EARTH'S GLOBAL ENERGY BUDGET

BY KEVIN E. TRENBERTH, JOHN T. FASULLO, AND JEFFREY KIEHL

An update of the Earth's global annual mean energy budget is given in the light of new observations and analyses. Changes over time and contributions from the land and ocean domains are also detailed.

Weather and climate on Earth are determined by the amount and distribution of incoming radiation from the sun. For an equilibrium climate, OLR[1] necessarily balances the incoming ASR, although there is a great deal of fascinating atmosphere, ocean, and land phenomena that couple the two. Incoming radiant energy may be scattered and reflected by clouds and aerosols or absorbed in the atmosphere. The transmitted radiation is then either absorbed or reflected at the Earth's surface. Radiant solar or shortwave energy is transformed into sensible heat, latent energy (involving different water states), potential energy, and kinetic energy before being emitted as longwave radiant energy. Energy may be stored for some time, transported in various forms, and converted among the different types, giving rise to a rich variety of weather or turbulent phenomena in the atmosphere and ocean. Moreover, the energy balance can be upset in various ways, changing the climate and associated weather.

Kiehl and Trenberth (1997, hereafter KT97) reviewed past estimates of the global mean flow of energy through the climate system and presented a new global mean energy budget based on various measurements and models. They also performed a number of radiative computations to examine the spectral features of the incoming and outgoing radiation and determined the role of clouds and various greenhouse gases in the overall radiative energy flows. At the TOA, values relied heavily on observations from the ERBE from 1985 to 1989, when the TOA values were approximately in balance. In this paper we update those estimates based on more recent observations, which include improvements in retrieval methodology and hardware, and discuss continuing sources of uncertainty.

State-of-the-art radiative models for both longwave and shortwave spectral regions were used by KT97 to partition radiant energy for both clear and cloudy skies. Surface sensible and latent heat estimates were based on other observations and analyses. During ERBE, it is now thought that the imbalance

AFFILIATIONS: TRENBERTH, FASULLO, AND KIEHL—National Center for Atmospheric Research,* Boulder, Colorado
*The National Center for Atmospheric Research is sponsored by the National Science Foundation
CORRESPONDING AUTHOR: Kevin E. Trenberth, National Center for Atmospheric Research; P.O. Box 3000, Boulder, CO 80307-3000
E-mail: trenbert@ucar.edu

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[1] A list of all acronyms is given in the appendix.

of ISCCP-FD (calculated on its native equal-area grid), and the three reanalyses NRA, ERA-40, and JRA (which have been regridded to a common T63 grid), with two major parts to the table for the TOA (Table 1a) and surface (Table 1b). Estimates of ocean heat content during ERBE (Levitus et al. 2005) suggest that there was little or no change and this applies then to the global net (Fasullo and Trenberth 2008a). Accordingly, the reanalyses are seriously out of balance by order 10 W m^{-2} and all produce net cooling. The NRA has a known bias in much too high surface albedo over the oceans (Kalnay et al. 1996) that is especially evident in the ocean TOA values (Table 1) and cloud distribution and properties are responsible for substantial errors in both ASR and OLR (Bony et al. 1997; Weare 1997; Trenberth et al. 2001). In ERA-40 OLR is too large by 5–30 W m^{-2} almost everywhere, except in regions of deep convection, and the global bias was 9.4 W m^{-2} in January 1989 (Trenberth and Smith 2008a). Problems with clouds also mainly account for the biases in JRA (Trenberth and Smith 2008b).

At the surface, values are provided for the latent and sensible heat fluxes (LH and SH) as well as the radiative terms, and the net overall is the sum of the solar downward, the net LW upward, and the LH and SH fluxes (upward). The downward land flux associated with global warming (that accounts for melting land ice, etc.) is estimated to be less than about 0.01 PW, or 0.07 W m^{-2}. Thus, in the reanalyses (Table 1b), the net downward flux into the ground is too large to be plausible. Over oceans (Table 1b), to the extent that the net TOA globally is approximately zero for the ERBE period, the ocean warming should also be small and so the net surface flux over ocean is largely a measure of the errors. From Trenberth

SPATIAL AND TEMPORAL SAMPLING

Although we are primarily interested in the global mean energy budget in this paper, it is desirable to assess and account for rectification effects. For example, in KT97, we used a single column model constrained by observations, to represent the average fluxes in the atmosphere. We compared results at TOA with those from the NCAR CCM3 and found good agreement, so that the spatial structure was accounted for. At the surface, the outgoing radiation was computed for blackbody emission at 15°C using the Stefan–Boltzmann law

$$R = \varepsilon\sigma T^4, \qquad (1)$$

where the emissivity ε was set to 1.

If we define a global mean as T_g, then $T = T_g + T'$, where the T' refers to departures from the global mean in either time or space. Therefore, $T^4 = T_g^4(1 + T'/T_g)^4$. We expand the bracket and take the global mean, so that the T' and T^3 terms vanish, and then

$$T^4 = T_g^4(1 + 6[T'/T_g)^2 + (T'/T_g)^4]. \qquad (2)$$

The ratio T'/T_g is relatively small. For 1961–90, Jones et al. (1999) estimate that T_g is 287.0 K, and the largest fluctuations in time correspond to the annual cycle of 15.9°C in July to 12.2°C in January, or 1.3%. Accordingly, the extra terms are negligible for temporal variations owing to the compensation from the different hemispheres in day versus night or winter versus summer. However, spatially time-averaged temperatures can vary from −40°C in polar regions to 30°C in the tropical deserts. With a 28.7-K variation (10% of global mean) the last term in (2) is negligible, but the second term becomes a nontrivial 6% increase.

To compute these effects more exactly, we have taken the surface skin temperature from the NRA at T62 resolution and 6-h sampling and computed the correct global mean surface radiation from (1) as 396.4 W m^{-2}. If we instead take the daily average values, thereby removing the diurnal cycle effects, the value drops to 396.1 W m^{-2}, or a small negative bias. However, large changes occur if we first take the global mean temperature. In that case the answer is the same for 6-hourly, daily, or climatological means at 389.2 W m^{-2}. Hence, the lack of resolution of the spatial structure leads to a low bias of about 7.2 W m^{-2}. Indeed, when we compare the surface upward radiation from reanalyses that resolve the full spatial structure the values range from 393.4 to 396.0 W m^{-2}.

The surface emissivity is not unity, except perhaps in snow and ice regions, and it tends to be lowest in sand and desert regions, thereby slightly offsetting effects of the high temperatures on LW upwelling radiation. It also varies with spectral band (see Chédin et al. 2004, for discussion). Wilber et al. (1999) estimate the broadband water emissivity as 0.9907 and compute emissions for their best-estimated surface emissivity versus unity. Differences are up to 6 W m^{-2} in deserts, and can exceed 1.5 W m^{-2} in barren areas and shrublands.

Similar rectification effects may occur for the back radiation to the surface, so that for KT97 the errors tend to offset, but the surface radiation exchanges should be enhanced by about 6 W m^{-2}.