Using Atmospheric Budgets as a Constraint on Surface Fluxes

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(Manuscript received 27 November 1995, in final form 15 February 1996)

ABSTRACT

Possible methods for estimating surface fluxes include (i) use of bulk fluxes and in situ observations, (ii) use of model parameterizations to interpret specified inputs and compute surface fluxes, and (iii) various indirect methods, which rely on the fact that the mass and surface heat, energy, and momentum budgets must balance and so, given computations of all the other components in the various budget equations applied to fields either within the ocean or the atmosphere, fluxes may be inferred as a residual. This paper reviews the third approach using indirect methods and outlines the advantages associated with the use of global atmospheric analyses from four-dimensional data assimilation (4DDA). The time mean increment required in producing analyses in 4DDA is identical to the systematic short-term (6 h) assimilating model forecast error and is most likely due to errors in the model physics. Therefore, the analyses include a desirable fix, which allows the sum of the ''physics'' to be deduced from ''dynamics.'' The focus is on the heat and moisture budgets to infer surface heat fluxes and freshwater fluxes, but with the recognition of the need to balance the mass budget as well. The diurnal cycle of the vertically integrated mass budget for July 1985 and January 1996 from National Centers for Environmental Prediction (formerly the National Meteorological Center) reanalyses is presented, revealing the strong semidiurnal tide and highlighting the need for at least four-times-daily data. The new results reveal that gross violations of the mass budget continue to be present, but these can be allowed for. A discussion is given of other sources of errors contributing to the heat and moisture budgets.

1. Introduction

Communication between the atmosphere and ocean is through the surface fluxes of heat, freshwater, and momentum, and knowledge of the atmosphere–ocean exchanges is critical for atmosphere–ocean modeling. Atmospheric general circulation models (GCMs) are developed using specified lower-boundary conditions [sea surface temperatures (SSTs), etc.], which provide a strong constraint on the modeled atmosphere. Ocean GCMs are developed with either a specified surface atmosphere or specified fluxes, and/or relaxation to observed surface temperatures and salinity. These provide a strong constraint on the ocean component. Thus, the component models may have systematic errors, and consequently coupled atmosphere–ocean models drift to a new model climate if left alone. Even small errors in one component model can be amplified by feedbacks. Widespread use has been made of unphysical ''flux correction'' to offset such drift and keep the modeled surface conditions similar to those observed. The fluxes corrected include heat, freshwater, and momentum (or a subset of these). However, the observational evidence to evaluate the models and the corrections is insufficient at present.

Aside from intensive field measurements using eddy correlation techniques, estimates of the surface fluxes are not direct in the sense that the fluxes are measured and analyzed into global gridded fields. Instead, estimates are made, as discussed further below, through (i) use of in situ observations along with empirical ''bulk'' formulas, (ii) use of a model atmosphere to interpret other specified inputs and compute surface fluxes, and (iii) various indirect methods, which rely on the fact that the surface heat, energy, and momentum budgets must balance and so, given computations of all the other components in the various budget equations applied to fields either within the ocean or the atmosphere, fluxes may be inferred as a residual. All of these methods suffer from difficulties, arising, for instance, from imperfect and uncertain formulas, sampling issues, model parameterizations, and accumulation of errors. Ideally, the strengths of all methods should be combined. If independent estimates are made of all the terms in a budget equation with an appropriate assignment of errors and the constraint requiring a balance is imposed, then errors must begin to cancel and optimal values may be assigned. The key difficulty seems to be the appropriate assignment of errors. In this paper, the focus is on the indirect approach, with special attention given to an assessment of error sources.

* The National Center for Atmospheric Research is sponsored by the National Science Foundation.

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2. Methods

a. Bulk methods

Atmosphere–ocean fluxes may be inferred using bulk methods and observed surface variables (e.g., Hastenrath 1980, 1982; Hsiung 1985; Isemer and Hasse 1987; Oberhuber 1988; da Silva and Levitus 1994a,b). From these methods, it is possible to obtain reasonable estimates of patterns of sensible and latent heat exchange, but substantial (several tens of W m^{-2}) systematic errors arising from the parameterizations, sampling, and uncertainties in the exchange coefficients are probable (e.g., Weare and Strub 1981; Weare 1989). Estimates of surface radiation depend heavily on cloud observations from ships, which are poor from the surface, and the various empirical formulas can err by up to 32% (Dobson and Smith 1988), so this component is quite uncertain. Estimates of surface wind stress also appear to be reasonable, but again systematic errors exist (e.g., Trenberth et al. 1990). All of these fluxes strongly depend on adequate sampling, and insufficient sampling exists over most of the southern ocean in particular (Trenberth et al. 1990).

b. Use of models

Atmospheric GCMs are used as part of four-dimensional data assimilation (4DDA) systems, and the model can be used along with the analyses to provide surface fluxes as a derived outcome from the model parameterizations. For some fluxes, this requires a short integration. The success of this method critically depends upon how good the model is and whether it is in balance with the observed fields or not. Commonly a ''spin-up'' occurs in the model, in which a violent adjustment takes place, such as in the divergence and/or moisture field, perhaps through a convective process, indicating an incompatibility between the observed and model-preferred states. Biases, such as manifested in climate drift or in systematic forecast error, would show up as errors in the derived fluxes. For example, it is vital to get accurate shortwave radiation at the surface, but this depends critically on the cloud diagnosed in the model, and errors of several tens of W m^{-2} are common (see, e.g., Gleckler et al. 1995).

c. Indirect methods

An alternative method relies on indirect estimates of the surface exchanges as a residual. For the heat budget, for instance, the total surface flux is a residual of the top-of-the-atmosphere satellite-observed net radiation and the divergence of the energy transports from global atmospheric analyses. Given the surface flux, ocean heat transports may be estimated. Or given ocean heat transports, an alternative estimate of the surface fluxes could be obtained, and then the two estimates would need to be reconciled. Application of both indirect methods clearly depends critically on the fidelity of the atmospheric or oceanic analyses; thus far, the latter are fragmentary and will not be considered further. Similarly, the freshwater flux may be obtained from the moisture budget in the atmosphere and the surface momentum exchange from the momentum equations. Here, we will focus on the energy and moisture equations, as the surface wind stress is more commonly estimated directly from measurements.

The annual mean atmospheric heat budget was first used to provide surface flux estimates and deduced ocean heat transports by Vonder Haar and Oort (1973), by Oort and Vonder Haar (1976) for the Northern Hemisphere, and Trenberth (1979) for the Southern Hemisphere. These studies and estimates by Carissimo et al. (1985) and Savijärvi (1988) made use of rawinsonde data, but the uncertainties in the atmospheric heat transports are substantial. At 70°S in the Carissimo et al. and Savijärvi results, for instance, where there is no ocean, the residuals imply a large poleward heat transport by the ocean. Problems are especially evident arising from estimates of atmospheric divergence in low latitudes in the rawinsonde-based analyses, but this aspect has improved in global analyses.

Thus, an alternative approach is to use state-of-the-art global analyses to assess the atmospheric heat transports (Boer 1986; Masuda 1988; Michaud and Derome 1991; Trenberth and Solomon 1994). The quality of atmospheric analyses produced from 4DDA systems has continued to improve in spite of observational losses (e.g., Trenberth and Olson 1988; Trenberth 1992, 1995a). Consequently it has been generally considered desirable to use the most recent analyses available to compute the atmospheric heat transports. This has changed as ''reanalyses'' become available.

