values than their total fields. Other small systematic differences may arise from methods of compiling the vertical integral and numerical aspects in computing  $O_E$  (which are not described by Antonov et al. (2004)).

#### *f. Secondary terms*

In addition to the energy budget uncertainties, terms of secondary importance also exist including the heat flux associated with river runoff and other water flows from land areas to ocean, and the latent heat flux associated with seasonal sea-ice variability. Scale analysis can be used to show the net contributions of these terms to be small, however, compared to  $\nabla \cdot \mathbf{F}_A$  and  $F_S$ . For land, net river runoff is approximately  $4 \times 10^4 \text{ km}^3 \text{ yr}^{-1}$  (Trenberth et al. 2007). If the 850 hPa to surface temperature difference ( $\sim 10 \text{ K}$ ) is used to approximate the mean difference between the advected water vapor and river runoff, and a specific heat of  $4218 \text{ J kg}^{-1}$  is assumed, a net

heat flux of 0.05 PW from land to ocean is estimated. For sea-ice, estimates of Arctic sea-ice extent and volume (Koberles and Gerdes 2003) have a seasonal cycle of about  $8 \times 10^3 \text{ km}^3$ . Assuming that ice has a density of  $917 \text{ kg m}^{-3}$  and heat of fusion of  $3.35 \text{ kJ kg}^{-1}$ , the peak heat flux associated with the seasonal cycle is  $\sim 0.5 \text{ PW}$ . However, for the global ocean, the seasonal cycle in the Arctic is balanced by an approximately out-of-phase seasonal cycle in the Antarctic, for which reliable estimates of ice volume seasonality are not presently available. Thus, one would expect some degree of cancellation between Arctic and Antarctic sea-ice variability and the associated global-mean flux is likely to amount to less than a few percent of the observed TOA annual cycle.

#### *g. Regridding and standard deviations*

To provide a consistent delineation of land-sea boundaries among the various datasets described above, all fields are transformed to a grid containing 192 evenly spaced longitudinal grid points and 96 Gaussian-spaced latitudinal grid points using bilinear interpolation (i.e., to a T63 grid;  $1.875^\circ$  latitude). Spatial integrals are calculated using Gaussian weights over the T63 grid and a common land-sea mask is applied. Monthly mean values are used for all calculations. In quantifying annual cycle variability, the estimated population standard deviation  $S$  of monthly mean values is used and so it relates to the amplitude of the seasonality.

### **3. Mean fluxes for ERBE and CERES periods**

Figure 2 summarizes the best-estimate CERES period (March 2000 to May 2004) mean fluxes and  $2\sigma_I$  for the global, global land, and global ocean domains. The corresponding global mean terms for TOASRB, CERES ERBE-like, ERBE-tuned, and ERBE-raw fields are

summarized in Table 2. For  $SI$ , the same total solar irradiance (TSI,  $1365 \text{ W m}^{-2}$ ) is used for all estimates with negligible  $2\sigma_I$ , corresponding to solar insolation of  $174.1 \pm 0.0 \text{ PW}$  globally (but note the systematic uncertainty in Table 1). The TOA  $SI$  flux in all estimates is partitioned into  $45.7 \pm 0.0 \text{ PW}$  for land and  $128.3 \pm 0.0 \text{ PW}$  for ocean. The best-estimate of global mean  $ASR$  is  $122.1 \pm 0.2 \text{ PW}$  and corresponds to a global-mean albedo of  $0.299 \pm 0.001$ . Global-ocean  $ASR$  is  $92.0 \pm 0.2 \text{ PW}$  and global-land  $ASR$  is  $30.0 \pm 0.0 \text{ PW}$ . A global mean best-estimate  $OLR$  flux of  $121.7 \pm 0.4 \text{ PW}$  largely balances  $ASR$  resulting in  $R_T = 0.5 \pm 0.3 \text{ PW}$ . Over ocean,  $R_T$  ( $2.6 \pm 0.2 \text{ PW}$ ) is substantially greater than the global value, and albedo ( $0.283 \pm 0.002$ ) is substantially lower, than for the global mean. An associated imbalance exists between  $ASR$  and  $OLR$  ( $89.5 \pm 0.2 \text{ PW}$ ). Conversely, over land,  $OLR$  of  $32.2 \pm 0.1 \text{ PW}$  exceeds  $ASR$ , and albedo is higher ( $0.344 \pm 0.001$ ) than for the global mean.

A net divergence of energy (Table 3) from ocean to land of  $2.2 \pm 0.1 \text{ PW}$  is implied from the combination of the best-estimate TOA budget, the assumption of small annual tendencies in the atmosphere, and the established negligible mean flux and variability in  $F_S$  over land (Huang 2006). If divergence is calculated explicitly from the reanalyses during the ERBE period, NRA and ERA-40 estimates of the ocean to land flux are  $2.3 \pm 0.2 \text{ PW}$  and  $1.8 \pm 0.8 \text{ PW}$ , respectively (see Fig. 8 presented and discussed later). For the CERES period, during which time the ERA-40 fields are not available, an increased divergence of  $2.7 \pm 0.2 \text{ PW}$  in NRA is found (see Fig. 8).  $F_S^0$  (Fig. 2) is based on values from Willis et al. (2004) for 1996 to 2003. Values from 2003 through 2005 contain considerable uncertainties likely associated with inhomogeneities in the observational record, as data from new ARGO floats have become available, so that the *in situ* record is incompatible with satellite altimetry (Lombard et al. 2006). Explicit calculation of

$\delta A_E/\delta t$  (Table 3) supports the assumption of a negligible contribution from  $\delta A_E/\delta t$ , as during the ERBE period, the mean atmospheric tendencies reported by NRA and ERA-40 are  $0.0\pm0.1$  PW and  $0.0\pm0.0$  PW, respectively. During the CERES period, NRA reports a tendency of  $0.0\pm0.1$  PW. Thus, the differences between estimates of the ocean to land transport are likely associated with errors in TOA fluxes projected onto the land-ocean residual calculation, and errors in reanalysis fields. From comparison of ERA-40 and NRA during the ERBE period, the latter are order 0.5 PW.

#### 4. The global mean annual cycle

For the annual cycle at TOA, Fig. 3 shows the global-mean, global-ocean, and global-land  $R_T$  and albedo for the ERBE-tuned and CERES best-estimate FM1 and FM2 fields with  $\pm 2\sigma_1$  also shown (shading). Characteristics of the global, global-ocean, and global-land annual cycles, including uncertainty estimates, are summarized in Tables 4, 5, and 6, respectively. The annual mean has been removed from these values. As differences in seasonal characteristics among the data sources are small, only best-estimate characteristics are reported and only the ERBE-tuned and best-estimate fields are plotted.

Globally,  $R_T$  variations are characterized by an annual maximum in February of 4.3 PW, somewhat after the perihelion owing to the albedo changes from January to February. Then there is a rapid decline during boreal spring to a minimum of -5.2 PW in June, followed by an increase from June through the following February. Fluctuations in albedo exhibit a significant semi-annual cycle with maxima occurring in May and November. Over the global-ocean, whose areal extent is biased significantly towards the Southern Hemisphere,  $R_T$  peaks at 8.0 PW during

austral summer and achieves a minimum of -9.7 PW during austral winter. Uncertainty in the annual cycle of  $R_T$  over ocean is non-negligible. Over land, which is disproportionately in the Northern Hemisphere,  $R_T$  peaks at 4.5 PW in June and reaches a minimum of -4.6 PW in December.

Globally, the  $R_T$  maximum (Fig. 3) in February is comprised of an  $ASR$  maximum of 3.1 PW (Fig. 4) and an  $OLR$  minimum of -1.6 PW. Peak  $OLR$  of 2.2 PW occurs in August with the minimum in  $ASR$  at -3.8 PW in June, but the  $OLR$  changes correspond mostly to the land contribution, and coincide with the phase of surface temperature in the Northern Hemisphere. Uncertainty in  $ASR$ , particularly for  $\Delta_D$ , is substantially greater than for  $OLR$ . For the global-ocean (Fig. 4b), the seasonal cycle is dictated largely by  $ASR$ , which reaches a maximum of 7.6 PW in January and a minimum of -9.5 PW in June. The annual cycle of  $OLR$  over the global-ocean is small, with a -0.6 PW minimum in May and 0.7 PW peak in August. Over land, peak  $ASR$  of 5.7 PW occurs in June with a minimum of -5.6 PW in December. The range of uncertainty associated with  $\pm 2\sigma_1$  for  $R_T$ ,  $ASR$  and  $OLR$  (shading in Fig. 4) is larger than differences between ERBE-tuned and best-estimate FM1 and FM2 fields for all months. Thus the possibility that differences between the ERBE and CERES estimates result from the different time periods considered cannot be ruled out.

