

The flow of energy through the earth's climate system

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SUMMARY

The primary driver of the climate system is the uneven distribution of incoming and outgoing radiation on earth. The incoming radiant energy is transformed into various forms (internal heat, potential energy, latent energy, and kinetic energy), moved around in various ways primarily by the atmosphere and oceans, stored and sequestered in the ocean, land, and ice components of the climate system, and ultimately radiated back to space as infrared radiation. The requirement for an equilibrium climate mandates a balance between the incoming and outgoing radiation, and further mandates that the flows of energy are systematic. These drive the weather systems in the atmosphere, currents in the ocean, and fundamentally determine the climate. Values are provided for the seasonal uptake and release of heat by the oceans that substantially moderate the climate in maritime regions. In the atmosphere, the poleward transports are brought about mainly by large-scale overturning, including the Hadley circulation in low latitudes, and baroclinic storms in the extratropics, but the seamless nature of the transports on about monthly time-scales indicates a fundamental link between the two rather different mechanisms. The flows of energy can be perturbed, causing climate change. This article provides an overview of the flows of energy, its transformations, transports, uptake, storage and release, and the processes involved. The focus is on the region 60°N to 60°S, and results are presented for the solstitial seasons and their differences to highlight the annual cycle. Challenges in better determining the surface heat balance and its changes with time are discussed.

KEYWORDS: Atmospheric energy Hadley circulation Heat budget Ocean energy

1. INTRODUCTION

The general picture of the earth's energy balance has been reasonably well known for some decades, but quantification of the flows, storage and transports of energy, and conversions into different forms of energy, has been continually evolving. Satellite data, especially from the Earth Radiation Budget Experiment in the late 1980s, provided major constraints on incoming absorbed solar radiation (ASR) and outgoing long-wave radiation (OLR), as well as the role of clouds from the perspective from space at the top-of-atmosphere (TOA). Estimates of atmospheric energetics, summarized by Lorenz (1967), have improved substantially using globally gridded re-analyses (Trenberth *et al.* 2001; Trenberth and Stepaniak 2003a,b). Trenberth and Caron (2001) showed that the latest estimates of atmospheric poleward energy transports are larger than earlier ones by as much as a petawatt (1 petawatt (PW) is 10^{15} watts), which is of the order of 20% of the total, owing to much improved representation of storm tracks and atmospheric waves over the oceans. When combined with TOA radiation observations, results are

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now quite compatible within error bars of order 0.3 PW, with direct ocean observations along various sections from the World Ocean Circulation Experiment amongst others. Accordingly, a somewhat revised view of the processes involved in all aspects of the energy cycle is emerging, although considerable uncertainties remain concerning the surface and how the energy cycle may respond during climate change. This article provides an up-to-date overview of the earth's energy budget and how it is achieved, weighted toward an atmospheric perspective. Results are presented here for the solstitial seasons from 60°N to 60°S to emphasize the seasonal changes in the tropics and subtropics associated with the Hadley, Walker and monsoonal circulations. Climate change aspects are dealt with in Karl and Trenberth (2003).

The atmosphere does not have very much heat capacity but is very important as the most volatile component of the climate system in moving heat and energy around, with wind speeds in the jet stream often exceeding 50 m s^{-1} . The oceans have enormous heat capacity and, being fluid, also can move heat and energy around in important ways. Ocean currents may be $>1 \text{ m s}^{-1}$ in strong currents like the Gulf Stream, but are more typically a few cm s^{-1} at the surface. Other major components of the climate system include sea ice, the land and its features (including the vegetation, albedo, biomass, and ecosystems), snow cover, land ice (including the semi-permanent ice sheets of Antarctica, Greenland and glaciers), rivers, lakes, and surface and subsurface water. Their role in storage, transport and balance of energy on the earth is addressed below.

Changes in any of the climate system components, whether internal and thus a part of the system, or from the external forcings, cause the climate to vary or change. Thus climate can vary because of alterations in the internal exchanges of energy or in the internal dynamics of the climate system. Examples are El Niño–Southern Oscillation events, which arise from natural coupled interactions between the atmosphere and the ocean centred in the tropical Pacific. There is often a mini global warming following an El Niño, as a consequence of heat from the ocean affecting the atmospheric circulation and so changing temperatures around the world (e.g. Trenberth *et al.* 2002). Consequently, interannual variations occur in the energy balance of the combined atmosphere–ocean system and are manifested as important changes in weather regimes and climate around the world.

2. THE GLOBAL ENERGY BUDGET

The incoming energy to the earth system is in the form of solar radiation, and the average amount of energy incident on a level surface outside the atmosphere is one quarter of the total solar irradiance, or $\sim 342 \text{ W m}^{-2}$. About 31% of this energy is scattered or reflected back to space by molecules, tiny airborne particles (known as aerosols) and clouds in the atmosphere, or by the earth's surface, which leaves $\sim 235 \text{ W m}^{-2}$ on average to warm the earth's surface and atmosphere (Fig. 1; Kiehl and Trenberth 1997). To balance the incoming energy, the earth must radiate on average the same amount of energy back to space (Fig. 1), and 235 W m^{-2} corresponds to black body radiation at a temperature of $\sim -19^\circ\text{C}$ (254 K). Therefore the emitted thermal radiation occurs at about $10 \mu\text{m}$ in the infrared part of the electromagnetic radiation spectrum. Note that -19°C is much colder than the conditions at the earth's surface where the annual average global mean temperature is about 14°C , and is reached typically at an altitude of 5 km above the surface in midlatitudes.

Some of the infrared radiation leaving the atmosphere originates near the earth's surface and is transmitted relatively unimpeded through the atmosphere if there is no

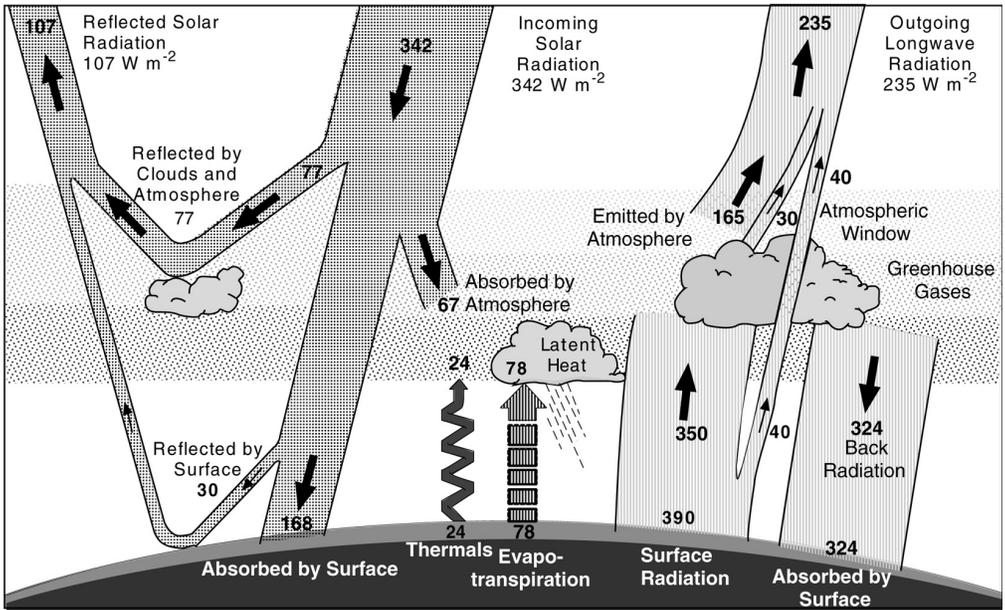


Figure 1. The earth's radiation balance. The net incoming solar radiation of 342 W m^{-2} is partially reflected by clouds and the atmosphere or at the surface, but 49% is absorbed by the surface. Some of that heat is returned to the atmosphere as sensible heating and most as evapotranspiration that is realized as latent heat in precipitation. The rest is radiated as thermal infrared radiation and most of that is absorbed by the atmosphere and re-emitted both upwards and downwards, producing a greenhouse effect, as the radiation lost to space comes from cloud tops and parts of the atmosphere much colder than the surface. From Kiehl and Trenberth (1997).

