CSE Water and Energy Budgets in the NCEP–DOE Reanalysis II

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ABSTRACT

During the past several years, the Global Energy and Water Cycle Experiment (GEWEX) continental-scale experiments (CSEs) have started to develop regional hydroclimatological datasets and water and energy budget studies (WEBS). To provide some global background for these regional experiments, the authors describe vertically integrated global and regional water and energy budgets from the National Centers for Environmental Prediction (NCEP)–U.S. Department of Energy (DOE) Reanalysis II (NCEPRII). It is shown that maintaining the NCEPRII close to observations requires some nudging to the short-range model forecast, and this nudging is an important component of analysis budgets. Still, to first order one can discern important hydroclimatological mechanisms in the reanalysis. For example, during summer, atmospheric water vapor, precipitation, evaporation, and surface and atmospheric radiative heating all increase, while the dry static energy convergence decreases almost everywhere over the land regions. One can further distinguish differences between hydrologic cycles in midlatitudes and monsoon regions. The monsoon hydrologic cycle shows increased moisture convergence, soil moisture, and runoff, but decreased sensible heating with increasing surface temperature. The midlatitude hydrologic cycle, on the other hand, shows decreased moisture convergence and surface water, and increased sensible heating.

1. Introduction

In the past, satellite observations of hydroclimatological variables were evaluated at a few limited validation sites and then extrapolated globally. Researchers would spend a lifetime studying a single hydroclimatological process or variable. Hydrologists would focus mainly on processes in small catchment basins. Atmospheric scientists did not fully appreciate the importance of continental-scale coupled land–atmosphere interactions. Recognizing the complexity of this extrapolation problem as well as the synergy of internationally coordinated hydrometeorological research at large and small scales, the World Climate Research Program’s Global Energy and Water Cycle Experiment (GEWEX) helped to develop five regional continental-scale experiments (CSEs), as well as an affiliated experiment over Africa, in order to scale up remote sensing and regional hydrology and meteorology to an eventual global hydroclimatological analyses.

The CSEs and affiliated experiments are loosely coordinated by the GEWEX Hydrometeorology Panel (GHP), which, as shown in Fig. 1, now includes nine representative world climate regions. The regions over the Americas include the Mackenzie (K; see Stewart et al. 1998), Mississippi (M; see Lawford 1999), and Amazon (A; see Marengo et al. 2002, manuscript submitted to J. Hydrometeor., hereafter MNS) river basins. In Europe, there is an experiment for the Baltic Sea [BALTEx (B); see Raschke et al. 1998] and four GEWEX Asian Monsoon Experiment (GAME; see GAME International Science Panel 1998) sites over the Lena (L) river basin, Huaihe River Basin Experiment [HUBEx (H)], Tibet (E), and tropical (T), regions. An affiliated experiment, the Coupling Tropical Atmosphere and Hydrological Cycle project (CATCH; see D’Amato and Lebel 1998), has begun over western equatorial Africa. About half of the CSEs are major river basins (Mackenzie, Mississippi, Amazon, Lena), one is an inland sea (Baltic), and the rest cover large-scale regions (CATCH, GAMEHUBEx, GAME-Tibet, GAME-Tropics).

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Fig. 1. Location of GEWEX CSEs, denoted by a single letter: Over the Americas are the Mackenzie River basin (K), Mississippi River basin (M), Amazon River basin (A). In Europe there is the BALTEX region (B), and in Africa there is the CATCH region (C). In Asia there are the four GAME locations [Lena River basin (L), Tibet (E), HUBEX (H), and Tropics (T)]. (River basin locations provided by N. Miller, LBL, and various CSE representatives.)

A major goal of the CSEs has been to develop datasets and model simulations that describe regional water and energy budget studies (WEBS) as a prelude to simulating and eventually predicting these budgets with regional and global climate models. To provide some general background for these regional experiments, we describe here the regional budgets from the National Centers for Environmental Prediction (NCEP)—U.S. Department of Energy (DOE) reanalyses II (NCEPRII), augmented by calculations for the water and energy fluxes from the NCEP—National Center for Atmospheric Research (NCAR) original reanalysis (NCEPR). In fact, many of the CSEs are examining various global reanalyses in order to understand what improvements in the global analyses and satellite observations are necessary for describing regional and eventually predicting global water and energy cycles (see, e.g., Trenberth and Guillemot 1994; Trenberth 1999; Higgins et al. 1996; Roads et al. 1992, 1994, 1997, 1998a, 1999; Roads and Betts 2000; Heise 1996; Maurer et al. 2001a, hereafter MOLRa; Maurer et al. 2001b, hereafter MOLRb; Oglesby et al. 2001; MNS; Strong et al. 2002). Most of these studies have concentrated upon the water mass equations. Here we also describe corresponding atmospheric and surface energetics.

As will be shown, the CSEs cover a range of land hydroclimates, ranging from the cold and dry high-latitude regions associated with the Mackenzie River basin in Canada and the Lena River basin in Siberia to the warm and moist tropical regions associated with the Amazon and GAME-Tropics regions in Southeast Asia. To a certain degree, all of the CSEs show a number of similar seasonal variations in the coupled land—atmosphere water and energy cycles. However, there are some differences between the Tropics and middle latitudes.

NCEPR and NCEPRII are described in section 2. We then discuss, in section 3, the basic equations of the water and energy cycles and also provide further details in the appendix about how these hydrometeorological processes are calculated from the analyses. In section 4, we discuss variations in the annual means of the water and energy cycle with respect to surface temperature. Geographic variations in the annual and seasonal means are then discussed in section 5. In section 6, we further discuss the seasonal mean basin averages. Some of the interannual variations are presented in section 7, and section 8 summarizes the results.

2. NCEPR and NCEPRII

Numerical weather prediction analyses attempt to use available data to constrain short-term model forecasts with a numerical weather prediction model to produce the best possible short-term forecast of various processes. These short-term forecasts remain close to available input observations, in some sense, and become even closer to available observations as the model’s physical and dynamical parameterizations better emulate the way nature operates. However, we must not expect too much yet. Analysis budget quantities are inferred through still quite imperfect model parameterizations. The analysis constraints of the model to observations can make significant but artificial contributions to water and energy
The GSM uses a simplified Arakawa–Schubert (SAS) parameterization (Arakawa and Shubert 1974), which was developed for the GSM by Pan and Wu (1995) following the Grell (1993) simplification. SAS removes large-scale instabilities by relaxing temperature and moisture profiles toward prescribed equilibrium values on a prescribed timescale. The SAS scheme also allows entrainment into the updraft and detrainment from the downdraft between the level from which the updraft air originates and the level of free convection (LFC). The level of maximum moist static energy between the surface and 400 hPa is used as the level from which the updraft air originates. Convection is suppressed when the distance between the updraft-air originating level and the LFC exceeds a certain threshold (150 hPa). Cloud top is determined as the first neutral level above cloud base.