Top-of-the-atmosphere (TOA) heat budget information derived from satellites is necessary to complete the atmospheric heat budget, and it is important to have data for the same months for the results to be valid locally (Trenberth and Solomon 1994).

d. Use of 4DDA analyses

The 4DDA analyses are produced by combining observations with a model forecast, which effectively carries forward in time all previous information. However, this tends to create imbalances within the model. In part, this is because of model biases or errors such as an incompatibility between the analyzed static stability and that preferred by the model convective parameterization. A spin-up of the model usually occurs in the subsequent forecast. As a result, the analysis increments applied each time an analysis is performed are unlikely to be random. In terms of the model compared with the analyses, for time averages, a balance exists between the dynamics, the physics, and the time-averaged increment (assuming that the overall tendency is negligible):

$$\overline{\text{dynamics}} = \overline{\text{physics}} + \overline{\Delta x},$$

where Δx is the increment and the overbar is the time average. It is highly desirable for the increment to be small. Note that the average increment is equivalent to the systematic short-term (6 h) forecast error of the 4DDA model.

The National Aeronautics and Space Administration (NASA)/Goddard reanalysis (Goddard Earth Observing System GE0S-1 Data Assimilation System, using the GEOS GCM) made estimates of the increment. As pointed out by Schubert et al. (1995), the residual approach assumes that the lhs is well known and estimates the total ''physics'' terms as the sum of the rhs. This assigns all the errors to the physics. However, the split is somewhat artificial, as the vertical motion terms, for instance, are included on the lhs despite being dependent on the physics of the model as they are diagnosed, not observed. Schubert et al. (1995) note that because the mean increments result from bias, interannual variability may be quite well known because it involves differences between years in which the bias is removed.

Kanamitsu and Saha (1995, 1996) have analyzed various atmospheric budgets using National Centers for Environmental Prediction (NCEP, formerly the National Meteorological Center) analyses and noted that the dynamics do not compensate the physical forcings, indicating an error in some physical forcings. Kanamitsu and Saha (1996) focused on the systematic errors, which means that in evaluating all the terms in the budget equations as a tendency, large local time tendencies result that do not match those observed. Nonzero tendencies can result from biased sampling, such as by not adequately sampling the diurnal cycle, but the source of the spurious imbalance is the time mean analysis increment and the biases in the model physics. These imbalances are found to be as large as the leading terms in the equations. Kanamitsu and Saha (1995) attempted to partition the systematic forecast error into various components.

For some quantities, it is possible to provide ground truth on sources of error. In particular, both Schubert et al. (1995) and Kanamitsu and Saha (1996) analyze the moisture budget over North America from the analyses and include model estimates of evapotranspiration E and precipitation P. The latter can be compared with measured rainfall. Results from both studies show that the systematic forecast error or analysis increment, expressed as a rainfall tendency, exceeds 2 mm day^{-1} over the United States in most months of the year, values comparable in magnitude to E or P. Errors in P of the same magnitude are present. For instance, in both studies, the dominant balance in the southeastern part of the United States is between the model precipitation and the analysis increment, indicating a spurious tendency for the models to produce rain and an inability to sustain the high levels of atmospheric moisture observed. It also seems that E may contain errors, for example in the western parts of the United States. In the NASA/Goddard analyses, both the model and analyzed P are small, but the E are not. Model evaporation into the atmosphere produces atmospheric moisture that is systematically removed by the 4DDA analysis increment. Presumably this relates to surface hydrology and excessive soil moisture availability.

The interpretation of the analysis increment, therefore, is that it is a much-needed adjustment for the accurate depiction of the state of the atmosphere and it arises in part from problems in the physics of the model. The analyses may not be fully accurate, especially in the representation of the divergent flow field, because the latter depends on the physics. But it therefore seems likely that the increment may be conservative. This would be the case, for instance, if there were an underestimate of precipitation by the model and, thus, an underestimate of latent heating and the associated divergent flow component. For a ''balanced'' set of analysis fields (e.g., from diabatic initialization or built into the analysis scheme), the analysis increment that includes the divergent flow would be underestimated. Then, inverting the process and deducing the physics as a residual would produce a similar underestimate. Nevertheless, it would be a much better estimate than that obtained from the model alone. Therefore, there seems to be considerable merit in the indirect approach to budgets, which use only the analyses to deduce the physics and forcing functions as residuals.

3. Budget equations

There are several formulations available of the different budgets, and, with perfect data, all should be equivalent. In practice, however, large differences arise from the differing formulations because of assumptions made about the data—such as the equation of continuity being satisfied—and it is desirable to explicitly spell out those assumptions to help understand the results and devise the best practical method for obtaining such estimates with minimal contamination from flaws in the data.

In using atmospheric data to evaluate budgets, it is important to realize the characteristics and shortcomings of the data. If fields are analyzed univariately, as is commonly the case using rawinsonde data, there may be no guarantee that the hydrostatic equation is satisfied and other dynamic constraints, such as the winds being close to geostrophic balance, can be grossly violated (Trenberth 1987). However, in that approach, the ω field is derived so that the equation of continuity is satisfied. Moreover, it is possible to introduce some dynamic contraints (Savijärvi 1988). Horizontal fluxes at stations can be accurately computed using rawinsonde data, but this information is difficult to transform into global analyses because of data gaps over the oceans. The global analyses from operational centers resulting from 4DDA have different characteristics. For instance, in the pressure-level archive from European Centre for Medium-Range Weather Forecasts (ECMWF), the fields are interpolated from model levels, and neither the hydrostatic equation nor mass conservation are exactly satisfied in the analyses (Trenberth 1991). The equation of continuity, in particular, can contain residuals of either term exceeding 50% (Trenberth 1991; Trenberth et al. 1995).

The global analyses are produced on model (sigma or hybrid) surfaces, which consist, in the simplest form, of a sigma (σ) terrain following coordinate in which the lowest level corresponds to $p = p_s$, where p is the pressure, p_s is the surface pressure, and $\sigma = p/p_s$. Hybrid η coordinates consist of σ near the surface, but with a gradual transition to pressure with height (often above about 100 mb).

a. Vertical integrals

The mass-weighted vertical integral of any quantity M integrated in the vertical over the mass of the atmosphere from the bottom ($p = p_s$, $z = 0$) to the top ($p = p_t = 0$, $z = \infty$), where, for practical reasons, it may often be necessary to recognize some other value than zero for p_t, is given by

$$\tilde{M} = \int_0^\infty \rho M \, dz = \frac{1}{g}\int_{p_t}^{p_s} M \, dp \qquad (1a)$$

and can be written as

$$\tilde{M} = \frac{1}{g}\int_{p_t}^{p_o} \beta M \, dp, \qquad (1b)$$

where p_o is now some fixed value of pressure greater than p_s everywhere, and we define $\beta = 0$, $p > p_s$ and $\beta = 1$, $p \leq p_s$. This has the potential advantage of having fixed integration limits (e.g., Boer 1982).

In hybrid model coordinates,

$$\tilde{M} = \frac{1}{g}\int_{\eta_s}^{\eta_t} \frac{\partial p}{\partial \eta} M \, d\eta, \qquad (2)$$

where η_t corresponds to $p = p_t$ and η_s corresponds to $p = p_s$. It is evident that $\partial p/\partial \eta$ plays the role of a density factor in these coordinates. In the event that σ is the vertical coordinate, this mass weighting factor becomes p_s.