cloud, and if it is in the part of the spectrum known as the atmospheric 'window' (Fig. 1). The bulk of the radiation, however, is intercepted and re-emitted both upwards and downwards from much higher and colder levels out to space, producing the greenhouse effect. The emissions to space occur either from the tops of clouds at different atmospheric levels (which are almost always colder than the surface), or by greenhouse gases present in the atmosphere which absorb and emit infrared radiation. Most of the atmosphere consists of nitrogen and oxygen (99% of dry air), which are transparent to infrared radiation. In the current climate, for clear skies the natural greenhouse effect is accounted for by water vapour $\sim 60\%$, carbon dioxide 26%, ozone 8% and other gases 6% (Kiehl and Trenberth 1997).

Clouds also absorb and emit thermal radiation, and have a greenhouse effect, but clouds are also bright reflectors of solar radiation and thus also act to cool the surface. While on average there is strong cancellation between the two opposing effects of short-wave and long-wave cloud heating, the net global effect of clouds in our current climate, as determined by space-based measurements, is a small cooling of the surface. A key issue of how clouds will change as climate changes is complicated by the strong influence of particulate pollution, which can be independent of climate change (see below). If cloud tops get higher, the radiation to space from clouds is at a lower temperature and so this produces a warming. However, more extensive low clouds would be likely to produce cooling because of the greater influence on solar radiation.

Aerosols occur in the atmosphere naturally from, for instance, blowing dust or pollen grains. The eruption of Mt Pinatubo in the Philippines in June 1991 added considerable amounts of aerosol to the stratosphere that, for about two years, scattered

solar radiation leading to a loss of radiation and a cooling at the surface (Hansen *et al.* 1996). Human influences on scattering and absorbing aerosols come from emissions into the atmosphere from burning of fuels, and result in gases that are oxidized to become highly reflective micron-sized aerosols, such as the milky white sulphate aerosols, or strongly absorbing carbonaceous aerosols such as soot. These aerosols can make important differences to the number and size of cloud droplets, precipitation formation, within-cloud heating and cloud lifetime (e.g. Rosenfeld 2000; Ramanathan *et al.* 2001; Kaufman *et al.* 2002). Light-absorbing aerosols can short-circuit the hydrological cycle by depositing heat directly into a layer that would otherwise be heated indirectly by latent heating in precipitation, originating from heat absorbed at the surface and lost through evaporation (see also Menon *et al.* 2002). Aerosols are rapidly (a week or less) removed from the atmosphere through the natural hydrological cycle and dry deposition as they travel away from their source. Nonetheless, atmospheric concentrations can substantially exceed background conditions in large areas around and downwind of the emission sources (IPCC 2001).

The values in Fig. 1 will no doubt be slightly revised as knowledge improves, but the overall picture is fairly well established. In Fig. 1 the residual errors were lumped into the surface sensible-heat flux, which is probably too large, while absorption of incoming radiation in the atmosphere (e.g. by aerosols and water vapour, see Ackerman *et al.* (2003)) should probably be increased somewhat. Of particular note at the surface is that the net loss of heat from evapotranspiration exceeds the net loss by radiation, highlighting the vital role of moisture in atmospheric energetics.

Climate can vary for multiple reasons and, in particular, human activities can lead to changes in several ways. The energy flowing through the earth system on an annual mean basis (Fig. 1) averaged over the globe at 235 W m^{-2} is equivalent to $\sim 120 \text{ PW}$. On a global scale, even a 1% change in the energy flows, which is the order of the estimated change to date (IPCC 2001), dominates all other direct influences humans have on climate (Karl and Trenberth 2003), although the latter can be important locally. Hence, the main way human activities compete with nature is if they interfere with the natural flow of energy from the sun, through the climate system and back out to space. This is achieved through changes in the composition of the atmosphere, especially generation of carbon dioxide through fossil fuel burning (see Karl and Trenberth (2003) for a more complete discussion), although changes in land use are important locally.

3. DISTRIBUTION OF RADIATION

For the earth, on an annual mean basis, there is an excess of solar over OLR in the tropics and a deficit at mid- to high latitudes that sets up an equator-to-pole temperature gradient. This, together with the earth's rotation, results in a broad band of westerlies in each hemisphere in the troposphere; embedded within these are large-scale baroclinic weather systems that, along with the ocean, act to transport heat polewards to achieve an overall energy balance, as described below.

The annual mean radiation and energy fluxes are given in Trenberth and Stepaniak (2003b). Here we focus mostly on the annual cycle, and present the values for the solstitial seasons of December–January–February (DJF) and June–July–August (JJA), and their differences. Mean values of ASR, OLR, and the net radiation are given in Figs. 2, 3 and 4. At the surface, deserts, snow and ice have a high albedo, and consequently absorb little incoming radiation. However, the main departures in the ASR (Fig. 2) from what would be expected simply from the sun–earth geometry are the signatures of persistent clouds. Bright clouds occur over Indonesia and Malaysia, across