NCEPR II (see Kanamitsu et al. 2002) used an updated version (1999 NCEP MRF) of the original GSM and, in addition to a few physical parameterization changes, a number of notable bugs in NCEPR were also fixed. For example, snow amount is now prescribed from operational files instead of using a fixed climatology (which was mistakenly used in NCEPR for a few years). Horizontal diffusion is correctly applied to pressure surfaces, rather than sigma surfaces, which results in better diffusion at high latitudes, and less spectral noise is now apparent in the precipitation and snowfields (see Roads et al. 1999; Serreze and Hurst 2000). The radiation is now computed on the full model grid instead of a coarser grid. The cloudiness–relative humidity relationship was refined. There are some other important differences in the boundary layer. In the NCEPR boundary layer, vertical transfer is related to eddy diffusion coefficients dependent upon a Richardson number–dependent diffusion process (Kanamitsu 1989). In the NCEPR II, a nonlocal diffusion concept is used for the mixed layer (diffusion coefficients are still applied above the boundary layer). Briefly, in the mixed layer, the turbulent diffusion coefficients are calculated from a prescribed profile shape as a function of boundary layer height and scale parameters derived from similarity requirements (Troen and Mahrt 1986).

Finally, unlike the NCEPR, the NCEPR II does not force the soil moisture to an assumed climatology. When the first reanalysis began, it was soon discovered that the soil moisture was drying out too fast, especially over the Amazon. The soil moisture was subsequently corrected via a relatively fast (60 day) damping to an assumed climatology. Unfortunately, this correction resulted in too large a seasonal cycle, mainly because the assumed wintertime soil moisture was too high (Roads et al. 1999; Roads and Betts 2000; MOLRa,b). This correction also contributed to some of the excessive precipitation noted in the southeastern United States. This correction also resulted in small interannual variations because of the fast damping time and thus relatively small-amplitude coupled land–atmosphere interactions. Basically, NCEPR soil moisture constraints
were similar in spirit to the flux corrections previously developed for the coupled ocean–atmosphere models, and problems similar to those associated with the ocean flux correction occurred for the soil moisture correction.

NCEPRII corrects, instead, the model soil moisture by adding the previous pentad (5 day) difference between the reanalysis precipitation and observed precipitation to the soil moisture. Basically, the correction assumes that if the predicted runoff is zero, $N_o = 0$, then the difference between the observed and predicted precipitation, $P_o - P_p$, is added to top soil layer (10 cm). If $P_o < P_p - N_o$, then $P_o - (P_p - N_o)$ is added to top soil layer. If $P_o > P_p - N_o$, no correction is made because it is assumed that $N_o$ is the reason $P_p - N_o$ is less than $P_o$. There are some additional constraints: no correction is made if soil is frozen; lower soil layer is adjusted if the top layer saturates or dries; the correction is lagged in time by 5 days. When observed global runoff becomes available in near–real time, additional corrections could perhaps be made. Additional terms to constrain the skin temperature might also help. We believe NCEPRII’s attempt to use observed rather than model precipitation to drive the soil moisture helps provide more realistic surface temperature and soil moisture means and interannual variability. In fact, NCEPRII soil moisture corrections are more similar in spirit to a global land data assimilation system, which will eventually use even more observed variables to develop offline soil moisture and snow initialization models until the global analysis models are better able to simulate processes, such as precipitation.

The snow initialization was also modified for NCEPRII. If the model guess has snow but the observations indicate the presence of snow cover, the model guess is used. If the model guess does not have snow but the observations have snow cover, 10 mm of water equivalent snow is artificially added. If the model guess has snow but the observation is snow free, the model snow is removed without feeding it back to soil wetness. The last two processes externally influence the snow and hence the surface water budget.

It should be noted that although we concentrated here upon the NCEPRII analysis for almost all processes, we did not have the resources to fully compute moisture and energy fluxes from four-times-daily quantities that were not readily available from the NCEPRII anyway (this was a huge effort just to do this for NCEPR, since the entire three-dimensional structure of the atmosphere must be saved every 6 h). We thus used previously calculated NCEPR calculations for the heat and moisture convergence in combination with NCEPRII quantities for all other processes. A few limited cases demonstrated that the differences (at least for the integrated moisture and dry static energy convergence) were not large, presumably since the variables and processes important for atmospheric heat and moisture convergence are highly constrained by atmospheric observations of temperature, wind, and moisture and are not overly affected by the parameterized physics, which are not really all that different between NCEPR and NCEPRII, anyhow.

Although the NCEPRII output continues to be updated, we concentrate here upon a fixed 12-yr period (1988–99) in order to at least compare the precipitation to the Global Precipitation Climatology project (GPCP) precipitation data inferred from satellite and rain gauge measurements (see Huffman et al. 1997). It should be noted that this particular precipitation dataset has some missing data at high-latitude grid points and that these were filled-in with precipitation observations from the similar Xie and Arkin (1997) dataset to obtain global averages; in comparison to the NCEPRII, there is really no difference between these two precipitation datasets. A runoff climatology was also available from the Global Runoff Data Center (Fekete et al. 1999) for comparison to the runoff from the reanalysis. We used the composite dataset, which merged available observations with a water balance model driven by observed precipitation. Additional global datasets such as the National Aeronautics and Space Administration (NASA) Water Vapor Project (NVAP) dataset (Randel et al. 1996) are also slowly becoming available, but the purpose of this paper is not to carry out a full comparison of the NCEPR products with all available global and point observations; instead the purpose is to describe the basic characteristics and interrelationships of the water and energy budgets in NCEPRII, recognizing that it still needs further validation and improvement.

3. Water and energy cycle equations

Water and energy cycles are time-varying 3D quantities. Taking vertical averages in the atmosphere and subsurface, we focus here on 2D horizontal variations. By using time means (monthly), we focus on seasonal to interannual timescales. Consider first the atmospheric and surface water mass conservation equations:

\[
\frac{\partial Q}{\partial t} = E + MC - C + RSQ';
\]

\[
\frac{\partial W}{\partial t} = P - E + N + RSW'.
\]