The seasonal cycle of  $\delta A_E/\delta t$  is non-negligible (Fig. 5), increasing during boreal spring and decreasing during late boreal summer and early autumn with a peak in  $\delta A_E/\delta t$  of 1.3 PW in June and a minimum of -1.8 PW in September. Over ocean, the relatively small annual cycle of  $\delta A_E/\delta t$  peaks in boreal summer and winter near 0.5 PW and reaches a minimum of -0.8 PW in September. The atmospheric total energy divergent flux  $\nabla \cdot \mathbf{F}_A$  over ocean (Fig. 5) peaks at 5.2

PW during boreal winter and achieves a minimum during summer of -1.1 PW. The mean ocean-land flux is thus dictated largely by exchanges during boreal winter. Over land the annual cycle of  $\delta A_E/\delta t$  is greater than that over ocean and  $\nabla \cdot \mathbf{F}_A$ , by constraint, balances that over ocean.

The annual cycle of  $F_S$  from (2) resembles that of  $R_T$  (with opposite sign) (Fig. 6 vs Fig 3), with an  $F_S$  maximum of 6.3 PW occurring in austral winter and minimum of -4.0 PW occurring in austral summer, based on CERES /NRA estimates over ocean and CLM modeled fields over land, as compared with 5.3 PW and -4.2 PW for  $R_T$ . The global-ocean annual cycle (Fig. 6), with a maximum of 7.1 PW and minimum of -4.5 PW, is approximately in phase with the global annual cycle. Estimates of  $F_S^o$  based on the residual methods of (2) (solid line), which relies primarily on  $R_T$  and  $\nabla \cdot \mathbf{F}_A$ , and (5), which relies on  $R_T$  and  $F_S^l$  from the CLM, agree closely through the annual cycle, providing an initial estimate of uncertainty associated with  $\nabla \cdot \mathbf{F}_A$  and the assumptions in (2) discussed in Section 2. The difference between the methods is less than, or comparable to, the uncertainty associated with  $\pm 2\sigma_1$  (shading) for all months, and the main differences occur in northern winter and may reflect uncertainties in land snow and ice (Qian et al. 2006).  $F_S^l$  (Fig. 6), is out of phase by approximately five months, with a maximum of 0.9 PW occurring in November and a minimum of -1.0 PW during May. Interannual variability is dominated by variability over ocean of order 0.3 PW as the CLM fields suggest negligible interannual variability over land ( $<0.1$  PW).

A comparison between inferred and direct estimates of  $O_E$  during the ERBE period (Fig. 7), along with their rates of change, reveals that the seasonal peak in  $O_E$  follows the austral summer, when the global-mean peaks in  $SI$  and  $R_T$  occur, and when the Southern Hemisphere experiences substantial net TOA and surface heating. The magnitude of the March-April peak in  $O_E$  varies

considerably among the datasets, with estimates from  $F_S$  in  $10^{22}$  J based on ERBE+ERA (not shown) and ERBE+NRA at 3.0, GODAS at 3.2, WOA at 4.2, and JMA at  $4.5 \times 10^{22}$  J. During austral winter,  $O_E$  falls sharply; reaching an estimated minimum for  $F_S$  (both ERBE-ERA and ERBE-NRA) of -2.8, for GODAS of -3.6, for WOA of -3.91, and for JMA of  $-4.01 \times 10^{22}$  J. The ocean temperature datasets thus imply a substantially larger annual cycle of  $O_E$  than do  $F_S$  estimates, with annual cycle standard deviation  $S$  of  $\delta O_E / \delta t$  greater than inferred from  $F_S$  ( $S=5.9$  PW for WOA,  $5.1 \pm 1.7$  PW for GODAS,  $6.1 \pm 2.0$  PW for JMA, and  $4.4 \pm 1.0$  PW for ERBE-NRA  $F_S$ , where uncertainty estimates represent  $2\sigma_1$ ).

## 5. Discussion

The estimates presented herein build upon the wide array of estimates currently available and reviewed by Kiehl and Trenberth (1997). Improvements have been made possible by both the newly available CERES fields, and recent considerations of the net planetary imbalances from global-ocean and global-land energy tendencies. At TOA, estimated sampling error ( $\pm 2\sigma_1$ ) associated with interannual variability of the global annual mean agrees well with early estimates (Ardanuy et al. 1992, Kyle et al. 1993). For many terms the sampling error is considerably smaller than the differences among the data at different times (Table 2) and hence the importance of errors in the retrieval process is highlighted as a fundamental limitation on current estimates. The best-estimate values of  $ASR$  (albedo) reported here lie above (below) the ERBE-tuned estimates and below (above) the TOASRB, ERBE-raw, and CERES ERBE-like estimates. Confidence in the best-estimate exceeds that in the raw retrievals as  $ASR-OLR$  differences in the un-tuned products are substantially greater than allowed for by existing planetary imbalances. The best-estimate albedo lies towards the lower end of the range of earlier estimates (presented

by Kiehl and Trenberth 1997). The global OLR value in Fig. 2 corresponds to  $239 \text{ W m}^{-2}$  which is larger than the  $235 \text{ W m}^{-2}$  estimated by Kiehl and Trenberth (1997) based on ERBE data without the adjustments in Appendix I. *ASR* of the ERBE-raw and CERES ERBE-like estimates contain significantly greater uncertainty due to interannual variability ( $2\sigma_1$ ) than for the CERES product or the tuned ERBE product, suggesting that the ERBE retrieval methods or discontinuities associated with the loss of NOAA-9 (Trenberth 1997) may artificially inflate  $\sigma_1$ . Other aspects of the TOA budget are obscured primarily by the limited span of the observational record. The significance of differences in monthly deviations from the annual mean is an example of such a quantity, as for all terms, ERBE-tuned and CERES best-estimate differences are less than  $2\sigma_1$  (Figs. 3 to 5).

Within the atmosphere the NRA reanalyses depict a mean ocean to land transport near 2 PW during the ERBE period (Fig. 8), consistent with inferences from TOA fluxes (Fig. 2). However, during the CERES period, the computed NRA land-ocean transport is  $2.7 \pm 0.2 \text{ PW}$  (Table 3), a value inconsistent with  $R_T^1$ . The ERA-40 estimates are less consistent over time than for NRA and the inconsistencies are associated, in part, with established shortcomings of ERA-40 fields during the 1990s (Uppala et al. 2005). In particular, major changes in the observing system occurred with the introduction of SSM/I data in 1987 (that were not used in NRA) and problems occurred following the Mount Pinatubo eruption in 1991 whose aerosol contaminated the radiances that were assimilated in ERA-40 (whereas NRA used retrievals) (Uppala et al. 2005). Other changes in Fig. 8 affected both reanalyses, albeit differently, in particular the change from TIROS-N (Television and Infrared Observation Satellite) operational vertical sounder (TOVS) to



the Advanced TOVS (ATOVS) in NRA from 1998 to 2001. Accordingly, the decadal variability in Fig. 8 is largely spurious and certainly is not reproducible.

However, the variability in the ocean-to-land energy transport from NRA and ERA-40 over time (Fig. 8) are among the most temporally extensive estimates of the atmospheric energy budget. They are compared (Fig. 8) with the transport as inferred from ERBE-tuned and CERES best-estimate fluxes per (3) using NRA estimates of  $\delta A_E/\delta t$ . Large interannual variability is mostly real and corresponds to ENSO in part, and includes the sharp increases in 1992 through 1994, and 1997/98 that coincide with El Niño events and the subsequent declines in 1999/2000 that accompany a transition to La Niña conditions (Trenberth et al. 2002). Lower frequency variability, including the increase in transport from the mid-1980s through the mid-1990s is also notable but can not be confirmed owing to the observing system changes, although these are smaller for NRA.