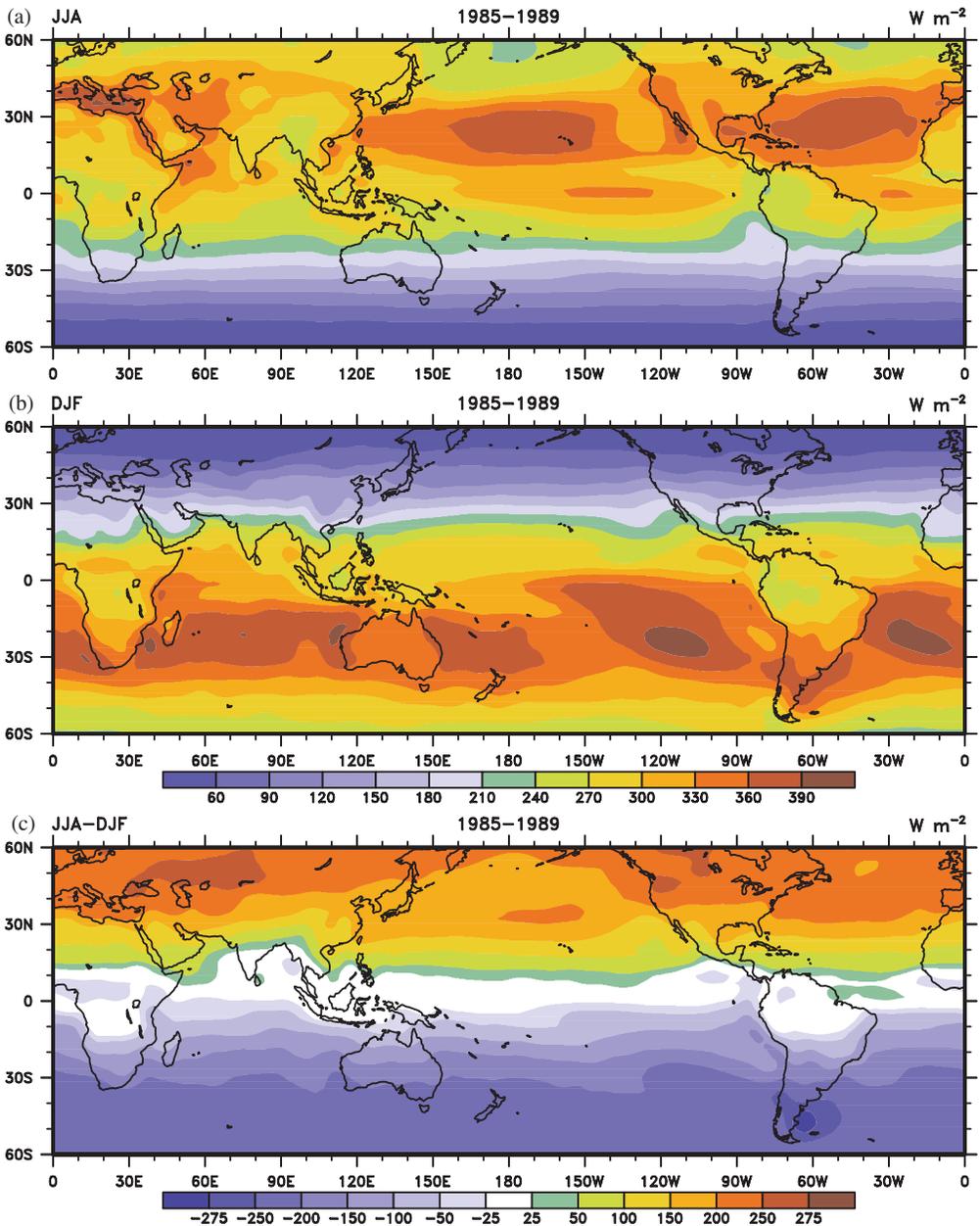


Figure 2. Top of atmosphere (TOA) absorbed solar radiation for the period of the Earth Radiation Budget Experiment, 1985-89 averaged for: (a) June, July and August (JJA), (b) December, January and February (DJF), and (c) their difference JJA-DJF. The contour interval in (a) and (b) is 30 W m^{-2} and in (c) is 25 or 50 W m^{-2} .

the Pacific near 10°N , and over the Amazon in the southern summer, contributing to the relatively low values in these locations; while dark oceanic cloud-free regions along and south of the equator in the Pacific and Atlantic and in the subtropical anticyclones absorb most solar radiation. The dominant annual cycle forcing is the change in distribution of incoming solar radiation due to the orbit of the earth around the sun. Figure 2 serves as an important reminder of just how large the seasonal variation is, because JJA versus

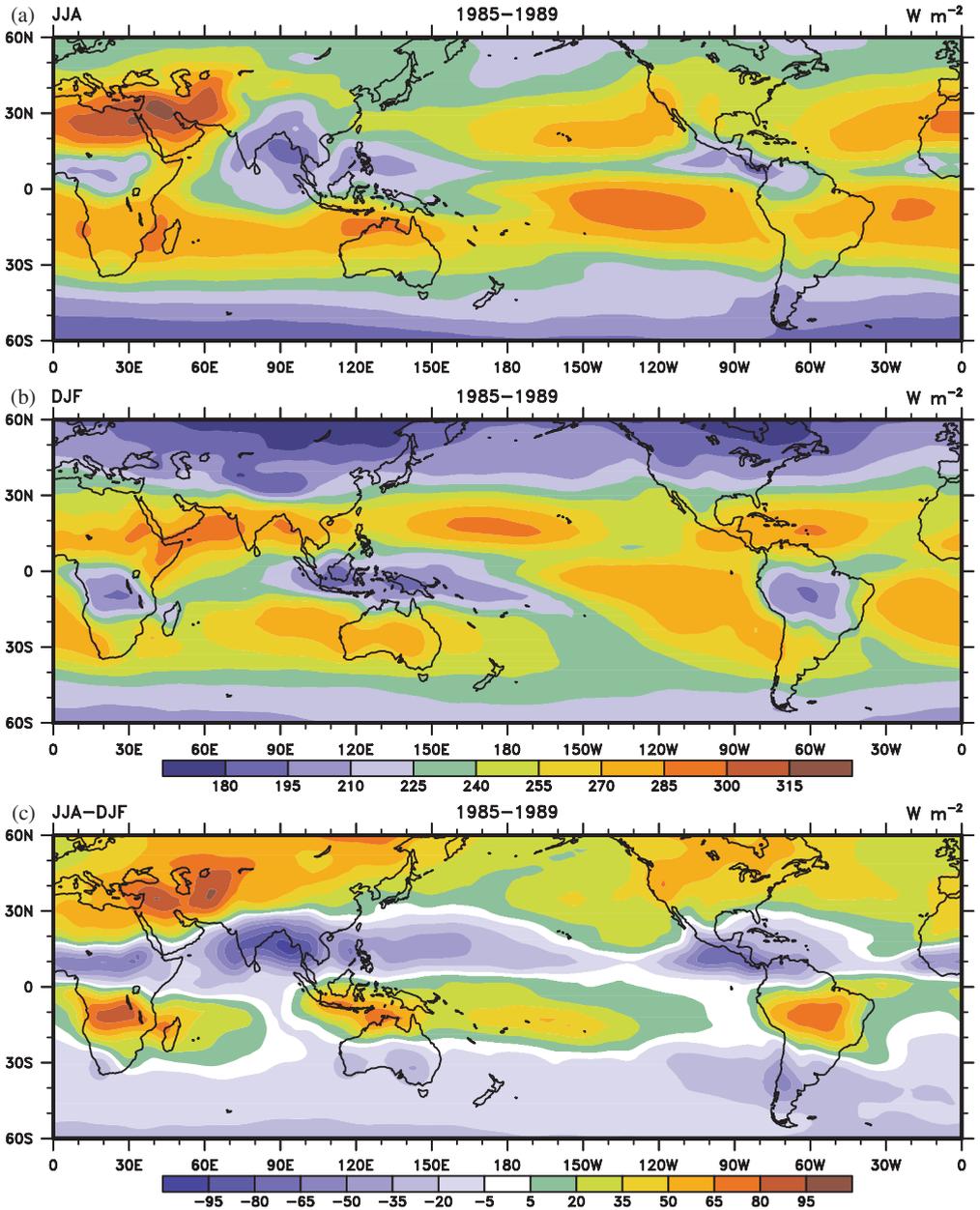


Figure 3. As Fig. 2, but for TOA outgoing long-wave radiation. The contour interval is 15 W m^{-2} .

DJF differences exceed 200 W m^{-2} . It is interesting that the zero line is not on the equator, but instead lies between 0 and 10°N for the most part.