The two state variables for these water mass conservation equations are $Q$, the vertically (pressure weighted) integrated specific humidity or precipitable water (see the appendix for details about the vertical integration), and $W$, the vertically integrated (2 m below the surface to the surface in the NCEP model) soil moisture ($M$) plus snow liquid water ($S$). The surface water, $W = M + S$, is computed only over land; over the ocean, it is more convenient to examine the salinity cycle (but this is not discussed here).
The water cycle described by these equations can be viewed rather simplistically in five steps. Under suitable conditions, liquid and solid water evaporate ($E$) from the ocean and land surface (which includes snow and vegetation) into the atmosphere. Water vapor is transported by atmospheric winds to other regions, and the convergence of this mass flux, $MC$ (see the appendix for details of how moisture convergence is calculated), will increase atmospheric water vapor over some regions while decreasing water vapor over other regions. Water vapor condenses into cloud particles, $C$. Cloud particles grow by condensation (or diffusion if solid), and then often by coalescence (aggregation if solid) and by accretion, into large liquid and solid drops, which fall as precipitation to the surface, $P$. If there is no horizontal cloud advection in a vertical column, there is as much water condensed as is precipitated and in the rest of the paper we assume the identity, $C = P$, where $C$ and $P$ are the vertical integrals. Although the contribution of cloud and precipitation evaporation to the total moisture budget is thought to be small, it can be important for influencing the dynamics, and most models now take into account at least the evaporation of rain through unsaturated layers. In any event, surface water is eventually increased by precipitation and decreased by evaporation. Continental rivers transport surface freshwater to other locations and the net divergence of this transport, $N$, will increase surface water in low-lying regions before discharging it into the oceans; in fact, most large-scale atmospheric models and the NCEP-PRII assume that the surface water is discharged immediately to the oceans. For a global long-term average, there is as much water precipitated as evaporated. For a long-term land average, there is as much atmospheric water converged over the land as is discharged by rivers to the oceans.

There are some additional artificial residual forcings, $RSQ'$ and $RSW'$, that appear in four-dimensional data assimilation (4DDA) analysis water budgets (e.g., Kanamitsu and Saha 1996). $RSQ'$ and $RSW'$ are denoted as residual in part because they are deduced as a balance of all other terms and also because they are not part of the natural processes and are only implicitly included in order to force the analyses' state variables close to observations. Unless some type of correction is done, current models will drift to their own model climate, instead of a climate close to the observations. In the atmospheric part of the analysis, this forcing is implicit by requiring the model to be adjusted to the available observations of atmospheric moisture, temperature, and winds. At the surface, some implicit adjustment must occur for snow, part (areal coverage) of which is an observed quantity. Additional adjustment occurs for the surface water, which uses observed precipitation instead of model precipitation to keep the soil moisture realistic. These adjustments are not random but instead are a systematic correction that results in substantial residual. Because of these residuals, one might think that re-

analyses cannot be used to study hydrometeorological budgets. However, it is worth pointing out that, since all models are designed to produce accurate budgets, and they will balance, independent of any residual, there are many other errors in models. In that regard, a continuous GCM, without any residual forcings and with perfect budgets, probably has larger errors in each of the individual terms than does an analysis with implicit residual forcings. An overall goal is to produce an analysis with small residuals and accurate estimates for each component of the budgets. In some sense these residuals then provide an estimate of the overall error in the budgets. As was discussed by Roads et al. (1998a), we had previously hoped that longer-term forecasts (say, 24 h) initialized from the reanalysis would decrease these residuals; unfortunately, even longer-term forecasts were required, which would eventually result in more unrealistic values for other components. Suffice it to say that there will not be a quick fix for these residual forcings, which are indicating fundamental errors in the model physical parameterizations.

In this paper, we combine the $RSQ'$ and $RSW'$ with the negative of the tendency terms ($RSQ = RSQ' - dQ/dt$ and $RSW = RSW' - dW/dt$). Please note that, while the natural atmospheric precipitable water tendency is small, the natural surface water tendency may be quite large and comparable to the other terms. Thus, we have to be careful when discussing $RSW$, with regard to whether this is a real phenomenon or part of the background analysis error.

Consider next the energy equations. The surface energy equation is simply the surface temperature equation:

$$C_{s} \frac{\partial \{Ts\}}{\partial t} = (QRS - LE - SH + G').$$

The atmospheric energy equation to a first approximation (mainly neglecting kinetic energy) is the atmospheric dry static or temperature equation and was described previously by Roads et al. (1997):

$$C_{a} \frac{\partial \{T\}}{\partial t} = (QR + LC + SH + HC + RST).$$

The surface energy input comes mainly from incoming solar and downwelling infrared radiation, moderated by reflected solar radiation and outgoing infrared radiation, $QRS$ (see the appendix for details). The net radiant energy that reaches the earth’s surface, $QRS$, is the source that controls temperature, drives evaporation, and is affected by atmospheric humidity and clouds.

Changes in the water phase can have a profound influence on the atmospheric and surface thermodynamic energy. Water cools its surroundings as liquid and solid water are converted into water vapor, $LE$. Globally averaged, this latent cooling must be balanced by the latent heat released when water vapor is converted to liquid and solid cloud particles, $LC$ (again, $LC = LP$ is as-
sumed here), which helps to balance the net radiative cooling of the atmosphere, QR. Because of the large latent heat involved in the condensation and evaporation of water molecules, water vapor is a very effective means of storing energy. The latent heat of fusion complicates the analysis. The latent heat required to melt snow should be balanced by the latent heat released when snow is formed initially. This exact relationship is usually not present in atmospheric models, which do not track the latent heat differences when snow is created from either the freezing of liquid drops or the conversion of vapor directly to ice. In particular, in the NCEPRII model, snow at the surface is assumed when the temperature above the surface reaches a certain minimum, but no latent heat is released when this happens. However, the fusion energy release relationship is tracked at the surface, where the latent heat of fusion is included in the melt process.

There are a number of other terms in the energy balance that are affected by the water phase change. The net equator-to-pole dry static energy transport or dry static energy convergence, HC, is positive and acts to balance the net atmospheric radiative cooling; HC also acts to balance the latent heat released, especially in the tropical regions. Cooling of the surface and heating of the atmosphere by turbulent transfers of sensible heat, SH, in the planetary boundary layer, is also governed by the latent heat release, since moist regions release more latent heat and thus require less sensible heat to achieve an energy balance.

Again, there are some additional terms (RST', G') that appear in analysis energy budgets that are not wholly related to natural processes. Heat storage in the land surface is thought to be small, but not negligible. NCE-PRRII includes a heat storage term, G', that releases heat to the atmosphere during the colder part of the year and stores heat during the warmer part of the year, and also includes the energy needed to melt snow. Again we combine these residual forcings with the negative of the tendency terms, although the tendency terms are thought to be small (RST = RST' – C_s dH/dt and G = G' – C_s dG/dt). Again the surface-heating tendency is the only tendency term thought to provide meaningful contributions on climate timescales. Although it is not shown here, we note that the surface heating tendency would have a strong influence on the sea surface temperatures (SSTs) if it were not for the surface currents, upwelling, and heat capacity that act to maintain their persistence.