While the reanalyses are susceptible to changes in their input data, early satellite platforms including Nimbus-7, the ERBE and ERBS (Kyle et al. 1993) contain significant uncertainty due to their limited sampling rate and spatial coverage, and their reliance on recalibration methods following major outages and gradual calibration drifts associated with sensor and orbital degradation over time (Trenberth 2002). While the ERBE and CERES retrievals capture aspects of the NRA transport, including the magnitude of the transport during the ERBE period and the decline in transport early in the ERBE period, other significant differences exist including especially the magnitude and variability of the transport during the CERES period. For these reasons, among others, it is thus not possible to distill the differences in decadal variability from NRA, ERBE, and CERES estimates into their real and spurious components, as natural

variability is likely to be large. The susceptibility of discrete observation periods, such as those of ERBE and CERES (shaded), or estimates based on intermittently observed data, such as the reanalyses, to low frequency variability is therefore real and needs to be quantified in any attempt to more generally assess the energy budget in the context of climate change. The present analysis shows that current datasets are not able to adequately achieve these standards and long-term variability in the budget remains a substantial and open science question.

Globally, the annual cycle of  $R_T$  is influenced primarily by  $ASR$  and secondarily, though importantly, by  $OLR$ . About 80% of the seasonality in  $SI$  projects onto  $ASR$ ; more than would be expected from a globally uniform and constant albedo (70%) owing to the interaction of spatial and temporal variability of albedo with  $SI$ . The phase of global mean  $ASR$  is determined primarily by the seasonal cycle of  $SI$ , which is governed by the Earth's orbit. The annual cycle of global-ocean albedo is predominantly semiannual, with peak values leading the solstices slightly, and minima in March and September coincide with the equinoxes and are associated with the cross-equatorial migration of the Inter-Tropical Convergence Zone. The global-ocean (Fig. 4b) seasonal cycle of  $OLR$  is small relative to that over land, and there is a near cancellation in  $OLR$  variability over ocean between hemispheres. In contrast, over land (Fig. 4c), both  $ASR$  and  $OLR$  contribute significantly to the annual cycle of  $R_T$ , and significant hemispheric asymmetries project onto the global mean. Albedo over land peaks in boreal winter and varies in a manner consistent with the enhancement of surface albedo due to increased winter snow and ice extent. Globally, the seasonal cycle of  $R_T$  is balanced primarily by  $F_S^0$  and the contribution of  $\delta A_E/\delta t$  is secondary (globally  $\nabla \cdot \mathbf{F}_A = 0$ ). For the global-ocean domain, therefore, a near balance occurs between TOA radiation and surface fluxes but also with important seasonal contributions from

transport of energy from ocean to land in northern winter, and vice versa in southern winter albeit much weaker. These aspects are clearer in Fasullo and Trenberth (2007) who report on the meridional structure of these changes.

The annual cycle of  $O_E$  estimates from WOA and JMA significantly exceed ( $\sim 30\%$ ) that which can be accounted for by both the TOA and surface energy budgets. Indeed, the differences between the ocean temperature datasets are themselves quite large and agreement between  $F_S$  and GODAS fluxes is better than for WOA and JMA. More detailed regional investigations of the spatial structure of differences in  $O_E$  (Fasullo and Trenberth 2007) reveal that the biggest problems lie over the southern oceans, where the scarcity of observations suggests shortcomings in the ocean datasets.

Although changes in  $R_T$  have been reasonably matched to changes in  $O_E$  (Hansen et al. 2005; Willis et al. 2004), systematic errors remain to be addressed. However, the mean fields and the links between TOA, atmospheric, and surface fluxes, particularly with regard to  $\nabla \cdot \mathbf{F}_A$  and  $O_E$ , are found to be insufficient for climate change purposes, as the uncertainties are too large. The likely uncertainties in  $O_E$ , especially in the Southern Hemisphere (Fasullo and Trenberth 2007), can alter apparent trends if the ocean observing system changes, as it has in recent years with ARGO floats (Gouretski and Koltermann 2007), and may be a source of the discrepancy between  $O_E$  tendencies and sea level changes (Lombard et al. 2007); a topic addressed more completely in Trenberth and Fasullo (2007). An additional and significant challenge in understanding the relationship between the energy budget and climate change is posed by the large temporal gaps in the observational record which render natural variability indistinguishable from calibration drift and error, and platform related hardware differences.

## 6. Concluding remarks

The budget established herein (Fig. 2) is thought to improve substantially upon previous estimates (summarized in Kiehl and Trenberth 1997 and Wunsch 2005) by including CERES retrievals, by adjusting TOA fluxes to recent estimates of the global-land and global-ocean imbalances, and by including a consideration of the annual cycle of all terms and an assessment of errors. Associated best-estimates of albedo and ocean to land fluxes are also derived. The annual cycle of  $R_T$  is found to be substantial and is augmented by variability in albedo and  $OLR$ . CLM estimates of  $F_S^l$  show it to be both an important contributor to the annual cycle of  $F_S$  globally and associated with a strong seasonal cycle in  $\nabla \cdot \mathbf{F}_A$  over land. The annual mean net flow from ocean to land is order 2% of the total incoming and outgoing radiation (122 PW or globally  $239 \text{ W m}^{-2}$ ) and 4.4 times the estimated net imbalance associated with global warming. Accordingly it is important for models to be able to capture the land and ocean albedo, and thus snow cover, vegetation, and cloud cover characteristics which all vary considerably monthly. The annual cycle of the net radiation has a range of 9.5 PW and thus is about 8% of the net flow through the system. Hence a 5% error in this is comparable to the climate change signal. Although the global net radiation is dominated by orbital effects, changes in planetary albedo are non-trivial and  $OLR$  contributes significantly to enhance the amplitude of the annual cycle.

The differences in the land and ocean energy budgets are considerable. There is an annual cycle in land energy storage of order 2 PW range, associated with flows of moisture onto land and the return river flows plus the changes in water storage in both liquid and solid forms. However, the small variability of  $L_E$  contrasts with that of  $O_E$ , which has a range of 11.6 PW (for

$F_s$ , Fig. 6) or nearly 10% of the net flow through the system. Accordingly, over 5 PW of energy flows from ocean to land in northern winter (Fasullo and Trenberth 2007).

The present study also identifies some key uncertainties that limit our present understanding of the budgets and the Earth's climate balance. Variability in  $F_S$  across the annual cycle is shown to be at odds with observations of  $O_E$ , a difference that is attributed largely to errors in ocean analyses. In addition, variability over time, as indicated by  $\sigma_I$ , is substantial for some terms, so that sampling issues are significant contributors to overall uncertainty. On the interannual timescale, uncertainties are significant for all aspects of the budget and relate to the challenges posed by episodic observation of the budget, in the context of platform calibration issues and hardware inconsistencies, intertwined with significant natural variations. Specifically, reanalyses are shown to be unable to resolve  $\nabla \cdot \mathbf{F}_A$  to the accuracy required to diagnose global-ocean  $F_S$  from TOA imbalances in a changing climate. Progress on the TOA issues is assured through ongoing work in the CERES project (Wong et al. 2006; Wielicki et al. 2006). In the interim, we have made plausible assumptions in adjusting these to provide new estimates of the global radiation budget and its annual cycle. In the atmosphere, reanalyses exhibit inconsistencies related to differences in the data assimilated and the data changes over time, and the need for improvements have been identified. The uncertainties are explored in more detail as we examine the meridional structure of energy storage and flows (Fasullo and Trenberth 2007) and a more in-depth analysis of the ocean (Trenberth and Fasullo 2007). Further refinements in these key uncertainties are likely to contribute substantially to a more complete understanding of the flow of energy through the climate system and its role in our changing climate.

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## **Appendix I: Method for adjusting ERBE-adjusted fluxes**

ERBE estimates are based on observations from three satellites (ERBS, NOAA-9, and NOAA-10) using both scanning and non-scanning radiometers. Data from ERBE are reported on a  $2.5^\circ$  grid and span from February 1985 to December 1989. While the NOAA-9 and 10 satellites were sun-synchronous and polar-orbiting, thus providing diurnally biased observations, the ERBS satellite was in a slowly precessing  $57^\circ$  inclined orbit that allowed for complete diurnal sampling in a 72-day period and thereby helped reduce the impact of diurnal aliasing during the ERBE period. Trenberth (1997) provides a detailed discussion of errors and processing of the original ERBE data (hereafter referred to as ERBE-raw) including adjustments to the ERBE data to ensure some homogeneity across the mission when NOAA-9 was lost and the observations were reduced to those from 3 to 2 satellites. In addition, a subsequent adjustment was to the albedo to enforce a global energy balance of zero, consistent with the changes in  $O_E$  for this period (Levitus et al. 2005) (referred to as ERBE-tuned).