On the other hand, OLR (Fig. 3) is more uniform with latitude and season, and it is well established that it varies primarily with deep convection, owing to the high cold cloud tops. Similarly, the dry cloud-free regions are where the most surface radiation escapes to space. The signatures of the Intertropical Convergence Zone (ITCZ), South Pacific Convergence Zone (SPCZ) and South Atlantic Convergence Zone (SACZ), and

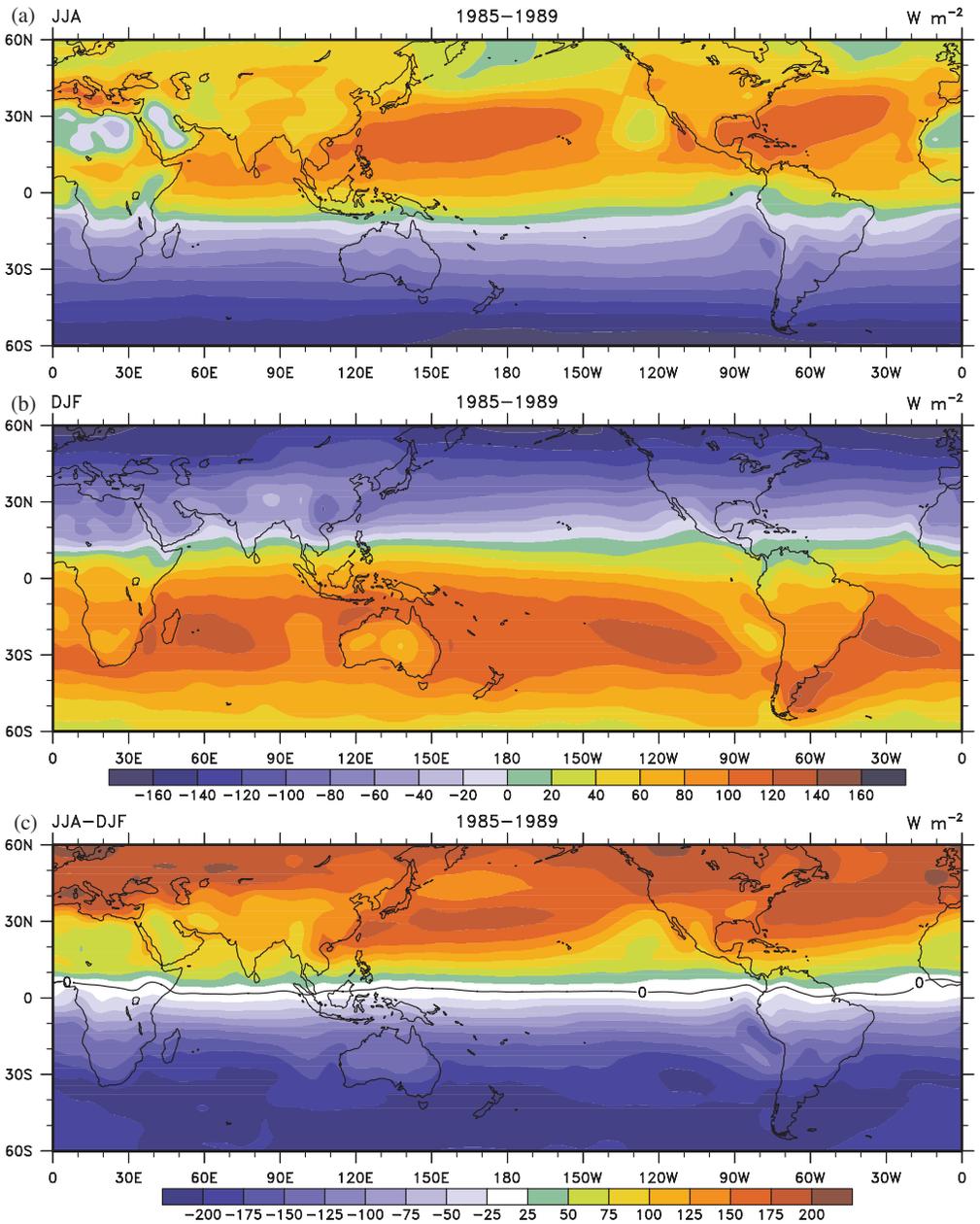


Figure 4. As Fig. 2, but for TOA net radiation. The contour interval in (a) and (b) is 20 W m^{-2} and in (c) is 25 W m^{-2} .

monsoon-related clouds are clear. Also, in the tropics there is a very strong relation between the JJA-DJF differences in OLR and the patterns of change of precipitation and vertical motion (not shown). In the extratropics, however, the OLR signature is more related to temperature.

The cloud signature in ASR is well matched by that in the OLR signal, and when the net radiation (Fig. 4) is considered there is remarkable but well-known cancellation

(Hartmann *et al.* 2001). In particular, the high convective clouds are bright and reflect solar radiation, but they are also cold and hence reduce OLR. The main remaining signature of clouds in the net radiation from earth is seen from the low stratocumulus cloud decks that persist above cold ocean waters, most notably off the west coasts of California and Peru. Such clouds are also bright but, as they have low tops, they radiate at temperatures close to those at the surface, resulting in a cooling of the planet. Note that the Sahara desert has a high OLR, consistent with dry cloud-free and warm conditions, but it is also bright and reflects solar radiation, and it stands out as a region of net radiation deficit. The net radiation seasonally is dominated by the ASR changes. Once again the zero line in the net radiation lies just north of the equator. It is this pattern of net radiation that ultimately directly drives the monsoons and their seasonal reversals.

4. HEAT STORAGE AND THE ROLE OF THE CLIMATE SYSTEM COMPONENTS

The different components of the climate system contribute on different time-scales to climate variations and change. The atmosphere and oceans are fluid systems and can move heat around through convection and advection, in which the heat is carried by the currents, whether small-scale short-lived eddies, large-scale atmospheric jet streams or ocean currents. However, many facets of the climate are determined simply by the heat capacity of the different components of the climate system. The total heat capacity depends on the mass of each substance involved, as well as its capacity for holding heat as measured by its specific heat.

The atmosphere does not have much capability to store heat. For the ocean, for a typical salinity of ~ 35 PSU, the density ρ is ~ 1028 to 1025 kg m^{-3} and specific heat $C_p \sim 3985$ to $3995 \text{ J kg}^{-1}\text{K}^{-1}$ for temperatures from 2 to 20°C ; the product ρC_p is more nearly constant than either of the two components. The heat capacity of the global atmosphere corresponding to a global mean surface pressure of 985.5 hPa (Trenberth and Smith 2004) and C_p of $1004 \text{ J kg}^{-1}\text{K}^{-1}$ thus corresponds to that of the ocean to a depth of 2.46 m or, taking the ocean area into account, 3.5 m. However, the depth of ocean actively involved in climate is much greater than that. Similarly, the ice sheets and glaciers do not play a strong role, while sea ice is important where it forms.

(a) *The atmosphere*

Energy in the atmosphere comes in several forms; basic equations are outlined extensively in Lorenz (1967). The incoming radiant energy is absorbed and transformed into sensible or latent energy. At the land surface heat may be manifested as increases in temperature and/or evaporation, and the partitioning depends on the available moisture and nature of vegetative ground cover. The hydrological cycle involves the transfer of water from the oceans to the atmosphere, to the land, and back to the oceans both on top of and beneath the land surface, and constitutes a key part of the energy cycle. In the tropics in summer land warms relative to the ocean, and sets the stage for monsoon development. Water is evaporated from the ocean surface, cooling the ocean. As water vapour, and thus latent energy, it is transported perhaps thousands of kilometres before it is involved in clouds and weather systems, and precipitated out as rain, snow, hail or some other frozen pellet back to the earth's surface. In that process it provides latent heat to the atmosphere.

Increases in temperature increase the internal energy of the atmosphere, which also causes it to expand. Consequently, it changes the altitude of the air and increases the potential energy. There is, therefore, a close relationship between internal and potential energy in the atmosphere, which together can be expressed as enthalpy, or sensible heat.