4. Annual global land means

Table 1 provides the annual mean values (1988–99) for the various water and energy processes averaged (cosine weighted) globally and also over land and ocean separately. Averages for the individual CSEs are also shown. In the NCEPRII, precipitable water has an annual average of 24.7 mm for the globe, with larger values over the ocean (27.1) than over land (18.7), which is consistent with the higher ocean surface temperatures (17.2 versus 14.7). Also somewhat consistent with the surface temperature is that the largest precipitable water values occur over the tropical Amazon, CATCH, and GAME-Tropics regions, and the smallest values occur over Tibet, which is slightly warmer on average than the Lena River basin but at higher elevation. Surface water has an average value of 538 cm over land, which is strongly influenced by extreme amounts occurring in

<table>
<thead>
<tr>
<th>A</th>
<th>B</th>
<th>H</th>
<th>L</th>
<th>K</th>
<th>T</th>
<th>M</th>
<th>C</th>
<th>GLB</th>
<th>LND</th>
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<td>Q (mm)</td>
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<td>LE (K day$^{-1}$)</td>
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<td>0.44</td>
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<td>P/Q (day$^{-1}$)</td>
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<td>100P/W (day$^{-1}$)</td>
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<td>3.31</td>
<td>6.00</td>
<td>2.23</td>
<td>2.78</td>
<td>12.86</td>
<td>5.42</td>
<td>6.12</td>
<td>4.17</td>
<td>1.88</td>
</tr>
<tr>
<td>GPCP (mm day$^{-1}$)</td>
<td>5.13</td>
<td>2.09</td>
<td>2.54</td>
<td>1.15</td>
<td>1.08</td>
<td>4.62</td>
<td>2.15</td>
<td>2.79</td>
<td>1.44</td>
<td>2.58</td>
</tr>
<tr>
<td>GRDC (mm day$^{-1}$)</td>
<td>3.18</td>
<td>0.67</td>
<td>0.55</td>
<td>0.51</td>
<td>0.38</td>
<td>2.19</td>
<td>0.46</td>
<td>0.43</td>
<td>1.34</td>
<td>0.77</td>
</tr>
</tbody>
</table>
the coldest regions, where the permanent ice cover dominates (Greenland, Antarctica). In most of the CSEs, soil moisture is the dominant form of surface water, and surface water values range from 60 cm over the Amazon to 39 cm over the CATCH areas. Precipitation has an average global value of 3.1 mm day$^{-1}$, with larger rates over the ocean, 3.4 mm day$^{-1}$, and smaller amounts over land, 2.3 mm day$^{-1}$. This is a bit higher than the GPCP observations, by about 0.5 mm day$^{-1}$, especially over the oceans where the GPCP precipitation is 2.8 mm day$^{-1}$. NCEPRII evaporation has the same average global value as the NCEPRII precipitation but even larger values over the ocean, 3.6 mm day$^{-1}$, and smaller values over land, 1.8 mm day$^{-1}$; the difference is contributed by the moisture convergence (MC), 0.7 mm day$^{-1}$, which is largest and smallest in the tropical monsoon regions, as well as the residual atmospheric moisture forcing (RSQ), $-0.2$ mm day$^{-1}$. The runoff (global average of 1 mm day$^{-1}$) is largest in the tropical regions. This runoff is contributed by the $(P - E)$ forcing of 0.5 mm day$^{-1}$ and a residual surface forcing (RSW) of 0.5 mm day$^{-1}$, which over balances the residual atmospheric forcing (RSQ $= -0.2$ mm day$^{-1}$), indicating that the NCEPRII model would approach a drier climatology if it wasn’t constantly being nudged by observed precipitation.

Better relative balances are seen in the atmospheric and surface energy balances. In the atmosphere, the average cooling rate of $-0.95$ K day$^{-1}$, which is slightly greater over the ocean than over land, is balanced by the latent heat released by precipitation (LP $= 0.78$ K day$^{-1}$), the heat converged into an area (note HC $\sim 0$ for the globe), although in tropical areas heat (dry static energy) is diverged in order to also balance the latent heat released by precipitation, and the sensible heating (SH $= 0.11$ for the land average). A residual forcing of 0.1 K day$^{-1}$ is also needed globally, with somewhat larger values in each of the CSEs. At the surface, the latent cooling is usually more important than SH (LE $= 0.45$ K day$^{-1}$) for the land average and is the major component balancing the net surface radiative heating (QRS $= 0.61$ K day$^{-1}$ for the land average). For the most part, G is small but negative, with the largest negative values occurring in high-latitude BALTEx, Lena River, and Tibet (B, L, E) regions with substantial snowmelt (again, the NCEPRII model includes the energy to melt snow in the surface energy balance).

How accurately are the budgets closed? When the original idea of budget closure was developed, it was related to the then inaccurate estimates of moisture convergence and how this moisture convergence was related to observed streamflow, since for a long-term average both should be equal (see, e.g., Roads et al. 1994). In that regard, we can see here that for a global land average, the NCEPRII moisture convergence is 0.69 and the “observed runoff” is 0.77, indicating the global water budget has been balanced to within 10%. Except over the Mississippi, BALTEx, and GAME-Tropics, the budget closes to much less accuracy, which indicates that the global land hydrologic budget is closed by compensating differences in different regions. Simply finding out the differences between analysis moisture convergence and observed runoff does not really answer the budget closure question. Other terms need to be considered. For example, the NCEPRII model has a global land runoff of 0.94, which has less of a balance with the moisture convergence. The global land NCEPRII precipitation is 1.8 mm day$^{-1}$, whereas the observed is closer to 2.0 mm day$^{-1}$. Even larger differences occur for certain seasons, especially summer. Although models are designed to accurately close budgets (otherwise they would become unstable) analyses are closed in part by nudging the model state variables close to observations, and the magnitude of this nudging thus becomes one additional estimate of how accurately the budget is closed. In that regard, it should be noted that the global land averaged atmospheric residual is $-0.21$, whereas the surface water residual is 0.46; together these residuals compensate each other and the total residual is equal to the difference of N and MC.

To illustrate the hydroclimatic characteristics of the CSEs, Fig. 2 plots the monthly and annual mean climatologies of various components of the water cycle with respect to surface temperature. That is, a 12-yr (1988–99) reanalysis monthly mean climatology (12 months) was developed from the 144 monthly means, and the annual means along with the monthly means were used to describe the hydroclimatic characteristics associated with certain mean temperatures. As will be shown, surface temperature does not necessarily define a unique hydroclimate, but it does provide an interesting and orderly counterpart to describing the hydroclimates regionally. Moreover, this order provides a possible qualitative indication as to what might happen in the future as the climate warms.

The coldest regions include the Lena and Mackenzie River basins as well as Tibet. Midlatitude regions include the Mississippi, BALTEx, and HUBEx regions. Tropical regions include the Amazon, GAME-Tropics, and CATCH regions. We also show the average monthly values for the land regions associated with each surface temperature. For each climatological month (January–December) each land grid point was categorized according to a discrete temperature value from $-30^\circ$ to $30^\circ$C, and then the various process values were spatially averaged (cosine of latitude weighted) over all land regions for each discrete temperature value. It should be noted here that monthly variations for the CSEs, as well as the values for individual grid points, tend to follow the averages denoted by the solid lines, and that individual CSEs can actually cover a large range of values during seasonal excursions. For simplicity, only the annual means for each CSE are shown.