During the course of our investigation it was discovered that the global corrections made in Trenberth (1997) should have discriminated between land and ocean, as analysis of those domains separately clearly revealed a discontinuity when NOAA-9 failed. Additional steps were therefore taken to remove this error. The failure of the afternoon orbiting NOAA-9 satellite was found to be associated with a spurious reduction in retrieved OLR over land, where the diurnal cycle is large and maximum OLR occurs during afternoon. Accordingly, an OLR increase over land following the NOAA-9 failure of  $2.95 \text{ W m}^{-2}$  is imposed. Along with compensating for diurnal aliasing, this correction has the beneficial effect of yielding continuity in the inferred ocean to land transport. The difference in global-mean OLR prior to and following the failure of

NOAA-9 (+0.37 PW) closely matches the difference reported by the temporally continuous ERBS retrievals (+0.39 PW), further justifying this approach.

The original data at T63 resolution from Trenberth (1997) are available at:

<http://www.cgd.ucar.edu/cas/catalog/satellite/erbe/means2.html>. The steps taken include:

1) **Adjust OLR:** An increase to OLR over land is prescribed to match the spurious discontinuity of  $+2.95 \text{ W m}^{-2}$ . OLR over ocean is not altered. It is then found that the associated ocean to land flux is approximately continuous through the NOAA-9 transition. Also the difference between the post- and pre-NOAA-9 failure global-mean OLR (+0.39 PW) is found to agree approximately with the 60°S-60°N difference from ERBS (+0.37 PW) (Wong et al. 2005).

2) **Retune reflected and absorbed solar radiation.** The adjustment is required to reset the  $\text{NET}=0$  as per Levitus et al. (2005): Absorbed and reflected solar radiation are adjusted in tandem such that they total to the original solar insolation. A tuning parameter,  $\alpha$ , is derived empirically to satisfy this criterion. If  $f$  is the ratio of the new global bias to the global mean  $ASR$  then a new  $ASR$  is derived from  $ASR_{new}=ASR(I-f)/\alpha$  which is solved empirically to give  $\alpha=1.0055$ . The  $ASR$  scaling contributes to a few instances for which  $ASR < 0$  and thus  $ASR$  is reset to 0.



## **Appendix II. Acronyms**

$A_E$ : Total Atmospheric Energy

ASR: Absorbed solar radiation

ATOVS: Advanced TOVS

CERES: Clouds and the Earth's Radiant Energy System

CLM: Community Land Model

ERBE: Earth Radiation Budget Experiment

ERBS: Earth Radiation Budget Satellite

ECMWF: European Centre for Medium Range Weather Forecasts

ENSO: El Niño–Southern Oscillation

ERA-40: ECWMF 40 year reanalysis

$F_A$ : Atmospheric Transport of Energy

$F_O$ : Oceanic Transport of Energy

$F_S$ : Net surface energy flux (turbulent + radiative)

FM1, FM2; FM3, FM4: CERES twin instruments Flight Models 1 and 2 on the *Terra* spacecraft  
and Flight Models 3 and 4 on the *Aqua* satellite

GODAS: Global Ocean Data Assimilation System

JMA: Japanese Meteorological Agency

$L_E$ : Total Land Energy Storage

MOM3: Modular Ocean Model v3

NCEP: National Center for Environmental Prediction

NCAR: National Center for Atmospheric Research

NOAA: National Oceanic and Atmospheric Administration

O<sub>E</sub>: Total Ocean Energy Storage

NRA: NCEP/NCAR reanalysis

OLR: Outgoing Longwave Radiation

PW: PetaWatt

R<sub>T</sub>: Net TOA Radiation

SI: Solar Irradiance

SSM/I: Special Sensor Microwave Imager

TIROS: Television and Infrared Observation Satellite

TOA: top-of-atmosphere

TOVS: TIROS Operational Vertical Sounder

T42: Triangular truncation at wave-number 42 (about 2.8° grid)

TOASRB: TOA Surface Radiation Budget

WOA: World Ocean Atlas

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## Figure Captions

Figure 1. Schematic of the energy balance illustrating the notation and direction of the fluxes.

The ASR is directed down, OLR is upwards, and their net  $R_T$  is downwards, while the surface flux  $F_s$  is upwards. The net fluxes into the atmosphere and ocean are mostly balanced by the atmospheric and oceanic divergences of the energy transports  $F_A$  and  $F_O$  (as shown), and the changes in energy storage (not shown). For land there is no internal transport of consequence and the surface flux is balanced by the storage changes.

Figure 2: CERES-period mean best-estimate FM1 TOA fluxes [PW] globally and for global-land (right) and global-ocean (left) regions.

Figure 3: a) Global, b) global-ocean, and c) global-land mean annual cycles of albedo (%) and  $R_T$  (PW) where shading represents  $\pm 2\sigma_1$  of monthly means and the annual mean has been removed.

Figure 4: a) Global, b) global-ocean, and c) global-land mean annual cycles for  $ASR$ ,  $OLR$ , and  $R_T$  (PW) where shading represents  $\pm 2\sigma_1$  of monthly means and the annual mean has been removed.

Figure 5: a) Global, b) global-ocean, and c) global-land estimates of atmospheric total energy divergence ( $\nabla \cdot F_A$ ) and tendency ( $\delta A_E / \delta t$ ) are shown in PW where shading represents  $\pm 2\sigma_1$  of monthly means and the annual mean is shown (dotted).

Figure 6: Global, global-ocean, and global-land estimates of net upwards surface flux ( $F_s$ ) are shown in PW where shading represents  $\pm 2\sigma_1$  of monthly means. Global ocean estimates are derived from (2) except for the dotted line, which is derived from (5).

Figure 7: Annual cycles of a)  $O_E$  ( $10^{22}$  J) and b)  $\delta O_E/\delta t$  (PW) are shown with ranges of uncertainty  $\pm 2\sigma_1$  for monthly means from the GODAS and JMA datasets (shading). The GODAS fields have been extended in latitude using JMA fields,  $O_E$  estimates were differenced to provide  $\delta O_E/\delta t$ , and  $F_S$  was integrated in time to provide  $O_E$  anomalies for the ERBE and NRA data.

Figure 8: Interannual variability in inferred net ocean to land energy transport in PW is shown by 12-month running means for ERBE and CERES best-estimate  $R_T$  over land and NRA  $\delta A_E/\delta t$  fields. Transports calculated directly from NRA and ERA-40 fields are also shown.

## Figures

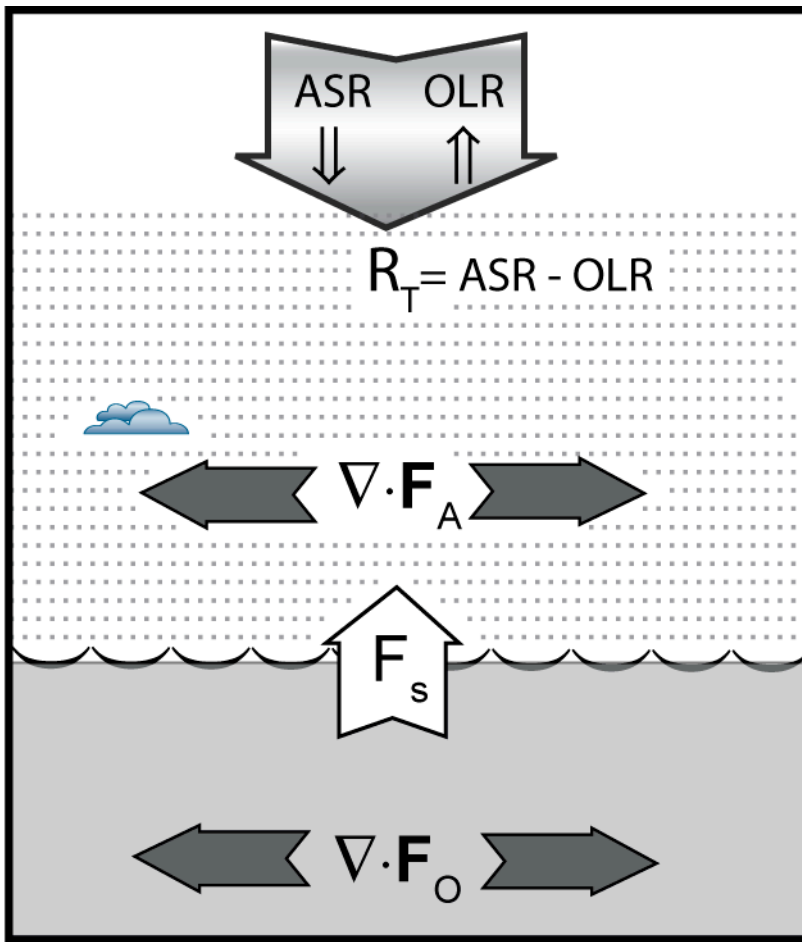


Figure 1. Schematic of the energy balance illustrating the notation and direction of the fluxes. The ASR is directed down, OLR is upwards, and their net  $R_T$  is downwards, while the surface flux  $F_s$  is upwards. The net fluxes into the atmosphere and ocean are mostly balanced by the atmospheric and oceanic divergences of the energy transports  $\mathbf{F}_A$  and  $\mathbf{F}_O$  (as shown), and the changes in energy storage (not shown). For land there is no internal transport of consequence and the surface flux is balanced by the storage changes.