Enthalpy is the sum of the internal energy and the product of the pressure and volume of the system. Conservation of energy requires that the change in internal energy is equal to the heat transferred to, less the work done by, the system, and is the enthalpy if the work done changes the volume at constant pressure. Once air starts to move, usually because temperature gradients give rise to pressure gradients that in turn cause wind, some energy is converted into kinetic energy. The total energy is the sum of the internal, potential, kinetic and latent energies and is a constant in the absence of diabatic processes. Transport of energy also requires full consideration of the work done, and is addressed for the atmosphere in section 5. However, the possibilities for conversions among all of these forms of atmospheric energy are a key part of what provides the richness of atmospheric phenomena.

(b) *The oceans*

The oceans cover 70.8% of the earth's surface, although with a much greater fraction in the southern hemisphere (SH, 80.9% of the area) than in the northern hemisphere (NH, 60.7%); through their fluid motions and high heat capacity they have a central role in shaping the earth's climate and its variability. The average depth of the ocean is 3795 m. The oceans are stratified in the opposite sense to the atmosphere, with warmest waters near the surface. Consequently, in the ocean convection arises from cooling at the surface and transport of heat upwards through colder and denser waters sinking and being replaced by lighter more buoyant waters. Another vital factor in convection is the salinity of the waters, because this also affects density. Consequently the densest waters are those that are cold and salty; these are found at high latitudes where not only is there a deficit in radiation and atmospheric temperatures are low, but also the formation of sea ice leads to a rejection of brine and increases the salinity of the surrounding waters.

The ocean surface cools through radiation, sensible heating of the atmosphere, and especially through moistening of the atmosphere in which the latent heat of vaporization cools the ocean, while eventual precipitation of atmospheric water vapour produces latent heating. Surface heating in summer warms the upper ocean, which becomes strongly stratified (as warm water is less dense) even though some heat is mixed to lower layers by mechanical stirring by winds. In winter, surface cooling results in denser surface waters which are apt to sink, triggering convection, and bringing warmer waters to the surface from below. Hence the oceans take up heat in summer and release it in winter and consequently the oceans are a great moderating effect on climate variations, especially changes such as those involved with the annual cycle of the seasons.

The mixed layer on average involves ~90 m of ocean. The thermal inertia of the ocean depends on the rate of ventilation of water between the mixed upper ocean layers and the deeper more isolated layers through the thermocline. Such mixing is not well known and varies greatly geographically. An overall estimate of the delay in surface temperature response caused by the oceans is 10 to 100 years. The slowest response should be in high latitudes where deep mixing and convection occur, and the fastest response is expected in the tropics.

It is difficult to make accurate estimates of the annual cycle in ocean heat content. Observations of monthly mean subsurface ocean heat content contain substantial uncertainties owing to poor sampling (cf. Levitus and Antonov (1997) for instance). Estimates of surface fluxes using bulk flux parametrizations also contain substantial sampling uncertainties and biases, and large imbalances occur when globally averaged (Trenberth *et al.* 2001). Accordingly, we have made use of the net surface fluxes computed as a residual of the TOA radiative fluxes and the divergence of atmospheric energy, as

described in the next section (see Fig. 5). This should be close to zero over land but it is not, and average rms values are $\sim 30 \text{ W m}^{-2}$ (Trenberth *et al.* 2001). The complexity of mountains and steep topography are evidently sources of problems that are likely to enhance the errors over land. Over the ocean, the net surface flux (Fig. 5) corresponds to the average divergence of ocean heat transports (plus residual errors, which are believed to be of order 20 W m^{-2} but are mostly randomly distributed). The implied ocean heat transports have been verified by comparisons with direct measurements in ocean sections (Trenberth and Caron 2001).

To convert these fluxes into contributions from ocean heat storage, uptake and release during the annual cycle, and ocean heat transport divergence, requires additional information on one component or the other. Changes in annual mean ocean heat storage, while important for sea-level rise and climate change, are still very small compared with the annual cycle. Therefore the annual mean surface flux (Fig. 5) is mostly balanced by ocean heat transports (as given in Fig. 6). There are substantial uncertainties in the SH where data are not as good as elsewhere, and the patterns are seen to be more irregular and even physically unlikely in some parts of the southern oceans. In the extratropics, the divergence of ocean heat transports does not change much with the seasons (Jayne and Marotzke 2001), and the departure of the seasonal mean from the annual mean provides an estimate of the ocean heat uptake and release (Fig. 5). The estimate of the annual cycle of the zonal mean net surface heat flux (not shown) for the ocean reveals about 160 W m^{-2} into and out of the midlatitude northern oceans and 120 to 140 W m^{-2} in the SH (Fig. 5). Of course the areal extent of the latter is much greater.

Surface cooling over the oceans is greatest from cold dry continental air outbreaks over western oceans, where boundary currents such as the Gulf Stream lie and turbulent surface fluxes can exceed 1000 W m^{-2} for short periods of time (e.g. Neiman and Shapiro 1993) and over 300 W m^{-2} when averaged over months and seasons. The solstitial extreme seasonal departures from the annual mean (Fig. 5) show how the seasonal northern ocean surface fluxes are localized along the east coasts over the Kuroshio and Gulf Stream regions and exceed 240 W m^{-2} , more than double the values over the eastern Pacific and Atlantic. These are on top of the annual mean values, which are about 140 W m^{-2} (Fig. 5) in these regions.

Wind blowing on the sea surface drives the large-scale ocean circulation in its upper layers. The ocean currents carry heat (Fig. 6) and salt along with the fresh water around the globe. Large seasonal changes in Ekman transports lead to a substantial annual cycle in poleward heat transports, especially throughout the tropics (Jayne and Marotzke 2001), that is stronger in summer than winter. This slightly enhances the annual cycle of ocean heat content, given by Antonov *et al.* (2004), over what occurs from surface fluxes alone in midlatitudes. The oceans store heat, absorbed at the surface, for varying durations and release it in different places, thereby ameliorating temperature changes over nearby land and contributing substantially to variability of climate on many time-scales. Additionally, the ocean thermohaline circulation, which is the circulation driven by changes in sea water density arising from temperature (thermal) or salt (haline) effects, allows water from the surface to be carried into the cold deep abyssal ocean where it is isolated from atmospheric influence and turns over only very slowly on time-scales of hundreds to thousands of years. Hence the ocean may sequester heat for periods of a thousand years or more.