Precipitable water (Fig. 2a) increases exponentially with increasing surface temperature, and the various CSE basins line up with the cold basins, Lena, Mac-
FIG. 2. Monthly mean (1988–99) water cycle processes as a function of surface $T$, for the globe. The cosine-weighted average of all the land grid points is shown by the solid line. Annual means are shown by the large capital letters designating the Mackenzie River basin (K), Baltic Sea basin (B), Mississippi River basin (M), Amazon River basin (A), CATCH (C), and the four GAME locations [Lena River basin (L), Tibet (E), HUBEX (H), and Tropics (T)]. (a) $Q$ (mm); (b) $W$ + $M$ (mm); (c) $P$ (mm day$^{-1}$); (d) $E$ (mm day$^{-1}$); (e) MC (mm day$^{-1}$); (f) $N$ (mm day$^{-1}$); (g) $P/Q$ (days$^{-1}$); (h) $P/W$, (days$^{-1}$). The dotted curve in (a) is the theoretical Clausius–Clapeyron curve. The dotted curves in other panels represent observations.

Mackenzie, and Tibet having the least amounts, and the tropical basins, GAME-tropics, Amazon, and CATCH having the greatest amounts. This relationship reflects the well-known Clausius–Clapeyron relationship (dotted line), which shows that the water-holding capacity of the atmosphere increases exponentially with surface temperature. Note that the precipitable water becomes very small in comparison to the actual holding capacity of the atmosphere, mainly because the warm regions of western Africa (CATCH) are relatively dry. In fact, the
decrease above 25°C for this and other variables below reflects the change to desert locations. Also, this theoretical precipitable water curve does not take into account the different vertical temperature structure associated with, for example, cold surface temperatures (which tend to have a more stable atmosphere) as well as the steeper lapse rates in desert regions. Instead, the theoretical curve assumes a constant lapse rate of 6.5 K km⁻¹ and thus provides only a first-order impression as to how the precipitable water should vary with respect to surface temperature. It should also be noted that the Clausius–Clapeyron curve shown here does not account for the saturation vapor pressure over ice. This is actually a complex situation for the atmosphere, which can often maintain supersaturation over water rather than ice because of the great difficulty in forming ice crystals directly from vapor. Despite all the minor problems, to first order the Clausius–Clapeyron relationship provides an upper bound on the atmospheric precipitable water.

Surface water (Fig. 2b) behaves differently. The colder regions have the greatest amounts of surface water, mainly because of snow accumulation and the water stored in the permanent snow-covered areas in Antarctica and Greenland. However, the surface water is also relatively larger in the wet areas of the Amazon and GAME-Tropics. This quasi-bimodal surface behavior shows up much more strongly in the surface runoff. Precipitation (Fig. 2c) also increases with increasing surface temperature, mainly because evaporation (Fig. 2d) increases with increasing surface temperature. However, precipitation is also dependent upon moisture convergence (Fig. 2e), which is bimodal in that it increases with increasing surface temperature for low and high temperatures but has a minimum between 15° and 20°C. (Slightly negative values are indistinguishable from the zero line and are balanced by the moisture residuals.) Because of the moisture convergence associated with wintertime extratropical cyclones in high latitudes, moisture convergence is strongest near freezing temperatures, although not as strong as the surface runoff (Fig. 2f), which is strongly peaked near the surface temperatures associated with snow melt. At the higher surface temperatures associated with subtropical zones, the moisture convergence decreases and then increases once again in the tropical monsoon regions, as does the runoff, before eventually decreasing at the highest surface temperatures, associated with tropical deserts. It should be pointed out here that, although the exact quantitative relationships deduced from precipitation and runoff are not the same as those deduced from the observations, there are no fundamental qualitative differences.

We also show a couple of measures of hydrologic cycle acceleration, one of which was originally described by Chahine et al. (1997; see also Roads et al. 1998b). In particular, an atmospheric rate (or time constant, which is the inverse of the rate) for the atmospheric hydrologic cycle can be defined from the ratio of the precipitation to the precipitable water. The decrease in this rate (Fig. 2g) indicates that, basically, the rate decreases with increasing surface temperature, mainly because the atmosphere has a much greater water-holding capacity at higher temperatures than an intrinsic ability to produce greater amounts of precipitation. However, there are signs of an atmospheric acceleration for some of the higher surface temperatures associated with warmer moist climates that may be associated with monsoon regions. A surface water acceleration rate can be defined in a similar fashion, except surface water is used instead of atmospheric precipitable water. In this case (Fig. 2h), there is an overall tendency for an acceleration to occur, especially in regions close to zero and 25°C. Suffice it to say that these acceleration measures are just one means of characterizing the hydrologic timescales associated with increased surface temperatures. For example, precipitation intensity and amount, in addition to other second-moment terms, must increase with increasing temperature.

Figure 3 shows the corresponding energy cycle processes as a function of surface temperature. Note first of all that the radiative cooling of the atmosphere is relatively constant (Fig. 3a) with respect to surface temperature. This radiative cooling is balanced mainly by the dry static energy convergence (Fig. 3e), which acts to counteract the radiative cooling at high latitudes. At lower latitudes, the dry static energy convergence decreases, as the latent heat released by precipitation (Fig. 3c) increases, and for the highest temperatures the net dry static energy convergence becomes negative since it must balance the latent heat release. Turbulent sensible heat transport (Fig. 3g) modifies the large-scale heat transport by transporting heat from the atmosphere to the surface in high latitudes and transporting heat from the surface to the atmosphere in lower latitudes.

The surface energy balance is strongly forced by the net radiative heating at the surface (Fig. 3b). Since the solar heating of the surface is greater than longwave cooling, the turbulent transport of sensible (Fig. 3g) and latent (Fig. 3d) heat from the surface to the atmosphere is also needed to cool the surface. The ground heating is close to zero, except near freezing, where the model takes into account the energy needed to melt snow.

5. Geographic variations

Figure 4 shows the geographic characteristics of the processes involved in the annual mean water cycle. Precipitable water (Fig. 4a), which is strongly temperature dependent, is largest in the Tropics and then decreases to near zero at the poles. There is a tendency to have larger amounts in ascending regions of the Walker circulation (tropical land masses) and lesser amounts in the subtropical descending regions (east equatorial Africa and ocean regions, which are not shown). Surface water (Fig. 4e) has similar variations in the Tropics and subtropics, but by contrast is much greater in mid to
ENERGY CYCLE

Fig. 3. Monthly mean (1988–99) energy cycle processes as a function of surface $T$, for the globe. The cosine-weighted average of all the land grid points is shown by the solid line. Annual means are shown by the large capital letters designating the Mackenzie River basin (K), Baltic Sea basin (B), Mississippi River basin (M), Amazon River basin (A), CATCH (C), and the four GAME locations [Lena River basin (L), Tibet (E), HUBEX (H), and Tropics (T). (a) QR (K day$^{-1}$); (b) QRS (K day$^{-1}$); (c) LC (K day$^{-1}$); (d) LE (K day$^{-1}$); (e) HC (K day$^{-1}$); (f) $G$ (K day$^{-1}$); (g) SH (K day$^{-1}$).