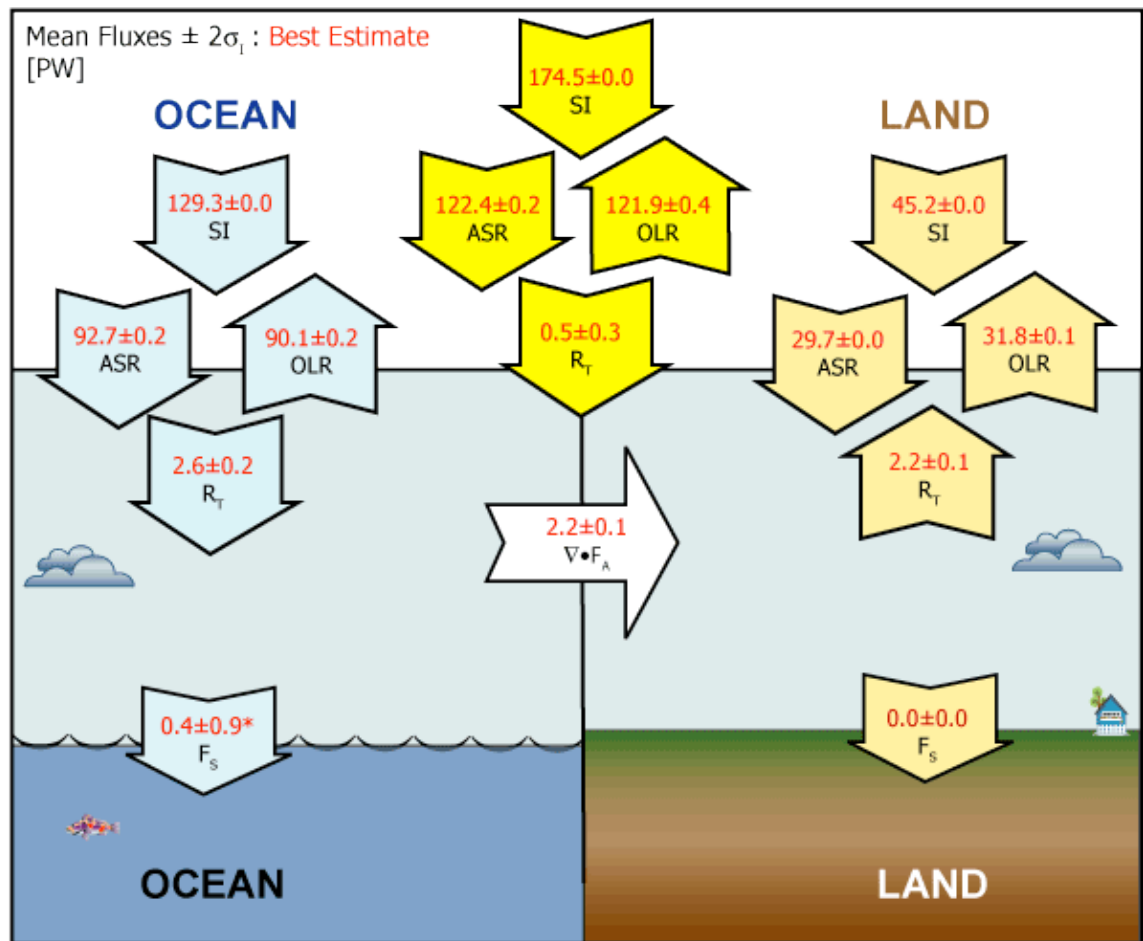


Figure 2: CERES-period mean best-estimate FM1 TOA fluxes [PW] globally and for global-land (right) and global-ocean (left) regions.

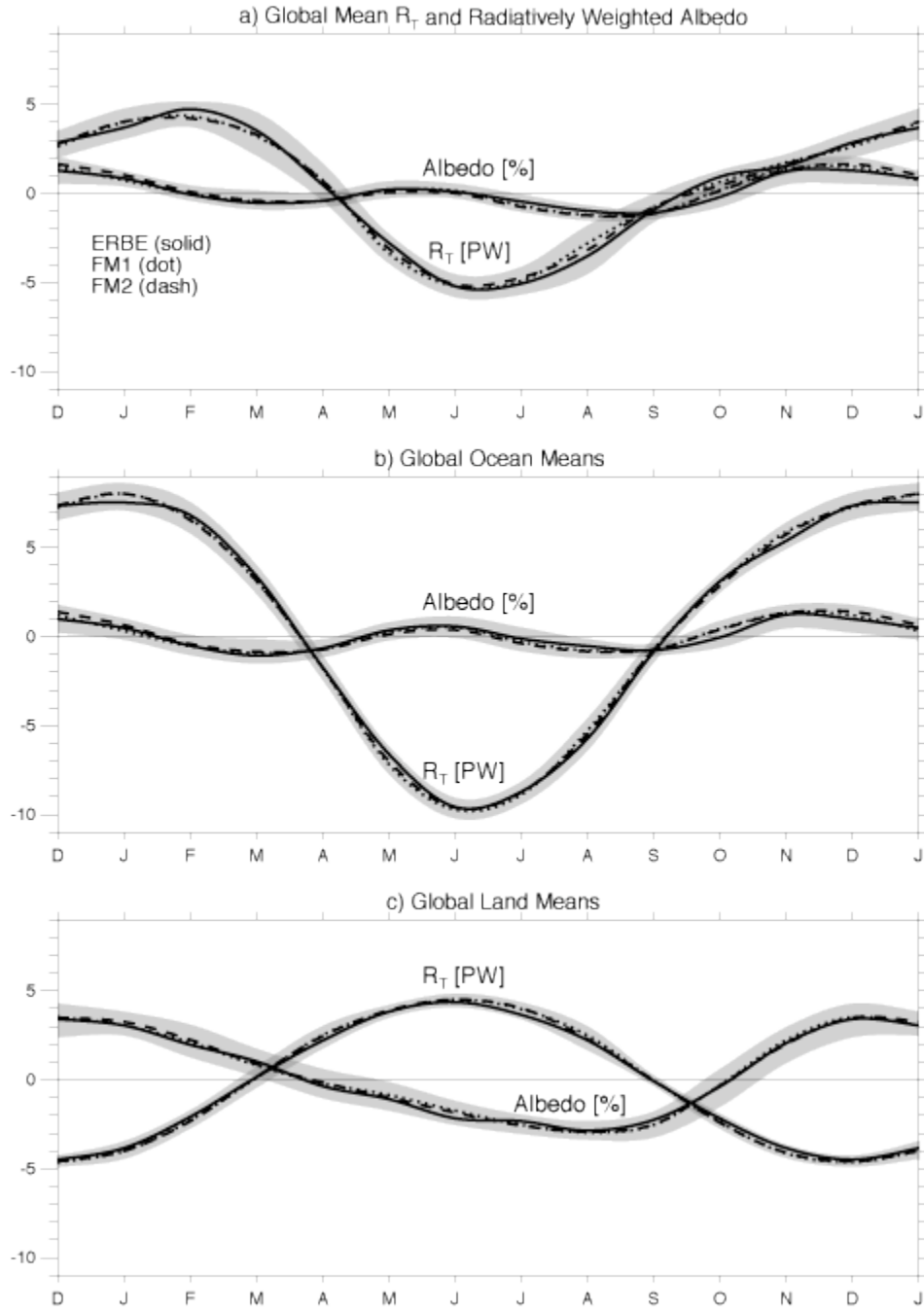


Figure 3: a) Global, b) global-ocean, and c) global-land mean annual cycles of albedo (%) and  $R_T$  (PW) where shading represents  $\pm 2\sigma$  of monthly means and the annual mean has been removed.