The main energy transports in the ocean (Fig. 6) are those of heat associated with overturning circulations as cold waters flow equatorwards at some depth while the return flow is warmer near the surface; see for instance Bryden and Imawaki (2001). There is a small poleward heat transport by gyre circulations whereby western boundary currents,

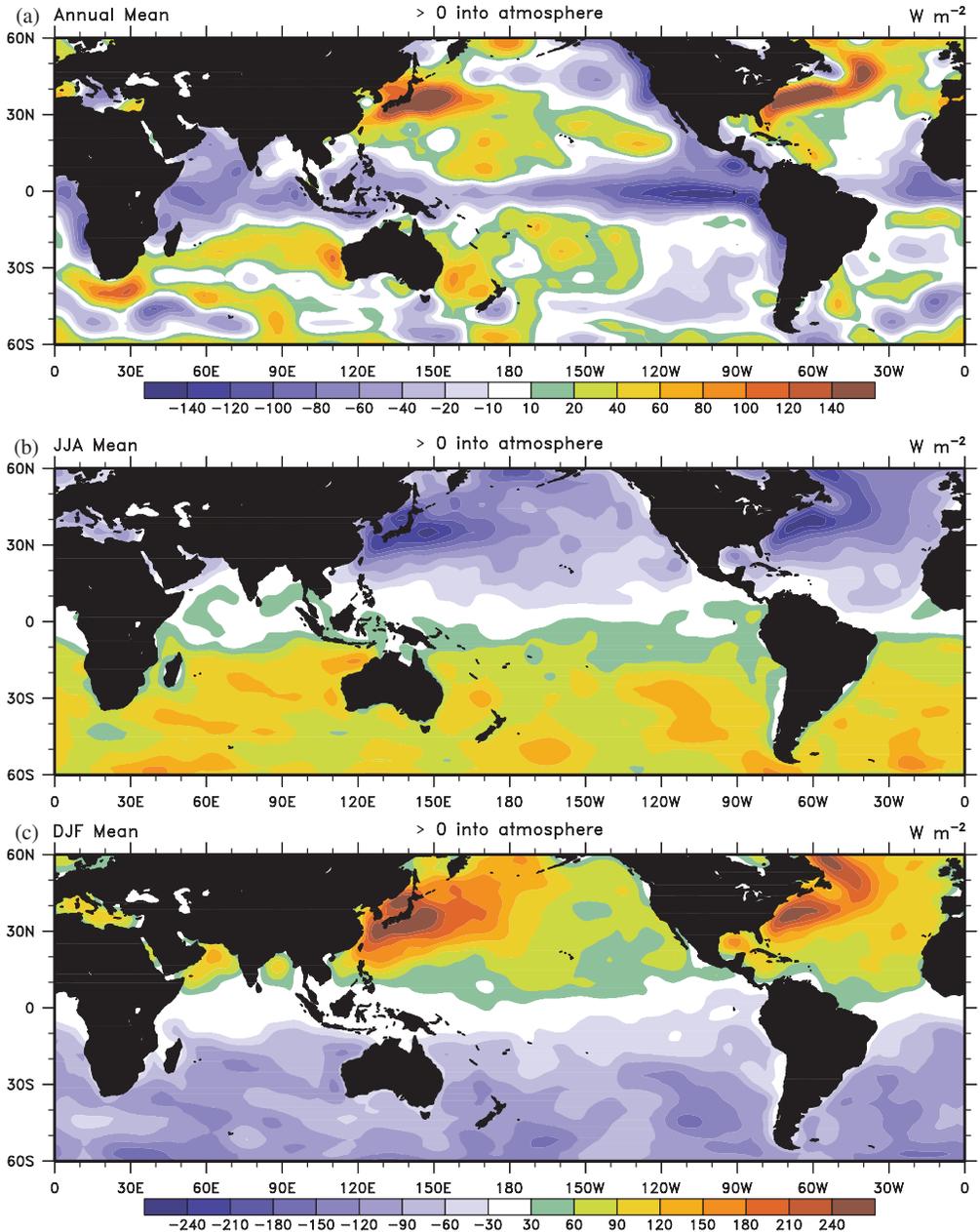


Figure 5. (a) Deduced net annual mean surface heat flux into the ocean surface for the Earth Radiation Budget Experiment (ERBE) period in W m^{-2} , based on Trenberth *et al.* (2001). (b) Estimated departures from this annual mean averaged over June, July and August (JJA), and (c) as (b) but averaged over December, January and February (DJF). The contour interval in (a) is 20 W m^{-2} and in (b) and (c) is 30 W m^{-2} .

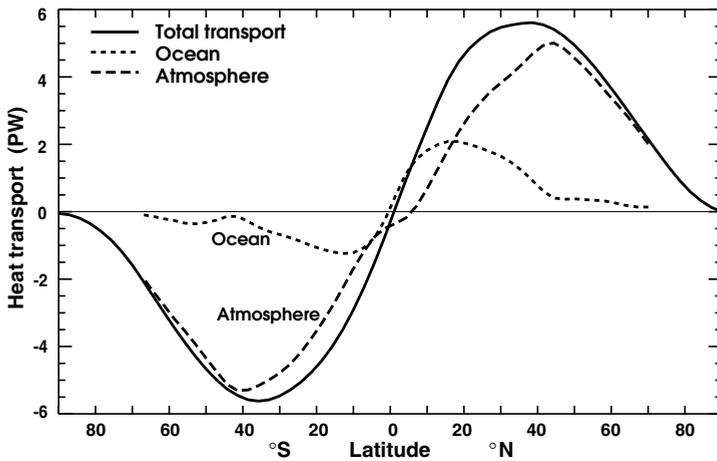


Figure 6. Poleward heat transports by the atmosphere (large-dashed line), ocean (small-dashed line) and the two combined (solid line) in petawatts (from Trenberth and Caron 2001).

such as the Gulf Stream in the North Atlantic or the Kuroshio in the North Pacific, move warm waters polewards while part of the return flow comprises colder currents in the eastern oceans. However, a major part of the Gulf Stream return flow is at depth, and the most pronounced thermohaline circulation is in the Atlantic Ocean. In contrast, the Pacific Ocean is fresher and features shallower circulations. A key reason for the differences lies in salinity. Because there is a net atmospheric water vapour transport in the tropical easterly trade winds from the Atlantic to the Pacific across the Central American isthmus, the North Atlantic is much saltier than the North Pacific.

(c) Land

Heat penetration into land from surface fluxes of energy occurs mainly through conduction except where water plays a role, so that heat penetration is limited and slow. The specific heat of dry land is roughly a factor of four-and-a-half less than that of sea water (for moist land the factor is probably closer to two). Moreover, heat penetration into land is limited and only the top 2 metres or so typically play an active role (as a depth for most of the variations on annual time-scales, say). Accordingly, land plays a much smaller role in the storage of heat and in providing a memory for the climate system. Temperature profiles taken from bore holes (Pollack *et al.* 1998) into land (or ice caps), therefore, provide a blurry coarse estimate of temperatures in years long past. The land surface encompasses an enormous variety of topographical features and soils, differing slopes (which influence runoff and radiation received) and water capacity. Changes in phase of water, from ice to liquid to water vapour, affect the storage of heat. Soil moisture and surface waters can act through evaporative cooling to moderate temperatures, but the role of moisture is limited by its availability either in soils or through transpiration from plants. Plants can act to pump moisture out of the root zone into the leaves where it can be released into the atmosphere as the plant participates in photosynthesis. Changes in vegetation alter how much sunlight is reflected. Sensible-heat fluxes are relatively modest (Fig. 1) and evaporative fluxes provide more net cooling than long-wave radiation.

Surface air temperature changes over land occur much faster and are much larger than over the oceans for the same heating, and the observed variability of temperatures

over land is a factor of two to six greater than that over the oceans (Barnett 1978). In the extratropics over land in winter there is often a strong surface temperature inversion because of the cold land surface, and it makes for a very stable layer of air that can trap pollutants. The strength of an inversion is very sensitive to the amount of stirring in the atmosphere and can be greatly affected by human activities (e.g. Weller 1982). In contrast, over the oceans surface fluxes of heat into the atmosphere keep the air temperature within a narrow range. Thus, it is not surprising that over land month-to-month persistence in surface temperature anomalies is greatest near bodies of water (van den Dool *et al.* 1986). Consequently, for a given heating perturbation, the response over land should be much greater than over the oceans; the atmospheric winds are the reason why the observed factor is only in the range two to six.