In using pressure coordinates, substantial difficulties occur in dealing with the lower boundary and vertical integrals. These are discussed more fully in Trenberth and Solomon (1994). In utilizing ECMWF analyses, several approaches were tried to properly deal with the p_s values. The latter were computed for the real world orography (Trenberth 1992) and were also available from ECMWF for analyses on the artificial model surface, which corresponds to an enhanced envelope orography, often with p_s values lower by 50 to 100 mb or more in the vicinity of complex high mountains (Trenberth 1992). On p levels, therefore, values as archived are extrapolated below the envelope orography as part of the post processing at ECMWF. There is no "correct" way to do this, and our experience indicates that all such values are somewhat contaminated. It is therefore necessary to use the ECMWF p_s values and neglect that part of the real atmosphere below the artificial surface. Moreover, we have found that it is also desirable to ensure that no part of the extrapolated values enters into the calculations, or else extremely noisy fields result in the vicinity of mountains. Therefore, in all cases, the lowest layer was treated very conservatively by Trenberth and Solomon (1994) and sometimes omitted. This practical difficulty in obtaining reasonable results near complex terrain is a compelling reason why it may be better to carry out diagnostic analyses in model coordinates, but interpolating from p to model coordinates only compounds the problems, so it is necessary to obtain the original model-level data. Trenberth (1995b) outlines some difficulties that remain when using model coordinate data. For the diagnostics from p coordinates, it means that results cannot be considered accurate near mountains, and this aspect makes results over land less reliable than over the oceans.

b. Mass budget

In carrying out budgets of any atmospheric quantity, the values must be mass weighted, which places a premium on the requirement for the mass budget to be satisfied. Calculations of the mass budget using atmospheric data on pressure levels reveal that mass is not conserved by the analyses (Boer and Sargent 1985; Alexander and Schubert 1990; Trenberth 1991; Trenberth et al. 1995), and this lack of conservation constitutes a major error component in the other budgets. We have devised variational approaches to adjust the analyses so that mass is conserved in three dimensions (Trenberth et al. 1995). In the model coordinates in which the analyses are created, in the free atmosphere, an instantaneous balance can be assured by recomputing the vertical motion field using whatever numerical algorithms are being employed. However, the mass budget is not necessarily satisfied for vertically averaged and time-averaged data, and special attention is warranted for the surface pressure, with an adjustment, such as that given in Trenberth (1991), if necessary.

The general conservation equation is

$$\frac{\partial \tilde{M}}{\partial t} + \boldsymbol{\nabla}\cdot\frac{1}{g}\int_{p_t}^{p_s} M\mathbf{v} \, dp = S, \qquad (3)$$

where S is the source minus sink of M, vertically integrated in this column. This equation states that the flux of M out of the column is balanced either by a change in $\tilde{M}$ or by S.

The total mass of the atmosphere m in a column is

$$\tilde{m} = \tilde{m}_d + w = \frac{p_s}{g}, \qquad (4)$$

where $\tilde{m}_d$ is the mass of dry air and the precipitable water $\omega = (1/g)\int_0^{p_s} q\, dp$, where q is the specific humidity. Only the mass of dry air is conserved, so that $S = 0$ in that case.

Accordingly,

$$\frac{\partial \tilde{m}_d}{\partial t} + \nabla \cdot \frac{1}{g} \int_0^{p_s} (1 - q)\mathbf{v}\, dp = 0. \qquad (5)$$

For water vapor,

$$\frac{\partial w}{\partial t} + \nabla \cdot \int_0^{p_s} q\mathbf{v}\, dp = E - P, \qquad (6)$$

where E is the rate of evaporation and P is precipitation rate per unit mass, and we have ignored other forms of liquid and frozen water in the atmosphere. Combining these gives

$$\frac{\partial p_s}{\partial t} + \nabla \cdot \int_0^{p_s} \mathbf{v}\, dp = g(E - P). \qquad (7)$$

This equation is more accurate than that commonly used in meteorology, where the right-hand side is typically ignored. As shown by Trenberth (1991), such an approximation is usually justified to within about two orders of magnitude except when vertically integrated.

The commonly used equation of continuity is written as

$$\nabla \cdot \mathbf{v} + \frac{\partial \omega}{\partial p} = 0, \qquad (8)$$

although this ignores changes in moisture (Trenberth 1991). In model coordinates, this is

$$\frac{\partial}{\partial t}\left(\frac{\partial p}{\partial \eta}\right) + \nabla \cdot \mathbf{v}\left(\frac{\partial p}{\partial \eta}\right) + \frac{\partial}{\partial \eta}\left(\dot{\eta}\frac{\partial p}{\partial \eta}\right) = 0. \qquad (9)$$

The lower boundary condition is straightforward in model coordinates $\dot{\eta} = 0$, but much more involved in p coordinates $\omega_s = \partial p_s/\partial t + \mathbf{v}\cdot\nabla p_s$, at $p = p_s$. Vertically integrating (9), therefore, gives

$$\frac{\partial p_s}{\partial t} + \nabla \cdot \int_{\eta_s}^{\eta_t} \mathbf{v}\frac{\partial p}{\partial \eta}\, d\eta = 0. \qquad (10)$$

c. Moisture budget

The moisture budget is computed from (6) and contributes to the heat budget through latent heating, given by

$$\tilde{Q}_2 = L(P - E), \qquad (11)$$

where L is the latent heat of vaporization and Q_2 is the latent heat released through evaporation and condensation (Yanai et al. 1973). Trenberth and Guillemot (1995) recently evaluated the precipitable water and computed $E - P$ as a residual of the moisture budgets from the global analyses of the ECMWF, the NCEP, and NASA/Goddard, and compared results from 1987 to 1993, which are discussed later.

d. Energy budget

The intent is to obtain the diabatic heating in the atmosphere and use the observed top-of-the-atmosphere radiative fluxes to deduce the flux of heat through the bottom of the atmosphere. One such method uses the so-called apparent heat source Q_1 and the apparent moisture sink Q_2 arising from the thermodynamic and moisture equations (e.g., Yanai et al. 1973, 1976, 1992; Fortelius and Holopainen 1990). The term ''apparent'' is used because Q_1 and Q_2 include contributions from unresolved eddies. Other methods use the full energy equations in one form or another (e.g., Boer 1982, 1986; Boer and Sargent 1985).

Energy in the atmosphere is usually considered in the form of kinetic energy k, internal energy $I = c_v T$, and potential energy P_e (e.g., see Boer 1982). It is readily shown that $\mathbf{E} = I + k + P_e$, when integrated over the entire mass of the atmosphere, is conserved in the absence of heating and friction.

The vertically integrated P_e is given by

$$\tilde{P}_e = \int_0^{\infty} gz\rho\, dz = \frac{1}{g}\int_0^{p_s} \Phi\, dp$$
$$= \frac{1}{g}\int_0^{p_s} (RT + \Phi_s)\, dp = \frac{1}{g}\frac{R}{c_v}\tilde{I} + z_s p_s \qquad (12)$$

after integrating by parts and using the equation state and the hydrostatic equation, where Φ_s, the surface geopotential, is not a function of pressure and z_s is the surface geopotential height.

The kinetic energy equation is obtained by taking the dot product of the two horizontal equations of motion with $\mathbf{v}$, and converting from the advective form to flux form results in

$$\frac{\partial k}{\partial t} + \nabla \cdot k\mathbf{v} + \frac{\partial \omega k}{\partial p} = -\mathbf{v}\cdot\nabla\Phi + \mathbf{v}\cdot\mathbf{F}, \qquad (13)$$

where $k = 1/2(u^2 + v^2)$ and $\mathbf{F}$ is friction. The thermodynamic equation can be written as

$$c_p\left[\frac{\partial T}{\partial t} + \mathbf{v}\cdot\nabla T + \omega\left(\frac{\partial T}{\partial p} - \kappa\frac{T}{p}\right)\right] = Q_1, \qquad (14)$$

where Q_1 is the diabatic heating, $\kappa = R/c_p$, and c_p is the specific heat at constant pressure. Strictly speaking, κ, R, and c_p vary with the amount of moisture in the atmosphere, but it is a good approximation to treat them as constant for our purposes. Adding these gives the total dry energy equation

$$\frac{\partial}{\partial t}(c_p T + k) + \nabla \cdot (s + k)\mathbf{v} + \frac{\partial}{\partial p}(s + k)\omega$$
$$= Q_1 - Q_f, \qquad (15)$$

where $s = c_p T + \Phi$ is the dry static energy and Q_f is the frictional heating arising from dissipation of kinetic energy.[1] The rhs corresponds to the nonfrictional heating.