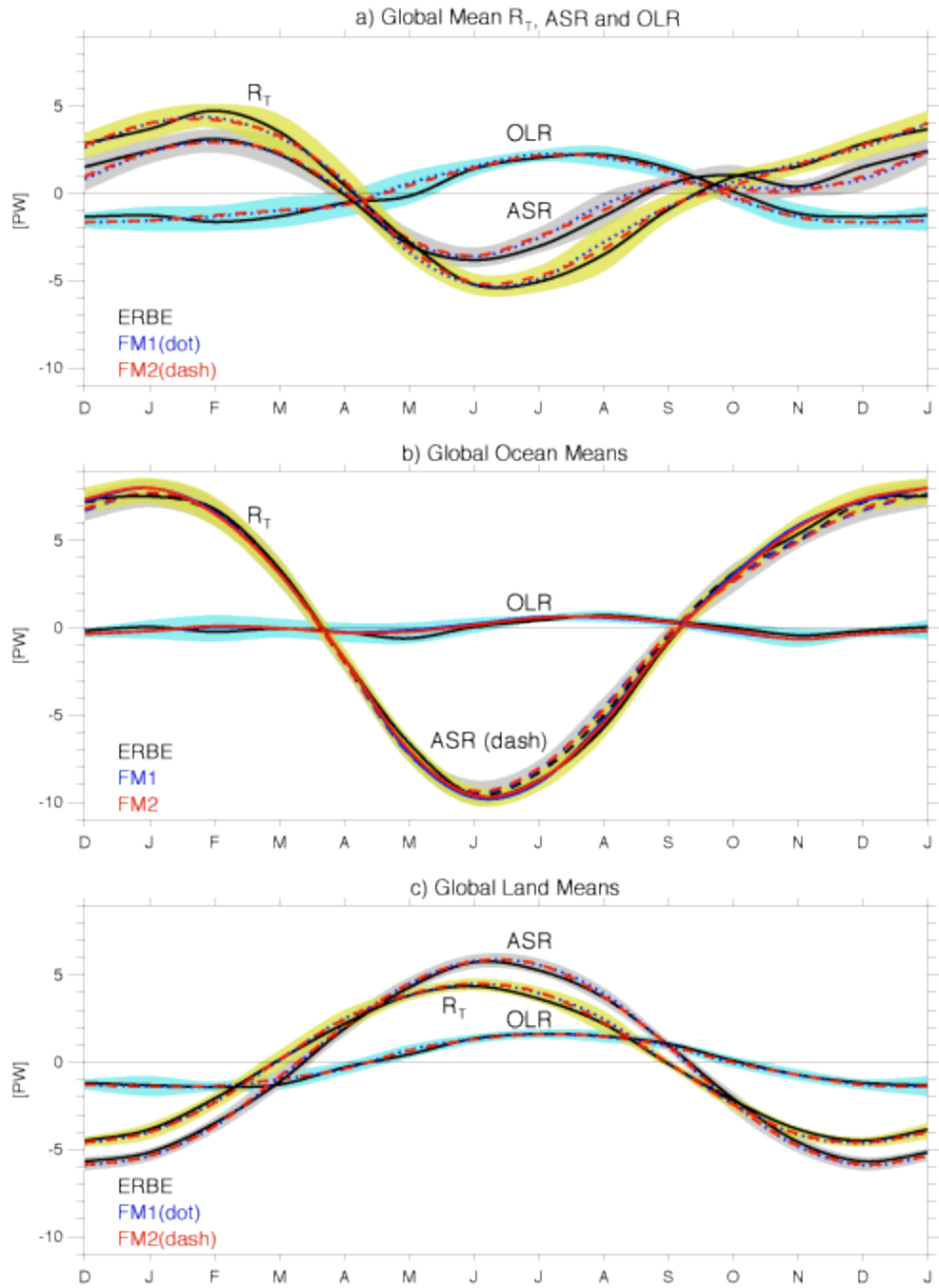


Figure 4: a) Global, b) global-ocean, and c) global-land mean annual cycles for  $ASR$ ,  $OLR$ , and  $R_T$  ( $\text{PW}$ ) where shading represents  $\pm 2\sigma_I$  of monthly means and the annual mean has been removed.

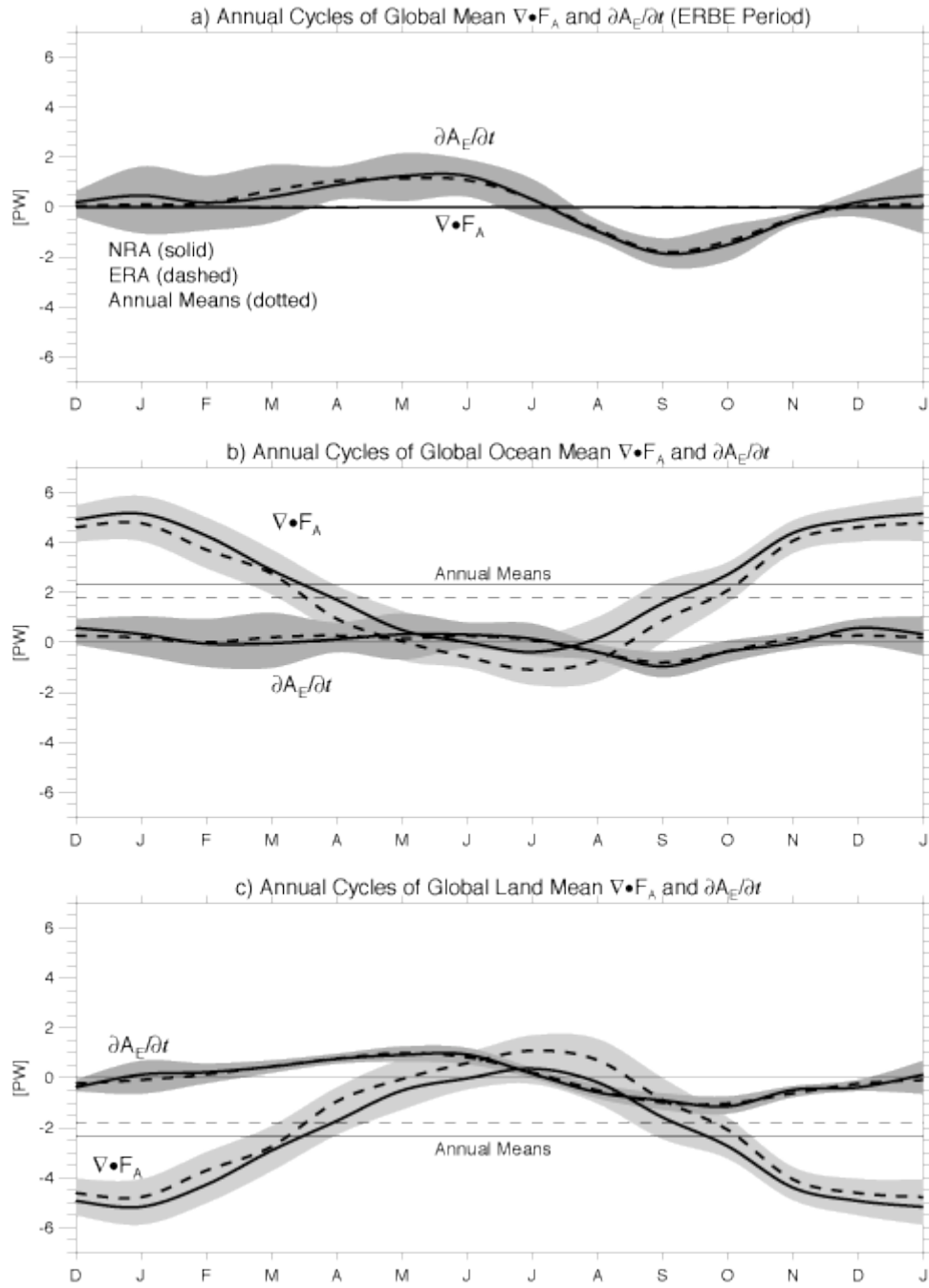


Figure 5: a) Global, b) global-ocean, and c) global-land estimates of atmospheric total energy divergence ( $\nabla \cdot \mathbf{F}_A$ ) and tendency ( $\partial A_E / \partial t$ ) are shown in PW where shading represents  $\pm 2\sigma_I$  of monthly means and the annual mean is shown (dotted).