high latitudes, especially in regions where snow provides a major contribution to the surface water. Precipitation (Fig. 4b) is related to precipitable water but tends to show stronger variations, especially in middle latitudes. For the most part, land precipitation corresponds to evaporation (Fig. 4f), except on the westernmost land regions of midlatitude continents, where moisture convergence (Fig. 4c) plays an important role, as it does in the tropical regions. For the most part, regions of strong moisture convergence are related to regions of...
strong surface runoff (positive over most land regions), although the western slopes of Tibet (Himalayas) have strong surface runoff that does not appear in the moisture convergence. This anomalous surface runoff is balanced to a certain extent by the residual term, which is provided in part by correcting the model soil moisture affected by model precipitation with observed precipitation (Kanamitsu et al. 2002). The residual precipitable water tendency is also large and, as will be shown later, is related in part to the erroneous model precipitation. Finally, the precipitable water and surface water timescales are shown in Figs. 4d and 4h. As discussed previously, the precipitable water timescales are fairly fast in high-latitude regions, whereas the surface water timescales are fairly fast in low latitudes. This indicates a relative slowing of the atmospheric hydrologic cycle with respect to an increase in temperature but a relative acceleration of the surface hydrologic cycle. It should be noted that these regional results are only qualitatively correct and need to be further evaluated with available observations.

Figure 5 shows the processes involved in the annual mean energy cycle. In the atmosphere, longwave radiation dominates over the solar heating, and the net radiation (Fig. 5a) acts to cool the atmosphere almost uniformly over land, with the smallest cooling occurring

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**Fig. 4.** Annual mean atmospheric water cycle. (a) $Q$ (mm); (b) $P$ (mm day$^{-1}$); (c) MC (mm day$^{-1}$); moisture flux vectors, max = 2 kg; (d) 10P/Q (day$^{-1}$); (e) $W$ (mm); (f) $E$ (mm day$^{-1}$); (g) $N$ (mm day$^{-1}$); (h) 100P/W (day$^{-1}$).
in dry subtropical regions since only sensible heating (Fig. 5g) balances the radiative cooling there. In other regions, the cooling is balanced in part by latent heat released by precipitation (Fig. 5b), sensible heat, and also the dry static energy convergence (Fig. 5c). The dry static energy convergence, which includes the adiabatic cooling, acts to heat the atmosphere in high latitudes and balances the radiative cooling as well as the sensible heat transfer to the surface in high latitudes. The dry static energy convergence also acts to cool the atmosphere, especially in the tropical regions of intense precipitation. At the surface, solar radiation dominates the infrared cooling, and the net surface radiation (Fig. 5e) acts to heat the surface, which is cooled by sensible heating in low latitudes and latent cooling (Fig. 5f) associated with surface evaporation. The ground heat flux (Fig. 5h) also acts to cool the surface, especially in regions associated with snowmelt (the relative larger amounts along the edge of the continents is a masking problem). The residual atmospheric temperature forcing (Fig. 5d), caused by the tendency for the analysis model to drift toward its own climatology, is not small and is discussed further for the next figure.

Figure 6 shows how the annual budget residuals compare to the difference between the model precipitation and the observed (Xie and Arkin 1997) precipitation (Figs. 6b,e). Of particular interest here is the noticeable relationship of the RSQ and RST residuals (Figs. 6a,d).
to the precipitation differences of the reanalysis from observations, indicating that improving the precipitation physics will reduce the atmospheric moisture and temperature residuals. However, precipitation is obviously not the only problem, and contributions from other terms are also indicated. For example, the surface residuals (Figs. 6c,f) tend to be related to snowmelt in southern Alaska and British Columbia and western Tibet, indicating that a proper accounting of the snow budget is still needed in the analysis model for at least the surface water budget. Runoff differences (Fig. 6g) also indicate that improving this feature might help lead to a reduction in the residual surface forcing.

Figure 7 shows the June–August (JJA) variations (with the annual mean removed) in the water cycle. Of special interest here is the overall increase in precipitable water (Fig. 7a), precipitation (Fig. 7b), and evaporation (Fig. 7f) over most of the Northern Hemisphere continental regions and the decrease over the Southern Hemisphere continental regions. The major exception occurs over the regions with a Mediterranean climate, which, in addition to the Mediterranean region, include California and Chile. Surface water variations are more complicated. There is a decrease of surface water (Fig. 7e) in northern midlatitudes, where evaporation is greater than precipitation, consistent with the strong moisture
Moisture divergence (Fig. 7c) is clearly one of the distinguishing characteristics between low-latitude monsoon climates and midlatitude climates. During the summer, regions associated with summer monsoons have a strong increase in moisture convergence, whereas midlatitude regions have a strong decrease in moisture convergence, which may be related to the inverse relationship found earlier by Higgins et al. (1997) between Mississippi River basin precipitation and Mexican monsoon precipitation. It should be noted here that some midlatitude regions do show an increase in moisture convergence during the summer, although these do not fall within the various CSEs. Note, for example, the
increase in moisture convergence on the northeast coast of Asia as well as the northeast coast of the Americas and even the west coast of Greenland. By contrast, runoff (Fig. 7g) is almost always greater during the summer (except for in the Mediterranean climates), reflecting perhaps snowmelt tendencies as well as greater precipitation tendencies.

Another distinguishing characteristic between tropical monsoons and midlatitude regions is the tendency of soil moisture to decrease (Figs. 7e,h) during the summer in midlatitude regions but increase in tropical monsoon regions. The balance between this moistening and drying can also become quite complicated in climate models that attempt to model anthropogenic climate changes; some climate models show an incursion of monsoon variations into higher latitudes, whereas others show an amplification of current midlatitude and tropical climate regimes (as manifested by different soil moisture tendencies in different general circulation models).

Figure 8 shows the JJA seasonal variations in the energy cycle, which resemble in many ways the moisture budget, due to the strong influence of the latent heat released when condensation and precipitation (Fig. 8b) occur. This increased latent heating is augmented by the slight decrease in radiative cooling (Fig. 8a) and increase in sensible heating (Fig. 8g) that occurs in a summertime atmosphere, all of which are balanced by...
the increased dry static energy divergence (Fig. 8c) and to a lesser extent by the residual temperature forcing from the analysis (Fig. 8d). At the surface, the increased JJA solar forcing (Fig. 8e) is diminished by the increased sensible and latent (Fig. 8f) cooling. Snowmelt also contributes to the surface energy balance, which is combined here with the ground heating term (Fig. 8h).