Vertically integrating and combining the dry and moist energy equations through the atmospheric column (see Trenberth and Solomon 1994 for details) gives

$$\frac{\partial}{\partial t}\frac{1}{g}\int_0^{p_s}(c_pT + k + Lq + \Phi_s)\,dp + \nabla\cdot\frac{1}{g}\int_0^{p_s}(h + k)\mathbf{v}\,dp = \tilde{Q}_1 - \tilde{Q}_2 - \tilde{Q}_f, \quad (16)$$

where $h = s + Lq$ is the moist static energy. Note that the form of (16) differs from (3) in that it is not simply the divergence of internal potential plus kinetic energy in the second term; the difference arises from the pressure work term in the thermodynamic equation.

The first term in (16) is the change in storage in the atmosphere of internal, potential, kinetic, and latent energy, the second term is the total atmospheric heat transport divergence. In Vonder Haar and Oort (1973) and Oort and Vonder Haar (1976), the k term is dropped from (16).

The vertical integral through the atmospheric column allows the heating to be broken up into components (see Fig. 1):

$$\tilde{Q}_1 - \tilde{Q}_f = R_T - R_s + H_s + \mathrm{LP}, \quad (17)$$

where R_T and R_s are the net downward radiation through the top of atmosphere and the earth's surface, H_s is the sensible heat flux through the surface, and P is the precipitation rate. The lhs corresponds to the nonfrictional heating. Thus,

$$\tilde{Q}_1 - \tilde{Q}_f = R_T + L(P - E) + F_s, \quad (18)$$

where

$$F_s = LE + H_s - R_s \quad (19)$$

is the net upward flux through the surface. Using (11),

$$\tilde{Q}_1 - \tilde{Q}_2 - \tilde{Q}_f = R_T + F_s. \quad (20)$$

On an annual mean basis, the changes in storage terms are small and often negligible, so that (16) may be written as

$$\nabla\cdot\mathbf{F}_A = R_T + F_s, \quad (21)$$

where $\mathbf{F}_A = (1/g)\int_0^{p_s}(h + k)\mathbf{v}\,dp$ is the atmospheric energy transport. Over land, ignoring heat stored in the ground, F_s should be close to zero for the annual mean, so

$$\nabla\cdot\mathbf{F}_A^{\mathrm{land}} = R_T \quad (22)$$

and the balance lies between the atmospheric energy divergence and the net top-of-the-atmosphere radiation. Within the ocean, again ignoring heat storage changes,

$$\nabla\cdot\mathbf{F}_O = -F_s, \quad (23)$$

where $\mathbf{F}_O$ is the vertically integrated divergent ocean heat flux. In principle, therefore, F_s can be derived from either (21) or (23).

Given estimates of R_T from satellite measurements and using computed values for terms on the left-hand side of (21) or (16), F_s can be estimated as a residual. Over the oceans, (23) then allows estimates to be made of the ocean heat transport. Moreover, F_s can be compared with independent estimates made using bulk flux formulations of the surface fluxes.

An attraction of vertically averaged budgets is the apparent absence of the difficult-to-calculate terms involving vertical motions. The fluxes are critically dependent on the divergent wind. Most commonly, the approach outlined above has not been followed locally, and instead (21) and (23) have been zonally averaged to obtain total meridional ocean heat transport as residuals (Vonder Haar and Oort 1973; Oort and Vonder Haar 1976; Trenberth 1979; Masuda 1988; Carissimo et al. 1985; Savijärvi 1988; Michaud and Derome 1991; Keith 1995). This assumes that (22) holds, and nearly all studies have made this assumption. However, (22) is often violated by several tens of W m^{-2} (Trenberth and So-

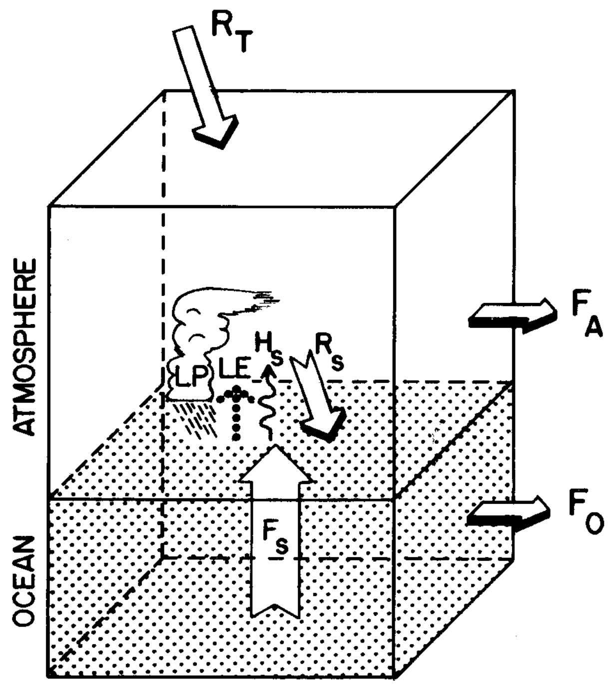

FIG. 1. Schematic of the balance of energy fluxes within the atmosphere, ocean, and interface. Here, R_T and R_s are radiative fluxes at the TOA and surface, respectively; H_s is the surface sensible heat flux; LE is the latent heat loss from the ocean; LP is the latent heat gain in the atmosphere through precipitation; F_s is the net surface flux from ocean to atmosphere; and F_A and F_O are horizontal fluxes within the atmosphere and ocean, respectively.

[1] Here, Q_1 includes a contribution from Q_f that was incorrectly handled in Trenberth and Solomon (1994); it was ignored as small rather than cancelled.

lomon 1994), and a spurious residual may indicate how well this method is actually performing. The implication is that some of the "ocean" transport is occurring over land. Equation (23) should be solved for the ocean region only with appropriate boundary conditions, and this was done by Trenberth and Solomon (1994). It is therefore desirable to examine F_s and F_O regionally much more carefully and critically than has generally been the case in the past.

4. Top-of-the-atmosphere radiation

A discussion of these datasets is given by Hurrell and Campbell (1992). Estimates of errors from satellite measurements of top-of-the-atmosphere radiation from the Earth Radiation Budget Experiment (ERBE) were reviewed by Trenberth and Solomon (1994) as giving an rms uncertainty estimate of 7.8 W m^{-2} for the three satellite combination versus 11 W m^{-2} for one satellite, with larger uncertainty in the absorbed solar radiation (ASR).

We have recently reprocessed the ERBE record from February 1985 to April 1989 to accommodate the discontinuity that occurred with the loss of *NOAA-9* in January 1987, to correct for the global mean imbalance, and to fill in missing data. While ERBE results are believed to be the most definitive on the TOA radiation, they cover only a limited period. Much longer time series are available from *Nimbus-7,* which allows estimates to be made of the sampling variability associated with interannual variability (Ardunuy et al. 1992; Kyle et al. 1993). The annual mean global net TOA radiation has varied from late 1978 to 1986 by about 1 W m^{-2}, which could partly be associated with the buildup of greenhouse gases in the atmosphere which may partly be associated with changes in heat storage within the the climate system, such as those associated with El Niño events.

The missing data all occur for the ASR and, thus, for net radiation near the delimiter of the incoming radiation, where it is difficult to obtain an accurate albedo. To fill the missing data, we firstly took advantage of the fact that the areas missing varied from year to year, and we formed a climatology of the available data on albedo. A least squares fitted first harmonic was then derived for each point, and the missing climatological mean values determined. These were used to fill in the missing albedo. Finally, a nine-point smoother was applied to the replacement data and to data within two grid points of a replacement value. The ASR and net radiation were then derived. The mean annual cycle probably involves more than a single harmonic, but use of the least squares fitting approach with more than one harmonic could occasionally result in albedos for missing points that exceeded unity. Therefore, the approach used is conservative, but it produces quite reasonable numbers. Because the missing points are weighted by small incoming radiation, the impact on the ASR is not very great, but it is highly desirable to do this step rather than treat the data as missing.