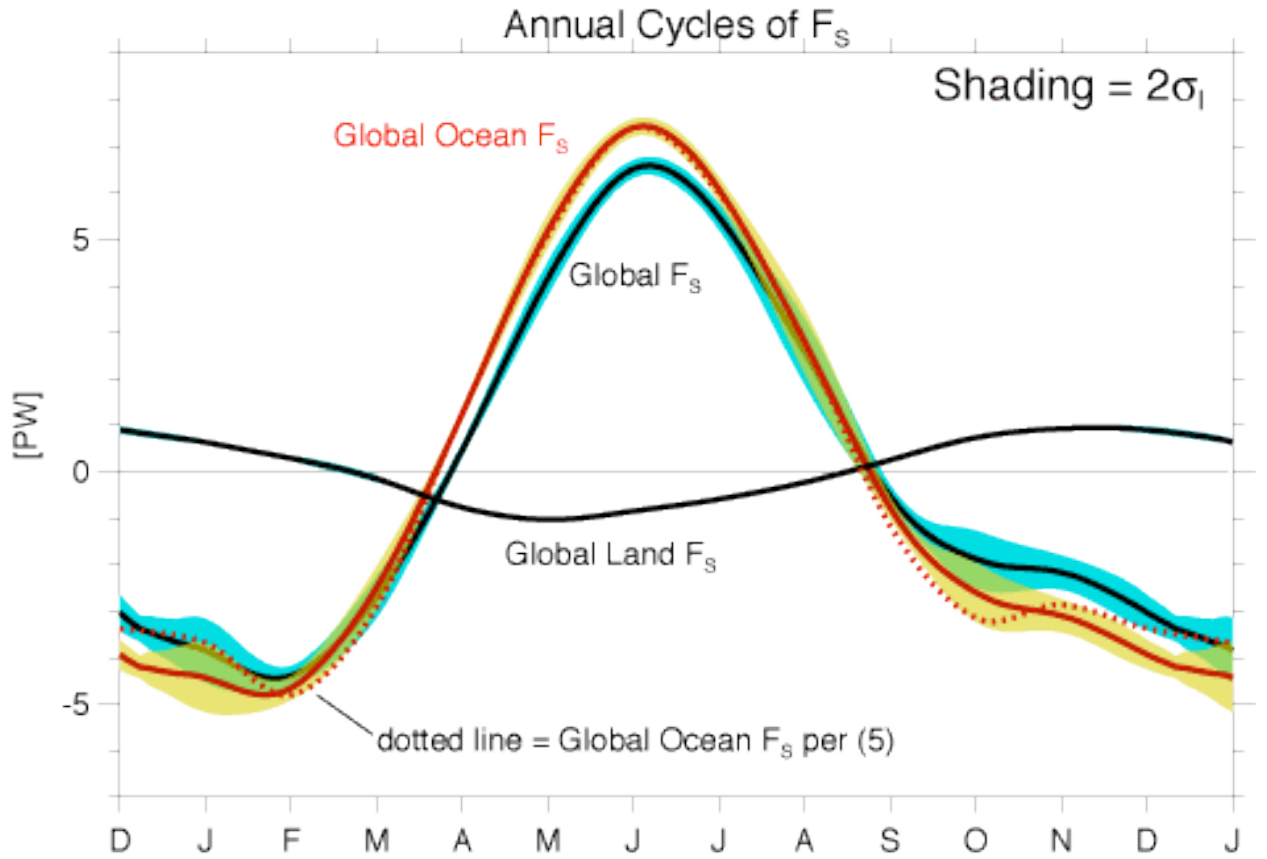


Figure 6: Global, global-ocean, and global-land estimates of net upwards surface flux ( $F_s$ ) are shown in PW where shading represents  $\pm 2\sigma_l$  of monthly means. Global ocean estimates are derived from (2) except for the dotted line, which is derived from (5).