6. Seasonal basin variations

To further examine the seasonal variations we next show seasonal plots of basin averages for the various CSEs. Figure 9 shows the seasonal atmospheric water cycle. Note that evaporation, which moistens the atmosphere, is positive year round and reaches its maximum during the summer. Precipitation also reaches its maximum value during the summer, but since it dries the atmosphere it is shown for this budget as a negative number. (Note that in comparison to observations, NCEP-PRR precipitation provides a useful simulation, albeit with too large summertime precipitation.) The major difference between the CSEs is that low-latitude monsoon regions (Amazon (remember summer in the Southern Hemisphere is DJF) CATCH, GAME-Tropics, GAME-HUBEX, and even Tibet) tend to have increased moisture convergence during the summer, whereas the other midlatitude regions (Mississippi, Mackenzie, Baltic, Lena) tend to have decreased moisture convergence, and even moisture divergence, during the summertime. For the low-latitude regions associated with the monsoons, there is a clear relationship with increased precipitation and moisture convergence as well as evaporation during the summertime.

Moisture divergence implies that evaporation is greater than precipitation during the summertime, which it would be except that there is a positive residual in some places (e.g., the Mississippi River basin) equivalent to the moisture divergence. Since this residual can in some sense be taken as the negative of the analysis tendency, we see that this seasonal pattern represents spindown (decrease in atmospheric water) during the summer and spinup during the winter. These seasonal spindown/spinup patterns are a known problem with current convective schemes (see Roads et al. 1997). At least the residual has the same sign as the moisture convergence in lower latitudes, which, again, exhibits stronger moisture convergence during the summer. Somewhat disconcerting is that the magnitude of the residual forcing can be as

**Fig. 9.** Seasonal atmospheric water cycle for the continental-scale regions (mm day$^{-1}$): −$P$ (thin dashed line); −GPCP (thick dashed line); MC (thick dotted line); E (thin dotted line); RSQ (thin solid line).
large as the moisture convergence and, moreover, has the opposite sign of moisture convergence in middle latitudes. Despite its possible influence upon the moisture divergence, we believe the residual forcing is more related to the analysis precipitation problems and only coincidentally resembles the moisture divergence seasonal cycle, which is also constrained by wind and moisture observations.

Figure 10 shows component of the surface water cycle. As shown previously in Table 1, the surface runoff and moisture convergence approximately balance. However, the residual forcing can still be quite large and comparable to the runoff. Note, for example, the Mackenzie, which has a strong positive RSW, which balances the runoff. In fact, all the midlatitude regions have this strong positive residual, peaked near the springtime, indicating that the surface moisture tendency should have a strong drying trend during this time, or that there is a large difference between the observed and forecast precipitation in these regions. Tropical monsoon regions have the opposite tendency, in that they have a negative RSW during the summertime and a positive RSW during the wintertime.

Figure 11 shows the corresponding atmospheric energy cycle. Atmospheric radiative cooling is relatively constant but has a slight decrease during the summer. Since the latent heat released by precipitation also increases during the summer, a balance is reached through the heat divergence, which is weak during the winter but reaches a maximum negative value during the summer in all regions. The residual RST is positive during the winter and becomes negative during the summer, which is similar to the impact upon the moisture equation by the precipitation. Again, correcting the errors in the precipitation field should reduce the residuals in the moisture and energy budgets.

Figure 12 shows the surface energy cycle. Here the dominant balance is the heating by the net surface radiation, which is dominated by downwelling solar radiation. For most regions, except Tibet and CATCH, the latent heat of evaporation is greater than the sensible heat, which can cool the surface in low latitudes during the summer and heat the surface in high latitudes during the winter. For the most part, the CSEs have a good surface energy balance. However, as mentioned previously, the snowmelt contribution to the surface energy balance is not negligible, especially in high latitudes and during the spring, G makes a noticeable contribution to the surface energy balance.
7. Summary

To provide a global background for the CSE regional WEBS studies, we described here vertically integrated global and regional water and energy budgets from the NCEP/NCAR. To first order we can discern important mechanisms in the reanalysis. For example, water and energy cycles can be characterized by surface temperature. Water vapor, precipitation, evaporation, surface and atmosphere radiative heating (less atmosphere cooling), and atmospheric heat divergence increase with increasing surface temperature. Moisture convergence and other surface features are bimodal because of different seasonal relationships in middle-latitude and monsoon regions. Surface water decreases with surface temperature in middle latitudes but increases in tropical monsoon regions. This surface water variation has a strong impact upon the sensible heating, which increases over the drying midlatitude land surface and decreases over the moist monsoon land regions. Surface runoff is also bimodal with surface temperature, in part because much runoff occurs near melting temperatures as well as during the rainy monsoon seasons. Moisture convergence also has two preferred modes. During winter, the atmosphere is more efficient in transporting moisture to high-latitude regions. During summer, increase in moisture convergence is a large contributor to the monsoon rainfall, whereas summertime moisture divergence is common over the midlatitude continents.

Future work must not only begin to examine regional characteristics in greater detail, and how they might have changed under different conditions, but must also examine regional characteristics that may be influenced by remote influences. For example, there are numerous low-frequency features (not shown) in the surface water that vary from region to region that cannot be understood through the midlatitude and monsoon classification used in this paper. For example, the causes of 1) the recent drying trend for the Amazon, Mackenzie, BALTEx, GAME-Tibet, and Lena River basins; 2) the recent moistening trends of the Mississippi after the 1988 drought and then the subsequent drying during the late 1990s; and 3) the strong interannual variations in the GAME-HUBEX, GAME-Tropics, and CATCH regions, as well as the variations in other regions, are subjects for future research. In that regard, it should be mentioned that the large influence of the residual terms,
especially for interannual variations (see Roads et al. 1997), makes these kinds of studies particularly difficult with the reanalysis.

Finally, it is worthwhile to emphasize that many global datasets are not yet available to fully validate many of the water and energy processes described here in the NCEPRII. Additional modeling and observational studies will be required to fully understand regional and global water and energy cycle interactions and variations on seasonal to interannual timescales.

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APPENDIX

Budget Computations

a. Normalization of water and energy equations

In the water budget equations, all vertically integrated reanalysis water budget terms [kg (m² s⁻¹)] are multiplied by 8.64e⁻³ 10⁴ s day⁻¹, which provides individual values in kg (m² day⁻¹) or mm day⁻¹ (using a density of 1000 kg m⁻³ for water). In the atmospheric temperature or energy budget equation, all terms (W m⁻²) are multiplied by 8.64⁻³ 10⁴ (C_p g⁻¹), to provide normalized values in units of K day⁻¹. Here C_p = 1.0046 × 10³. It should be noted here that the p used for this normalization was taken from the monthly mean pressure files only and was not included in the instantaneous calculations, which, however, did include the instantaneous pressure in the flux calculations (see below). The normalization is simply for presentation purposes. The surface energy terms are also multiplied by a constant atmospheric normalization in order to provide values in K day⁻¹. That is, C_eH = C_e(p/g) = 10⁷ W kg⁻¹ K⁻¹.
kg m$^{-2}$. This normalization definition is only a simplification, in order to easily compare surface energy variations with atmospheric energy variations. The heat capacity of the surface is certainly different from the atmosphere, and adequate accounting of this surface heat capacity is needed in order to properly represent diurnal and seasonal cycles. However, as is the case for many global climate models, the heat capacity of the surface was ignored since only an average heat balance is being analyzed here.