Because of changes in satellites during ERBE, it is difficult to estimate the true interannual variations in net radiation, Consequently, we ensure that the annual net radiation is constrained to be $<\pm0.5$ W m^{-2}, consistent with the *Nimbus-7* results. For the first 2 yrs of ERBE data, the global mean outgoing longwave radiation (OLR) mean was 233.9 W m^{-2}, but it jumped to 236.5 W m^{-2} after the loss of *NOAA-9.* Presuming that the values with three satellites are superior, we have adjusted the OLR everywhere uniformly downward from February 1987 on, justifying this as a bias most likely arising from the inadequate sampling of the diurnal cycle. The net imbalance in annual mean net radiation was initially 4.2–6.0 W m^{-2}, and after the first adjustment to OLR, this was corrected for by applying a decrease in the albedo uniformly such that the 12-month running mean radiation budget balances. At the beginning and end of the ERBE record, the 12-month mean is one-sided rather than centered. Note that each month is adjusted separately, rather than using the 12-month mean. Because the adjustment is to albedo, the adjustment in the ASR is nonuniform in latitude and, instead, is greatest where the radiation is largest, consistent with the view that the imbalance is most likely associated with sampling of the diurnal cycle. Following these two fairly minor adjustments, the 12-month running mean net radiation for the revised ERBE data for most months is close to, but not exactly, zero (ranging from -0.14 W m^{-2} to $+0.14$ W m^{-2}).

The adjusted ERBE dataset is available from the National Center for Atmospheric Research (NCAR) (see the World Wide Web at http://www.cgd.ucar.edu/cas/catalog/).

5. Examples

a. Mass budget

Previous results have revealed the effects of the large and spurious changes in operational analysis systems at ECMWF and NCEP (see Trenberth 1995a), but these should be much improved in the reanalyses. As shown in Trenberth (1991), the order of the tendency term in (7) is 10×10^{-4} Pa s^{-1}, and the rhs contribution from changes in moisture in the column, for $E - P$ rates of $\sim$5 mm day^{-1}, would be 5×10^{-4} Pa s^{-1}. However, spurious mass tendencies are commonly 5×10^{-3} Pa s^{-1}, which corresponds to $\sim$130 mb month^{-1}. From (16), this would contribute to the heat budget an amount on the order of $(1/p_s) \tilde{h}\, \nabla \cdot \int_0^{p_s} \mathbf{v}\, dp$. Values of the vertical average of $h = (g/p_s\tilde{h}$ given in Fig. 2 for July 1985 and January 1986, range from about 300 to 350×10^3 J kg^{-1}. Therefore, a mass residual of 5×10^{-3} Pa s^{-1} translates to 150 to 175 W m^{-2} in the heat or energy budget.

We have examined the mass budget with reanalyses from NCEP using model coordinate data and evaluated

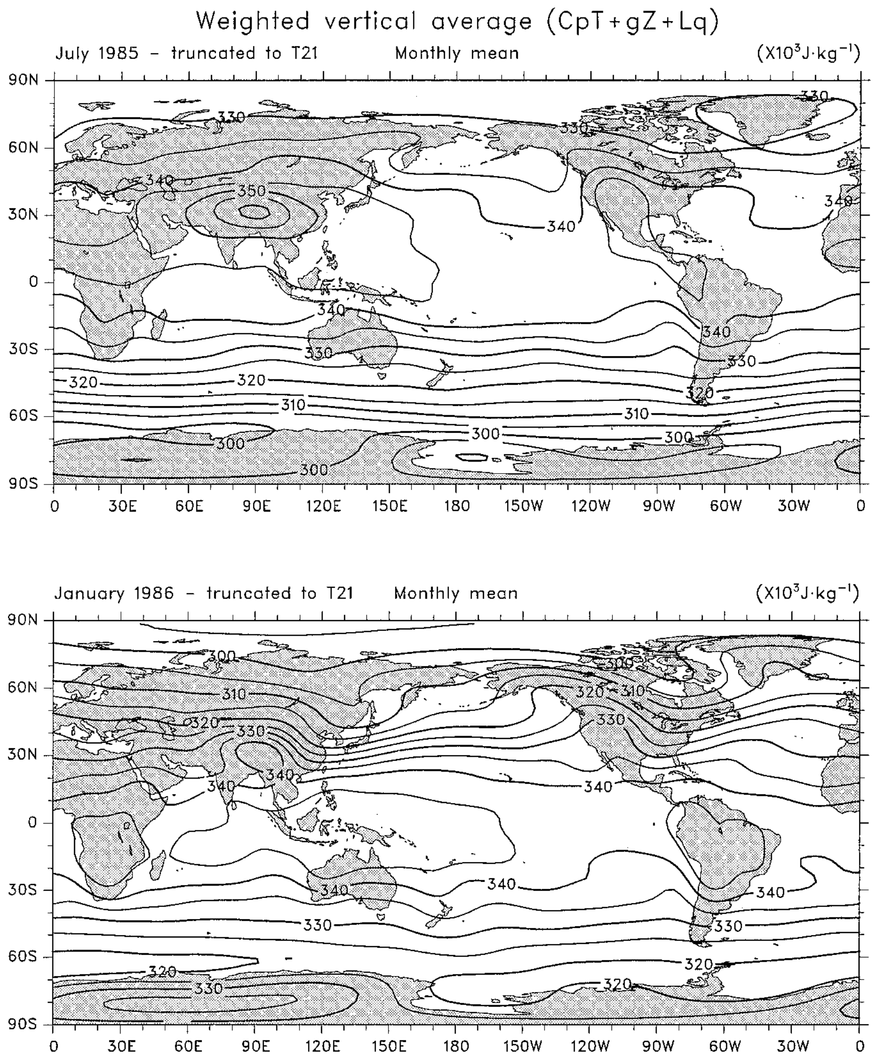

FIG. 2. The vertically averaged moist static energy for July 1985 and January 1986 from NCEP reanalyses. Units are 10^5 J kg^{-1}.

both the sum of the terms on the lhs of (7) [or equivalently (10)], the rhs of (7), and the residual for the entire equation. The latter is shown in Fig. 3 for July 1985 for all four times of day of the analyses. Figure 4 shows the results for July 1985 and January 1986 for the complete month sampled four-times daily, while Fig. 5 shows the corresponding rhs terms from the moisture budget. The latter is an order of magnitude smaller, so the lhs of (7) is indeed very similar to Fig. 3. As noted above, the magnitudes of these residuals are large, several orders of magnitude larger than accountable for by changes in moisture in the column and large enough to create errors of hundreds of W m^{-2} in heat and energy budgets unless corrected for.

Figure 3 clearly reveals the influence of the diurnal and especially the semidiurnal tide on the analyses with an amplitude of $\sim 20 \times 10^{-3}$ Pa s^{-1} in a large-scale wave 2 pattern. There is no evidence of such a pattern when all four-times-daily analyses are taken into account (Fig. 4), showing the need for data at least every 6 h to resolve such patterns and avoid aliasing. Analyses based solely on conventional once- or twice-daily data potentially contain a major spurious influence of the semidiurnal tide on results. The semidiurnal tide in these NCEP analyses is similar in magnitude to that in the operational ECMWF analyses and to observations of surface pressure (as summarized by Trenberth 1991; see also Trenberth et al. 1995). The observed amplitude of the semidiurnal tide throughout the Tropics is 1 to 1.4 mb, which translates to 17 to 25 $\times$ 10^{-3} Pa s^{-1} for the surface pressure tendency amplitude, consistent with Fig. 3.

b. Moisture budget

Residual computations of $E - P$ from NCEP and ECMWF operational analyses from 1987 to 1993 have