(d) Ice

Major ice sheets, like those over Antarctica and Greenland, potentially have a large heat capacity but, like land, the penetration of heat occurs primarily through conduction so that the mass involved in changes from year to year is small. Temperature profiles can be taken directly from bore holes into ice, and an estimate is that terrestrial heat flow is $51 \times 10^{-3} \text{ W m}^{-2}$ (Dahl-Jensen *et al.* 1998). On century time-scales, however, the ice sheet heat capacity becomes important. Unlike land the ice can melt, which has major consequences through changes in sea level on longer time-scales.

Sea ice forms in winter, reaches its maximum extent about the equinox—in March in the NH or September in the SH—and declines in thickness and extent in summer. The changes in phase involve the latent heat of fusion, and thus sea ice adds additional memory to the climate system in regions where it forms. Because the surface fluxes in Fig. 5 are a residual computation, they apply just as much in regions of sea ice as over the open ocean, although they are at the limits of the plots in Fig. 5.

5. HEAT AND ENERGY TRANSPORTS IN THE ATMOSPHERE

As we have seen, incoming radiant energy is absorbed. Convection and other processes act to redistribute energy in the vertical within the atmosphere and ocean. However, of central interest here is the lateral transport of energy, especially polewards, to balance the radiative forcing (Figs. 4 and 6). Hence, the atmospheric and ocean circulations are key players. In this section the atmospheric transports are quantified and processes examined.

In the atmosphere, in addition to advection of internal, potential, latent and kinetic energy, transport of energy also involves work done, and accordingly the total energy transport is most naturally partitioned into dry static energy (*DSE*), latent energy (*LE*), and kinetic energy (*KE*). The *DSE* consists of the sensible heat (*SH*) plus potential energy (*PE*), i.e. $DSE = SH + PE$, and these are combined as they are closely related. The moist static energy (*MSE*) includes the latent component: $MSE = DSE + LE$. The total transport of energy, \mathbf{F}_A , is a vector and also includes the kinetic energy, although meridional transports of the latter are fairly small; see Trenberth and Stepaniak (2003a) for full expressions. The divergence of vertically integrated transports is balanced by diabatic forcings and, ignoring tendencies, consists of atmospheric diabatic heating Q_1 minus the friction heating Q_f (where Q_f is very small, and henceforth ignored), and column moistening $-Q_2$. The column latent heating of the atmosphere (negative of moistening) is given by $Q_2 = L(P - E_s)$, where P is the net precipitation, L is the latent heat of condensation and E_s is the surface evaporation (denoting surface fluxes with a subscript *s*). Surface fluxes consist of upwards sensible heat, H_s , and surface

radiation, R_s , directed downwards, then the upward net surface flux of energy $F_s = H_s + LE_s - R_s$. The diabatic heating can be written as $Q_1 = R_T + F_s + L(P - E_s)$, where R_T is the downward radiation at the TOA. From the standpoint of atmospheric energy transport it is not Q_1 that matters but $Q_1 - Q_2$ because it includes the latent moist component and, ignoring the tendency terms:

$$\nabla \cdot \mathbf{F}_A = Q_1 - Q_2 = R_T + F_s.$$

In the atmosphere most of the transport occurs as *MSE*, although it is only in lower latitudes that *LE* contributes substantially. Trenberth and Stepaniak (2003a) document the global vertically integrated atmospheric heat budget as zonal means for the annual mean, mean annual cycle and the interannual variability for 1979–2001, based upon National Centers for Environmental Prediction/National Center for Atmospheric Research re-analyses. They further examine the transports and their divergences partitioned into contributions from within-month transients and quasi-stationary components; in the tropics the latter correspond to the large-scale overturning global monsoon and the embedded Hadley and Walker circulations. Zonal asymmetries are also important (Cook 2003) and arise from the distribution of land and surface topography in particular, and give rise to regional monsoons and subtropical anticyclones (Trenberth *et al.* 2000; Chen *et al.* 2001; Rodwell and Hoskins 2001). Trenberth and Stepaniak (2003b) present the geographic distributions of transports and their divergences from the annual mean. The quantities Q_1 , Q_2 and $Q_1 - Q_2$ are computed as residuals from the heat and moisture budgets in the atmosphere and the tendency terms are included (Trenberth *et al.* 2001; Trenberth and Stepaniak 2003a,b). Here we present results for DJF and JJA, and their differences.

Before considering the energy transports in detail, we examine the deduced diabatic heating sources and sinks and latent energy forcings of the atmosphere. It is often tempting to take these as forcings of the atmosphere but, as we shall see, they are just as much a consequence of the circulation as a cause.

Estimates of vertically integrated diabatic heating in the atmosphere Q_1 (Fig. 7) and latent heating component Q_2 (Fig. 8) reveal from the similarity of patterns that the dominant diabatic heating in the tropics comes from latent heating associated with precipitation. The heating is associated with regions of upward motion and the low-level convergence of moisture by the atmospheric circulation (see Fig. 8), as indicated by the vertically integrated divergent component of the total flux of water vapour transport (Trenberth and Stepaniak 2003a,b). These figures illustrate the atmospheric diabatic latent heating associated with the monsoons over Southeast Asia, Africa north of the equator, and central America in JJA, and northern Australia, Africa south of the equator, and northern South America in DJF, that are highlighted in the difference map. The latter also reveals the extensions over the oceans that make up key parts of the Hadley circulation.

The difference between these two quantities, $Q_1 - Q_2$, is directly related to the divergence of the total atmospheric energy transports (Fig. 9) on an annual basis when changes in storage of energy in the atmosphere can be ignored. If we now further consider the net radiation (Fig. 4) versus the net atmospheric heating (where this includes the latent component; Fig. 9), then for long-term annual means the difference between them gives the net surface flux of energy (F_s ; Fig. 5), discussed in detail in section 4.

The re-analyses of the atmospheric data reveal that the atmosphere is primarily responsible for the energy transports that compensate for the net radiation imbalances (Fig. 6). Annual mean poleward transports of atmospheric energy of 5.0 PW peak at