b. Vertical integrals

Pressure-weighted vertical integrals can be written in sigma coordinates as

$$\int_{\sigma_0}^{\sigma_1} p \, d\sigma/g.$$  

For example, $Q = \{q\} = p, \int_{\sigma_0}^{\sigma_1} qg \, d\sigma.$

An equivalent mass weighting also occurs for the surface (and below) terms. However, since the liquid (or solid) water density is constant, this is a vertical integration in height.

c. Moisture and energy convergence

The vertically integrated moisture convergence is

$$MC = -\nabla \cdot (\mathbf{v} \mathbf{q}).$$

The vertically integrated energy convergence or vertically integrated temperature convergence and adiabatic expansion is

$$HC = -\nabla \cdot (\mathbf{v}(CpT + \Phi)).$$

The above expression for dry static energy convergence is used here, instead of the separate temperature convergence and adiabatic expansion term, in order to avoid explicit calculation of the vertical pressure velocity. However, preliminary tests showed that both calculations are equivalent, which also indicated that the contribution of kinetic energy to this equation is negligible (see Roads et al. 1997). It should also be noted that the NCEP/1II uses virtual temperature as its state variable and thus the temperature needed for the energy equation and dry static energy convergence calculation must first be derived from $T_s = (1 + 0.61q)T$, where $T$ is the actual temperature, after first deriving the geopotential from the virtual temperature. For simplicity, we decided not to try to emulate exactly the virtual temperature equation in the analysis model and to instead concentrate on developing a useful energy equation, which is approximately conserved by the analysis.

These convergence terms were explicitly calculated using four-times-daily NCEPR data from advection forms of the equations. The advection form was used in order to emulate the actual analysis model calculation of these terms, which uses the advection form to eliminate numerical noise when three spectral quantities $(p, u, q)$ are multiplied together and then a derivative is taken. In advection form, the definitions become

$$\frac{g}{p_s}MC = -\left[ \mathbf{v} \cdot \nabla q + \frac{\partial q}{\partial \sigma} + \frac{q}{p_s} \left( \frac{\partial p_s}{\partial t} \right) \right],$$

$$\frac{g}{p_s}HC = -\left[ \mathbf{v} \cdot \nabla (CpT + \Phi) + \frac{\partial (CpT + \Phi)}{\partial \sigma} \right]$$

$$+ \frac{(CpT + \Phi)}{p_s} \left( -\frac{\partial p_s}{\partial t} \right).$$

Standard recursion relationships were used to compute the horizontal gradients. Note that in the advective form, the surface pressure is used to mainly provide normalization for the equations after the fact. This form of the equations thus requires that only quadratic quantities be calculated and therefore is less noisy than corresponding flux convergence terms involving the nonlinear product of three quantities.

The sigma velocity was derived from the pressure equation

$$\frac{\partial p_s}{\partial t} = -p_s \mathbf{v} \cdot \mathbf{\hat{v}} - p_s \mathbf{\hat{v}} \cdot \mathbf{\nabla} p_s - p_s \frac{\partial \sigma}{\partial \sigma},$$

where $\sigma(1) = \sigma(0) = 0$.

The winds needed for these computations were first calculated from the tabulated divergence and vorticity,

$$\mathbf{\nabla} \cdot \mathbf{\hat{v}} = \mathbf{\nabla} \cdot \mathbf{\hat{\omega}} = \mathbf{\hat{\omega}} \times \mathbf{\hat{v}} = \mathbf{\nabla} \cdot \mathbf{\hat{v}},$$

using standard spherical harmonic recursion relationships (see Roads et al. 1992). The mean velocity was then corrected (see Trenberth and Guillemot 1994 and Roa et al. 1992) so that the wind satisfied the integrated surface pressure equation,

$$\frac{\partial p_s}{\partial t} = -g \mathbf{\nabla} \cdot \{ \mathbf{\hat{v}} \},$$

for every four-times-daily observation $(dp/dt$ was calculated as a centered finite difference). As pointed out by Trenberth and Guillemot (1994), this relationship is really only true for the dry air mass, since the total mass can be changed by the weight of the water, which has sources and sinks; however, since the correction is small in any event, and since dry air mass is not conserved by the analysis, which instead explicitly conserves the total mass, we decided to use just the total surface pressure and not make any corrections for dry air mass.

Finally, it should be noted that it would certainly be better to calculate the convergence terms from continuous model accumulations since the four-times-daily computations can lead to sampling noise (see, e.g., E. Chen et al. 1996). In fact, Roads et al. (1998a) did just that by initializing the NCEPR version of the global...
model every day and carrying out short-term forecasts. They found that any differences between the more exact computation and the approximations described above appeared small for the vertically averaged budgets for the Mississippi River basin. Thus, despite all the potential problems with using four-times-daily NCEPR data to estimate moisture and dry static energy convergence, we still believe to first order that we have developed a useful approximation for the accumulated fluxes and that to first order we can thus deduce the true NCEPRII residuals.

d. Radiative heating and fluxes

The net surface solar radiation can be written in terms of the upward and downward components

\[ \text{NSW}(1) = \text{SWU}(1) - \text{SWD}(1), \]

where the terms in parentheses indicate the values at the surface \((\sigma = 1)\), SWU indicates the upward (reflected) solar flux, and SWD represents the downwelling solar radiation. The net surface infrared radiation is written similarly in terms of the upward and downward components:

\[ \text{NLW}(1) = \text{LWU}(1) - \text{LWD}(1). \]

The net solar radiation at the top of the atmosphere is written similarly in terms of upward and downward components:

\[ \text{NSW}(0) = \text{SWU}(0) - \text{SWD}(0), \]

where the terms in parentheses indicate the values at the top of the atmosphere \((\sigma = 0)\). Since there is negligible downward infrared radiation from space, the net infrared radiation at the top of the atmosphere is simply the upward component

\[ \text{NLW}(0) = \text{LWU}(0). \]

The surface radiative heating can thus be written in terms of the net solar and infrared fluxes,

\[ \text{QRS} = -[\text{NSW}(1) + \text{NLW}(1)], \]

which provides on the average about 100 W m\(^{-2}\) or 1 K day\(^{-1}\) but which varies greatly between equator and pole \((\sim 0)\). In the atmosphere, the net radiative heating is given by

\[ \text{QR} = [\text{NSW}(1) + \text{NLW}(1) - \text{NSW}(0) - \text{NLW}(0)], \]

which is a more uniformly negative number of about 100 W m\(^{-2}\) or \(-1\) K day\(^{-1}\).

REFERENCES