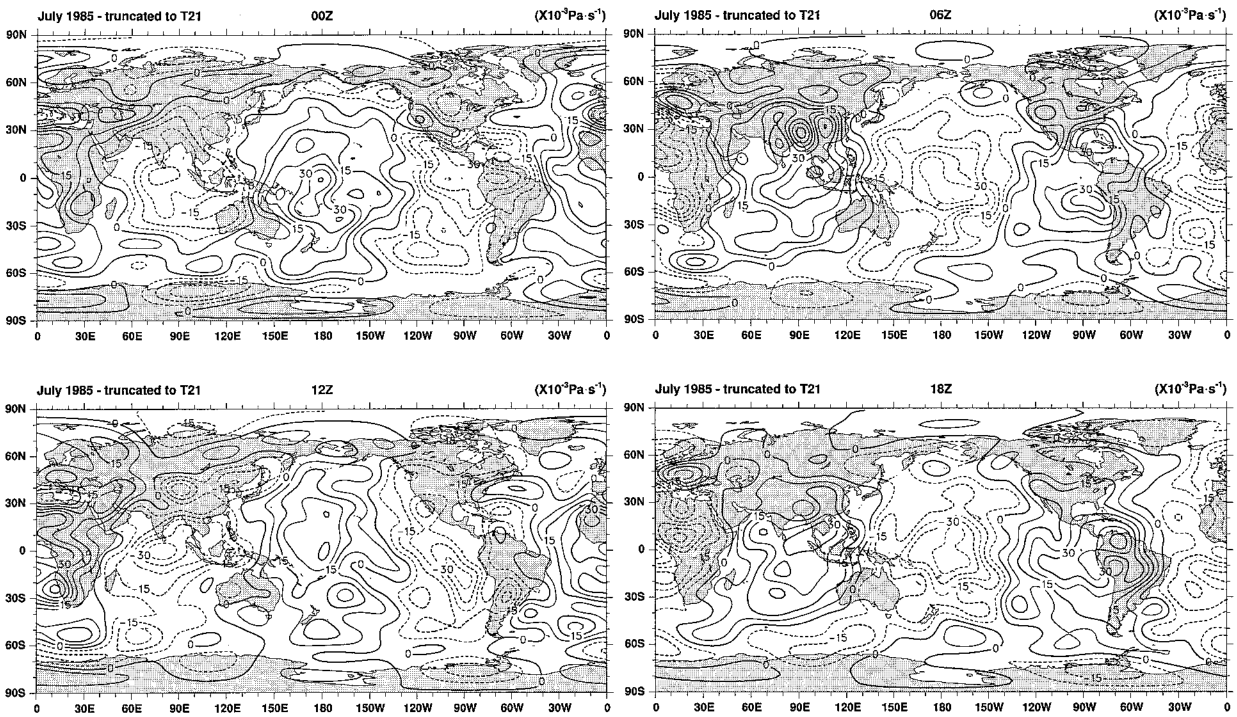

FIG. 3. The mass budget residual of (10) expressed as a surface pressure tendency in 10^{-3} Pa s^{-1} for July 1985 from NCEP reanalyses for 0000, 0600, 1200, and 1800 UTC. The fields have been truncated to T21 resolution.

been made by Trenberth and Guillemot (1995), who noted the problems with the results over land where it is not physically possible for E to be greater than P for an annual mean. This is one indication of remaining difficulties with the moisture budget. Trenberth and Guillemot evaluated the moisture budget by comparing NCEP and ECMWF results with other sources of information, including satellite data from the Special Sensor Microwave Imager precipitable water, reanalyses from NASA/Goddard, and precipitation from the Global Precipitation Climatology Project. In the subtropics, large positive, biases are present in precipitable water in the ECMWF and, to a lesser extent, the NCEP analyses, although they are partly due to the assimilation of biased Television Infrared Observation Satellite (TIROS) Operational Vertical Sounder (TOVS) retrievals. Over regions of good data, such as North America, the agreement in the moisture budgets is reasonable. But over the oceans, the disagreements are large and locally are of almost the same order as the terms themselves. The effects of the large and spurious changes in analysis systems at ECMWF and NCEP are manifested in the results. Discrepancies in the Tropics in $L(E - P)$ for the root-mean-square zonal mean discrepancies are 70–90 W m^{-2} (or for $E - P \sim 3$ mm day^{-1}), compared with rms values of 120 W m^{-2}—that is, 60% to 75% of the values themselves in the Tropics at T31 resolution (and about half as much at T15 resolution). These discrepancies are present throughout the 1987 to 1993 period and, if anything, get slightly larger with time (Trenberth and Guillemot 1995). Fortunately the uncertainties in the total heat budget are much less because of the cancellation between Q_1 and Q_2 as the moisture convergence is realized as latent heating.

The units of Fig. 5 translate into mm day^{-1} for $E - P$ when the values are multiplied by ~ 0.9. We have compared this result with corresponding values from ECMWF (as computed by Trenberth and Guillemot 1995), and there are strong similarities, but also notable differences, consistent with those found between recent ECMWF and NCEP operational analyses.

c. Energy budget

Because of difficulties with analyses, only one study thus far has done a heat budget calculation locally using consistent data at the top-of-the-atmosphere and within the atmosphere—that of Trenberth and Solomon (1994). The 1988 annual mean surface flux, as adjusted by Trenberth and Solomon over the region south of 30°S to give transports that go to zero at the ice edge, is shown in Fig. 6. Surpluses of heat into the ocean in the eastern tropical Pacific of over 120 W m^{-2} and in other parts of the Tropics are qualitatively in agreement with bulk flux computations. The main heat fluxes into the atmosphere from the ocean are off the east coast of Asia and North America, and those take place mainly in winter. Qualitatively this agrees with results from bulk flux computaions such as from Oberhuber (1988) and da Silva and Levitus (1994a).

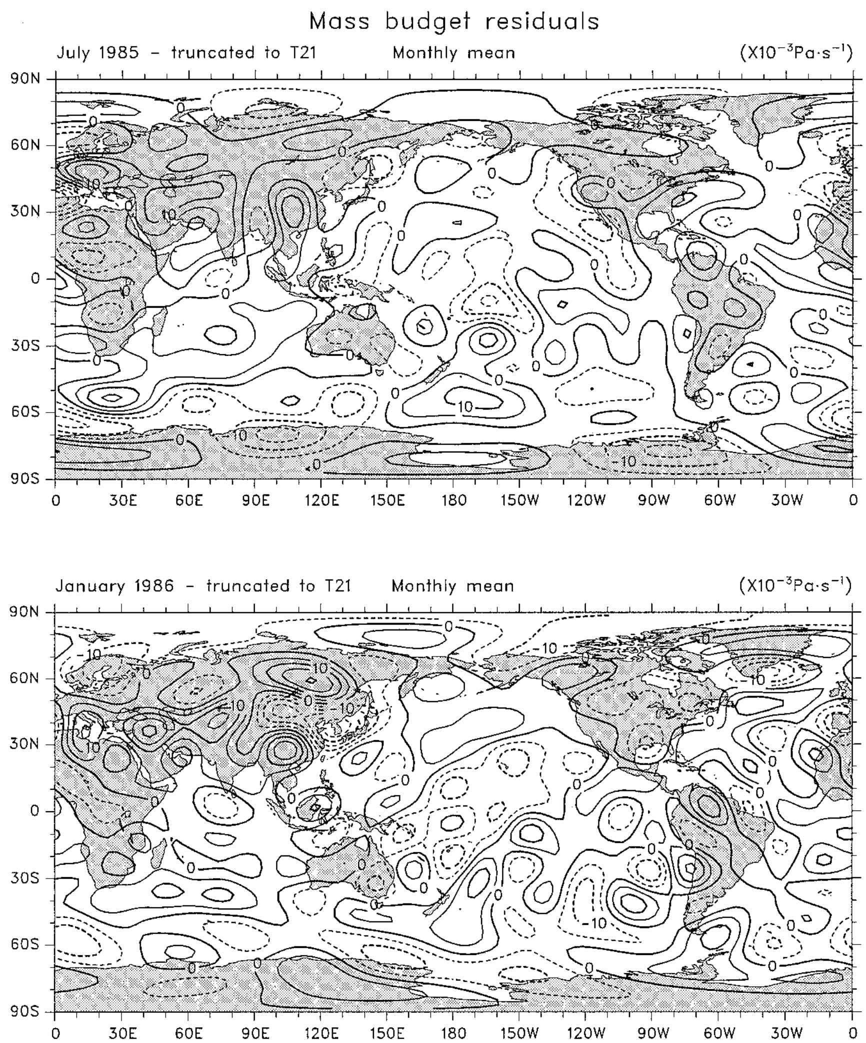

FIG. 4. The mass budget residual of (10) expressed as a surface pressure tendency in 10^{-3} Pa s^{-1} for July 1985 and January 1986 from four-times-daily NCEP reanalyses. The fields have been truncated to T21 resolution. Note that 4×10^{-4} Pa s^{-1} corresponds to about 10 mb $month^{-1}$.