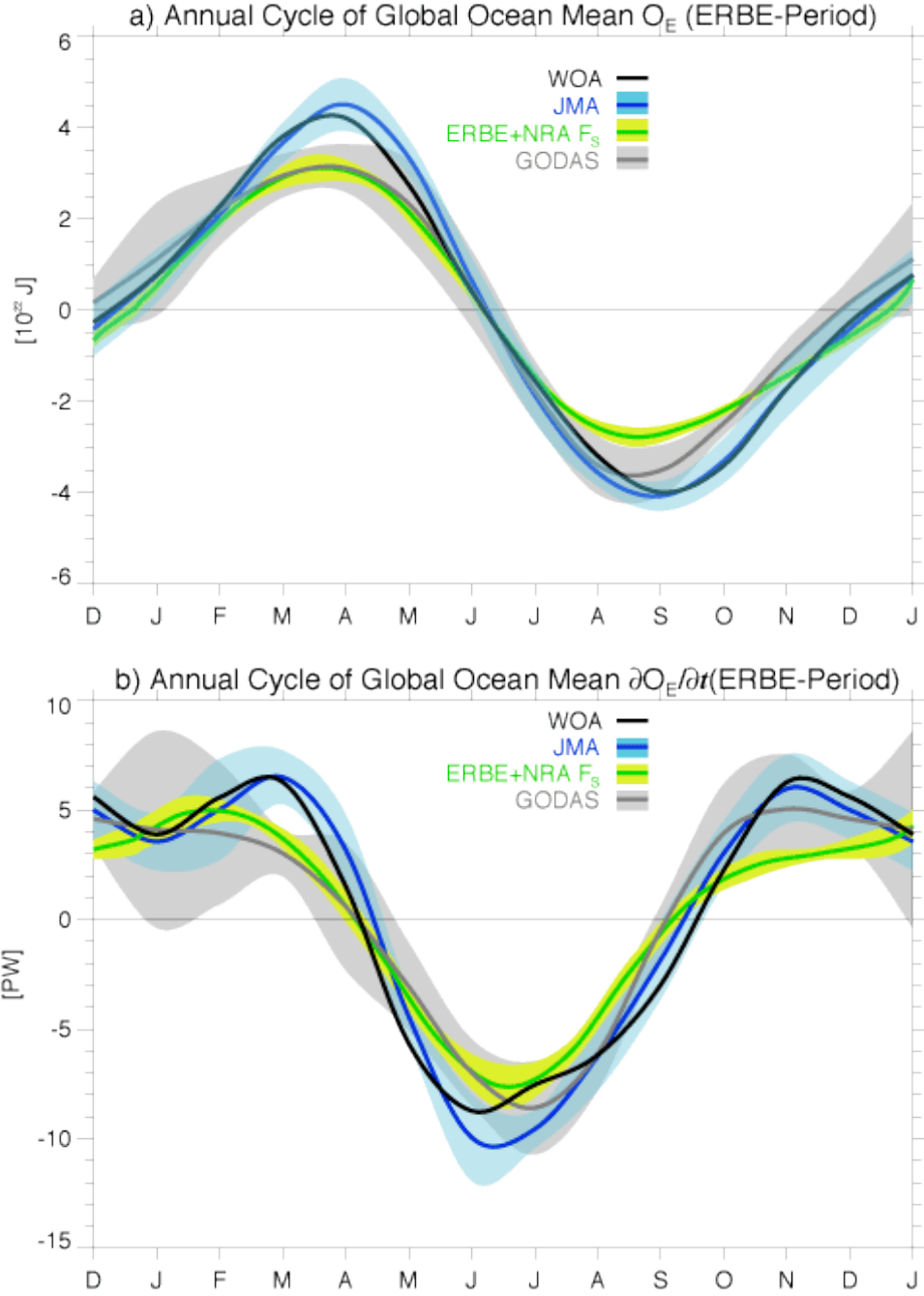


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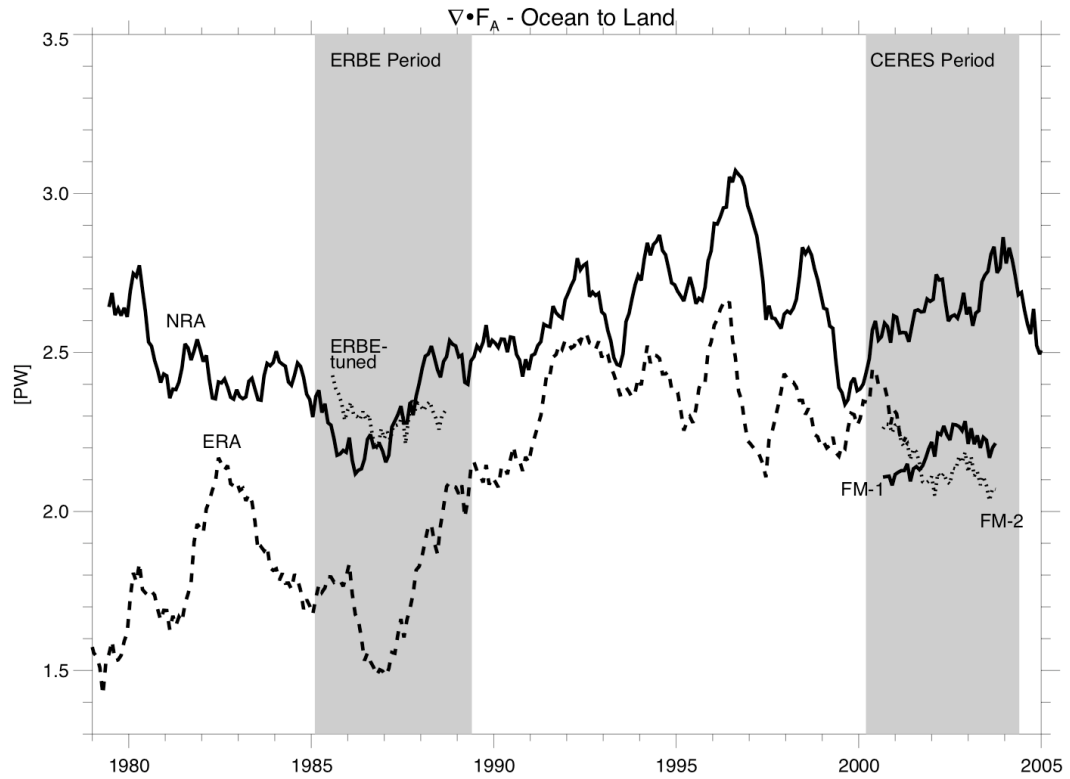


Figure 8: Interannual variability in inferred net ocean to land energy transport *is* shown in PW by 12-month running means for ERBE and CERES best-estimate  $R_T$  over land and NRA  $\delta A_E/\delta t$  fields. Transports calculated directly from NRA and ERA-40 fields are also shown.

Table 1: Maximum likely contribution to error in CERES annual mean  $R_T$  estimates from several sources. These errors are taken from Wielicki et al. (2006) and must add constructively to reconcile  $R_T$  from CERES with independent best-estimates of the global imbalance (e.g. Hansen et al. 2005, Huang, 2006). The terms are distinguished for shortwave (SW) and longwave (LW) components and units are  $W\ m^{-2}$ .

Error Source	SW	LW	$R_T$
Total Solar Irradiance (1361 vs 1365)	+1.0	0.0	+1.0
Absolute Calibration	1.0	1.0	2.0
Spectral Correction	0.5	0.3	0.8
Spatial Sampling	<0.1	<0.1	<0.1
Angle Sampling	+0.2	-0.1	+0.1
Time Sampling (diurnal)	<0.2	<0.2	<0.2
Reference Altitude (20 km)	0.1	0.2	0.3
Twilight SW Flux (-0.25 $Wm^{-2}$ )	<0.1	0.0	<0.1
Near Terminator SW Flux	+0.7	0.0	+0.7
3-D Cloud Optical Depth bias	+0.7	0.0	+0.7
<b>Cumulative Contribution</b>			<b>6.4</b>

Table 2: Global-mean TOA fluxes and attributes from a variety of sources. Uncertainty ranges represent the  $2\sigma_1$  interval. Energy values are in PW.

Source	$R_T$	ASR	OLR	Albedo
Best-estimate, this paper	$0.5 \pm 0.3$	$122.1 \pm 0.2$	$121.7 \pm 0.4$	$0.299 \pm 0.001$
[2000 Mar –2004 May]				
CERES TOASRB	$3.5 \pm 0.3$	$124.3 \pm 0.2$	$120.9 \pm 0.4$	$0.286 \pm 0.001$
[2000 Mar –2004 May]				
CERES ERBE-like	$2.1 \pm 0.3$	$123.9 \pm 0.6$	$121.8 \pm 0.3$	$0.288 \pm 0.003$
[2000 Mar –2004 May]				
ERBE-tuned	$0.0 \pm 0.4$	$119.5 \pm 0.1$	$119.5 \pm 0.5$	$0.313 \pm 0.001$
[1985 Feb –1989 Apr]				
ERBE-raw	$4.6 \pm 0.6$	$123.7 \pm 1.3$	$119.0 \pm 1.5$	$0.289 \pm 0.004$
[1985 Feb –1989 Apr]				

Table 3: Net atmospheric global-ocean total energy divergences and tendencies for the ERBE (Feb. 1985 to Apr. 1989) and CERES (Mar. 2000 to May 2004) periods. Uncertainty ranges represent  $\pm 2\sigma_1$  intervals.

	$\nabla \cdot \mathbf{F}_A$	$\delta A_E / \delta t$ (ocean)	$\delta A_E / \delta t$ (land)
NRA [1985 Feb –1989 Apr]	2.3 $\pm$ 0.2	0.0 $\pm$ 0.2	0.0 $\pm$ 0.1
ERA [1985 Feb –1989 Apr]	1.8 $\pm$ 0.8	0.0 $\pm$ 0.1	0.0 $\pm$ 0.0
NRA [2000 Mar –2004 May]	2.7 $\pm$ 0.2	0.0 $\pm$ 0.1	0.0 $\pm$ 0.1

Table 4: Characteristics of the best-estimate global mean annual cycle standard deviation  $S$ . Uncertainty estimates ( $2\sigma_I$  and  $\Delta_{DI}$ ) are provided where possible. Energy values are in PW.

Characteristic	$R_T$	ASR	OLR	SI	Albedo
$S \pm 2\sigma_I$	$3.4 \pm 0.1$	$2.1 \pm 0.1$	$1.5 \pm 0.1$	$4.2 \pm 0.0$	$0.010 \pm 0.001$
$\Delta_D$	0.3	0.5	0.0	0.0	0.000

Table 5: Characteristics of the best-estimate global-ocean annual cycle standard deviation  $S$ . Uncertainty estimates ( $2\sigma_I$  and  $\Delta_D$ ) are provided where possible. Energy values are in PW.

Characteristic	$R_T$	ASR	OLR	SI	Albedo
$S \pm 2\sigma_I$	$2.6 \pm 2$	$92.0 \pm 0.2$	$89.5 \pm 0.3$	$128.3 \pm 0.0$	$0.283 \pm 0.002$
$S \pm 2\sigma_I(S)$	$6.4 \pm 1$	$6.2 \pm 0.2$	$0.4 \pm 0.1$	$9.3 \pm 0.0$	$0.009 \pm 0.001$
$\Delta_D$	0.4	0.5	0.0	0.0	0.000

Table 6: Characteristics of the best-estimate global-land annual cycle standard deviation  $S$ . Uncertainty estimates ( $2\sigma_I$  and  $\Delta_{DI}$ ) are provided where possible.

Characteristic	$R_T$ [PW]	$ASR$ [PW]	$OLR$ [PW]	$SI$ [PW]	Albedo
$S \pm 2\sigma_I$	$-2.2 \pm 0.1$	$30.0 \pm 0.0$	$32.2 \pm 0.1$	$45.7 \pm 0.0$	$0.346 \pm 0.001$
$S \pm 2\sigma_I(S)$	$3.5 \pm 0.1$	$4.5 \pm 0.1$	$1.2 \pm 0.1$	$5.3 \pm 0.0$	$0.025 \pm 0.002$
$\Delta_D$	0.0	0.0	0.0	0.0	0.000