The net ocean heat transport locally resulting from this is given in Fig. 7 (from Trenberth and Solomon 1994). Note that this is only the divergent component of the heat transport; it does not include a rotational component such as may occur in gyres or the Antarctic Circumpolar Current. Trenberth and Solomon also provide results for the individual ocean basins. Results are encouraging, as can be seen for the Atlantic Ocean (Fig. 8), where poleward heat fluxes from that study are compared with results from an ocean model that incorporates a new mixing parameterization. Error bars are estimated as about ± 0.2 PW at 20°N (Trenberth and Solomon 1994). For the first time, with improvements in diagnostic techniques and models, oceanographic estimates and atmospheric estimates have come together to give reasonable agreement.

It is not clear how well the overall local heat budget is known; however, the results over land indicate that the uncertainty is on the order of ± 30 W m^{-2} on scales of 1000 km or so, where this should be regarded as a standard error.

6. Heat and moisture budget sources of errors

The following discussion of errors draws heavily on the analyses and comparisons performed by Trenberth and Solomon (1994) for the heat and energy budgets and those of Trenberth and Guillemot (1995) for the moisture budgets.

a. Top-of-the-atmosphere radiation

Although there are errors in the radiation measurements of significance, the estimate is that the errors bars are ± 5 W m^{-2} after corrections for the overall imbalances. The uncertainties in the various components of

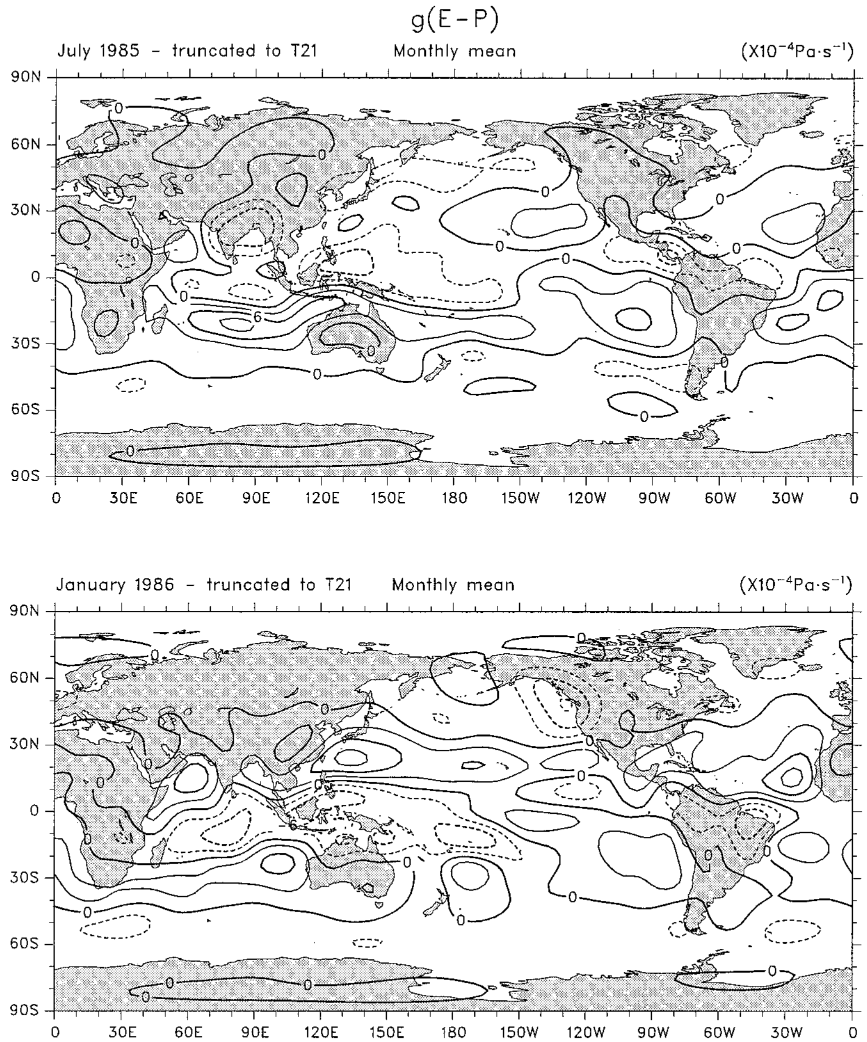

FIG. 5. The $g(E - P)$ from the NCEP reanalyses for July 1985 and January 1986 in 10^{-4} Pa s^{-1}.

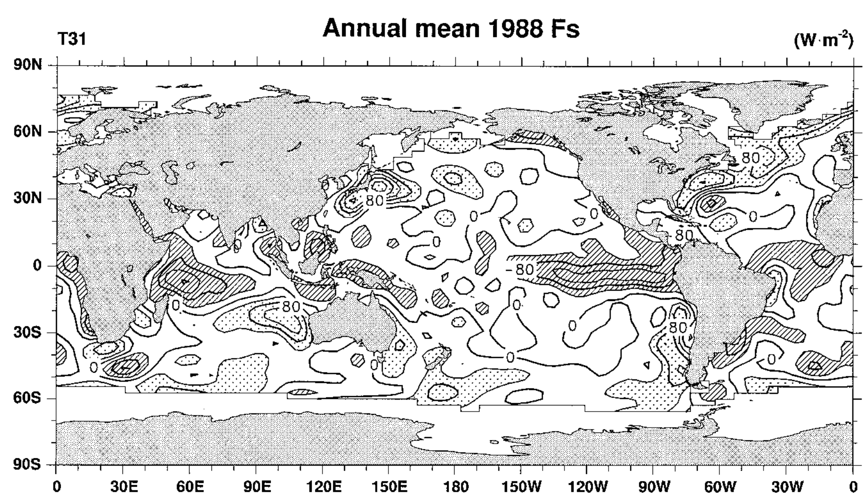

FIG. 6. The net surface energy flux F_s for 1988 in W m^{-2}, adjusted as described in Trenberth and Solomon (1994) south of 30°S to give zero transport at the southern ice edge. The contour interval is 40 W m^{-2}.

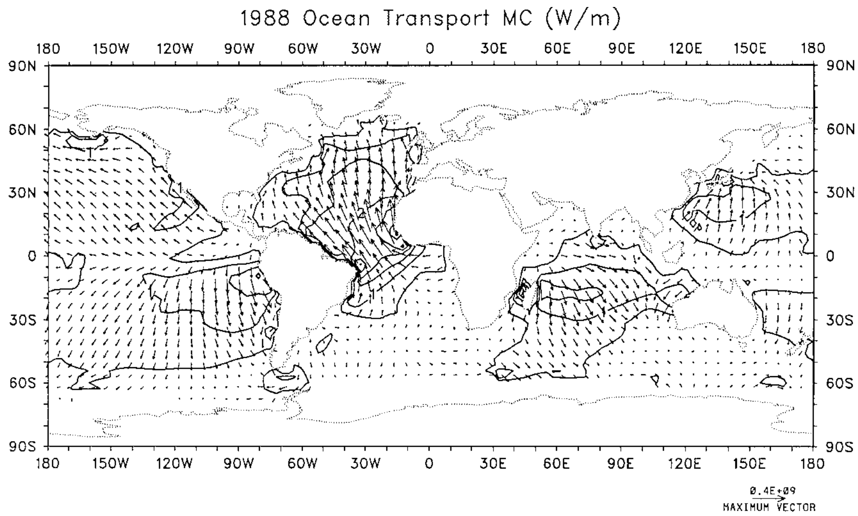

FIG. 7. The vectors of divergent ocean heat transport for 1988 from solving (20) using the field in Fig. 6. The simplified coastal outlines are indicated. Units are 10^8 W m^{-1}, and the vector for 4×10^8 W m^{-1} is indicated at the lower right. The contours indicate the magnitude of the vectors. From Trenberth and Solomon (1994).

the atmospheric transports, although difficult to quantify, probably dwarf these numbers.

b. Atmospheric divergence

In low latitudes, the estimates of atmospheric divergence and the associated large-scale overturning in the Hadley and Walker circulations were especially poor in earlier rawinsonde-based analyses, but have been improving steadily in global analyses. Changes in the strength of the analyzed Hadley and Walker circulations with changes in 4DDA systems are common, and the extent of the spinup in precipitation in the model employed in the 4DDA has also changed (see especially Trenberth 1992, 1995b). As shown by Trenberth (1992), prior to May 1989, the mass flux in the Hadley cell in the ECMWF operational analyses in July was about 200 $\times 10^9$ kg s^{-1}, but this increased to over 280 units after changes were made in May 1989, whereas the NCEP reanalysis retains the strength of the Hadley cell at about 160×10^9 kg s^{-1}. Sardeshmukh (1993) has shown the inconsistency of the ''analyzed'' divergence field with the vorticity budget and shown that it was too weak for the period 1982–85, in particular. Sardeshmukh and Liebmann (1993) have further demonstrated the inconsistencies between the ECMWF and NCEP divergence fields for 1988–89. Diabatic initialization of the analyzed fields imposes a somewhat artificial balance that may overstrengthen zonal mean meridional circulations (Errico and Rasch 1988), although it tends to reduce the divergence locally.

Because there is strong cancellation between the Q_1 and Q_2 contributions, the net effect of uncertainties in the tropical divergence on the total heat budget is somewhat ameliorated (Trenberth and Solomon 1994). There has been a clear improvement with time in the tropical analyses, although the discrepancies between NCEP and ECMWF remain very large. The critical dependence of the moisture budget on the the veracity of the velocity field and, specifically, on the horizontal divergence in the Tropics has been shown by Trenberth and Guillemot (1995). In the midlatitudes, quasigeostrophic dynamics ensure that the divergence field is better known, and uncertainties in the moisture budget stem roughly equally from discrepancies in moisture analyses and the velocity field.

c. The diurnal cycle

Most previous studies have not resolved the diurnal cycle, and once- or twice-daily analyses have been used. However, four-times-daily analyses mostly capture the diurnal cycle. A major part of the diurnal cycle and its semidiurnal component is probably real (see section 5); however, part of it arises because observations are taken mostly at 0000 and 1200 UTC and in many places only once per day, so that spurious components are introduced by the distribution of observations (see Trenberth and Solomon 1994; Trenberth 1995b). Because of the semidiurnal component, twice daily analyses contain a strong residual diurnal cycle component. For moisture, the diurnal cycle is not well captured by twice-daily data, and, while not so important for w, it makes significant differences for $E - P$. Therefore, none of the atmospheric budgets, and in particular the mass, moisture, and heat budgets, can be considered adequately determined unless the diurnal cycle is fully incorporated.

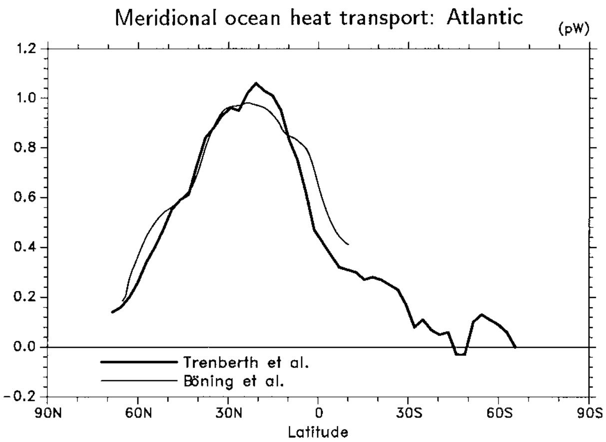

FIG. 8. Northward transport of heat in the Atlantic Ocean north of 10°S for a 1° ocean model (Böning et al. 1995), calculated by the stand-alone version of the ocean model component of the NCAR Climate System Model that uses the isopycnal mixing scheme of Gent and McWilliams (1990), compared with the observational patterns diagnosed by Trenberth and Solomon (1994).

d. Mass imbalances remain in all analyses

Mass imbalances are quite large (cf. Fig. 4) and will totally distort other budgets unless corrected for, although the impact is small on the moisture budget. Fortunately first-order corrections are relatively straightforward to do (Trenberth 1991), and this correction also removes most of the problems of not adequately resolving the diurnal cycle. However, the correction is to the vertical integral and does not take account of the vertical distribution of the correction (see Trenberth et al. 1995). The mass imbalance is much worse in the uninitialized analyses (Trenberth 1991). Vertical motions and divergence fields tend to be stronger but noisier in the uninitialized analyses.

e. Vertical resolution

Improved vertical resolution is important, although our calculations show that the differences are relatively small compared with the other items mentioned above. In particular, improved resolution is needed in the boundary layer for the moisture budget, and in the boundary layer and upper troposphere for divergence. Biases exist in *w* with vertical resolution, and vertical interpolation of *q* rather than RH tends to overestimate the moisture content of layers. However, the errors are systematic and do not have much effect on the moisture budget (Trenberth and Guillemot 1995).

f. Interpolation to p surfaces

The archives of global atmospheric analyses are generally made available in *p* coordinates. The archival postprocessing of interpolating from model coordinates, on which the analyses are performed, to *p* levels introduces errors (Trenberth 1991) and results, for instance, in the equation of continuity no longer being satisfied. Earlier, in section 3a, the difficulty in properly dealing with the lower boundary of the atmosphere in *p* coordinates was discussed. Discrepancies exist between the true surface of the earth and that depicted in the global analyses, and problems are apparent near mountains. It may be worthwhile to use model coordinates for future diagnostic analyses, although these too are not without substantial difficulties. In addition to the problem of surface depiction, the process of changing resolution (e.g., from T106 to T42) to facilitate processing large amounts of data is ill defined in model coordinates (Trenberth et al. 1993; Trenberth 1995b). Equations in terrain following coordinates have additional terms, making interpretation more cumbersome, and operations such as averages or decomposing winds into rotational and divergent parts take on different meanings on such coordinates. Model coordinates appear to have their main advantage in dealing with vertically integrated budgets.

7. Future plans

Plans are underway to carry out more comprehensive studies using the products of the reanalysis efforts and model coordinates, thus avoiding the problems associated with the lower boundary in *p* coordinates. Large discrepancies between results from different centers can still be expected for both the heat and moisture budgets, however, because of the strong dependency on moist physics in the assimilating model of the 4DDA and uncertainties related to the divergence field in the atmosphere.

Acknowledgments. I thank Amy Solomon and Chris Guillemot for help with the figures. This research is partly sponsored by NOAA under Grant NA56GP0247 and by NASA under NASA Order W-18,077.

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