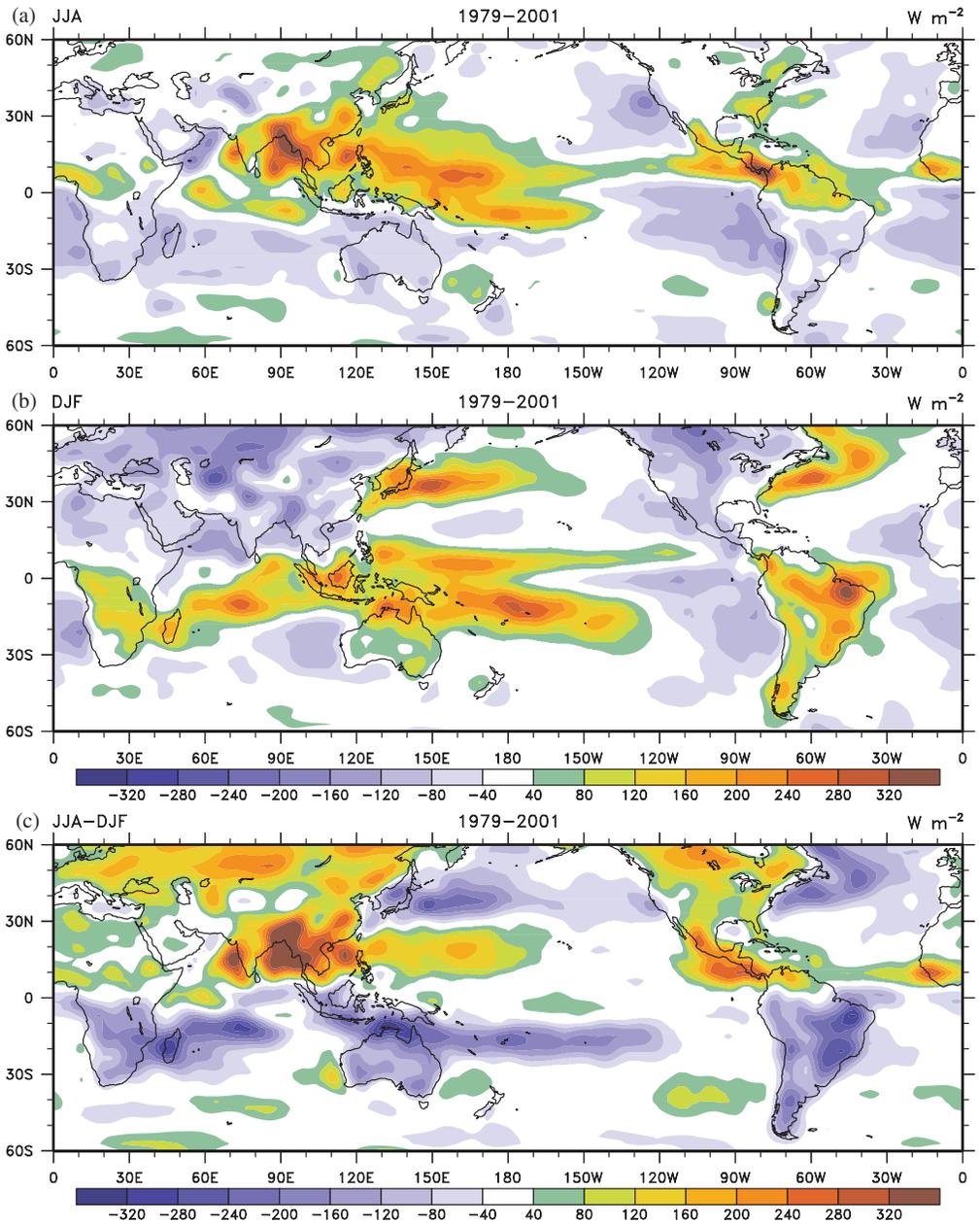


Figure 7. Vertically integrated atmospheric diabatic heating for 1979–2001 averaged over: (a) June, July and August (JJA), (b) December, January and February (DJF), and (c) their difference JJA–DJF. The contour interval is 40 W m^{-2} .

43°N , with similar values near 40°S . Poleward ocean heat transports, inferred as a residual but which agree with direct ocean estimates within error bars of order 0.3 PW (Trenberth and Caron 2001), are dominant only between 0 and 17°N . At 35° latitude, where the peak total poleward transport in each hemisphere occurs, the atmospheric transport accounts for 78% of the total in the NH and 92% in the SH; however, the poleward atmospheric transport peaks near 40° latitude (Trenberth and Caron 2001).

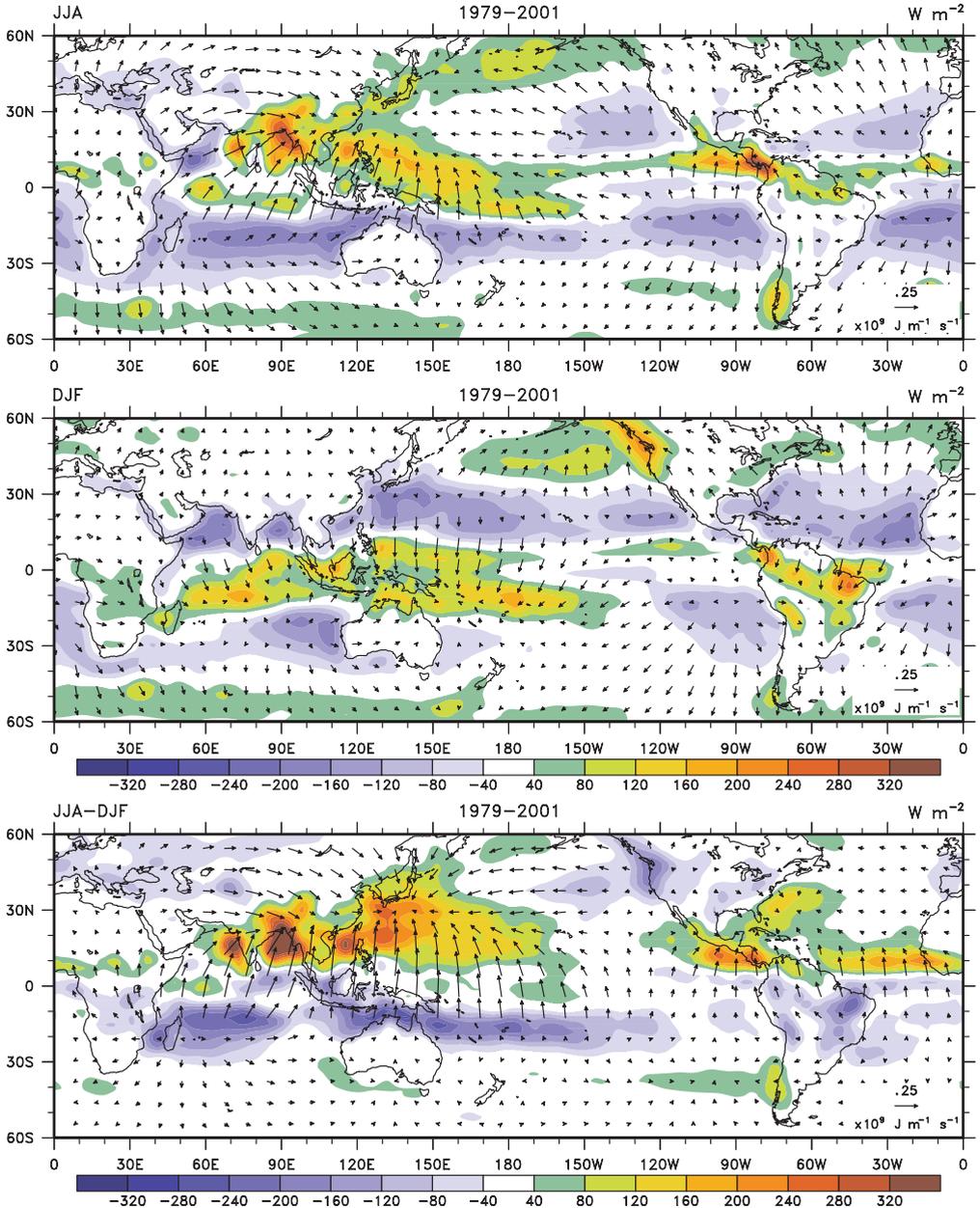


Figure 8. Vertically integrated atmospheric latent heating, $L(P - E_s)$, for 1979–2001 averaged over: (a) June, July and August (JJA), (b) December, January and February (DJF), and (c) their difference JJA–DJF. The contour interval is $40 W m^{-2}$. Also plotted in each case is the divergent component of the latent energy transports as vectors in $10^9 W m^{-1}$, with a reference vector in each lower right-hand corner. See text for details.

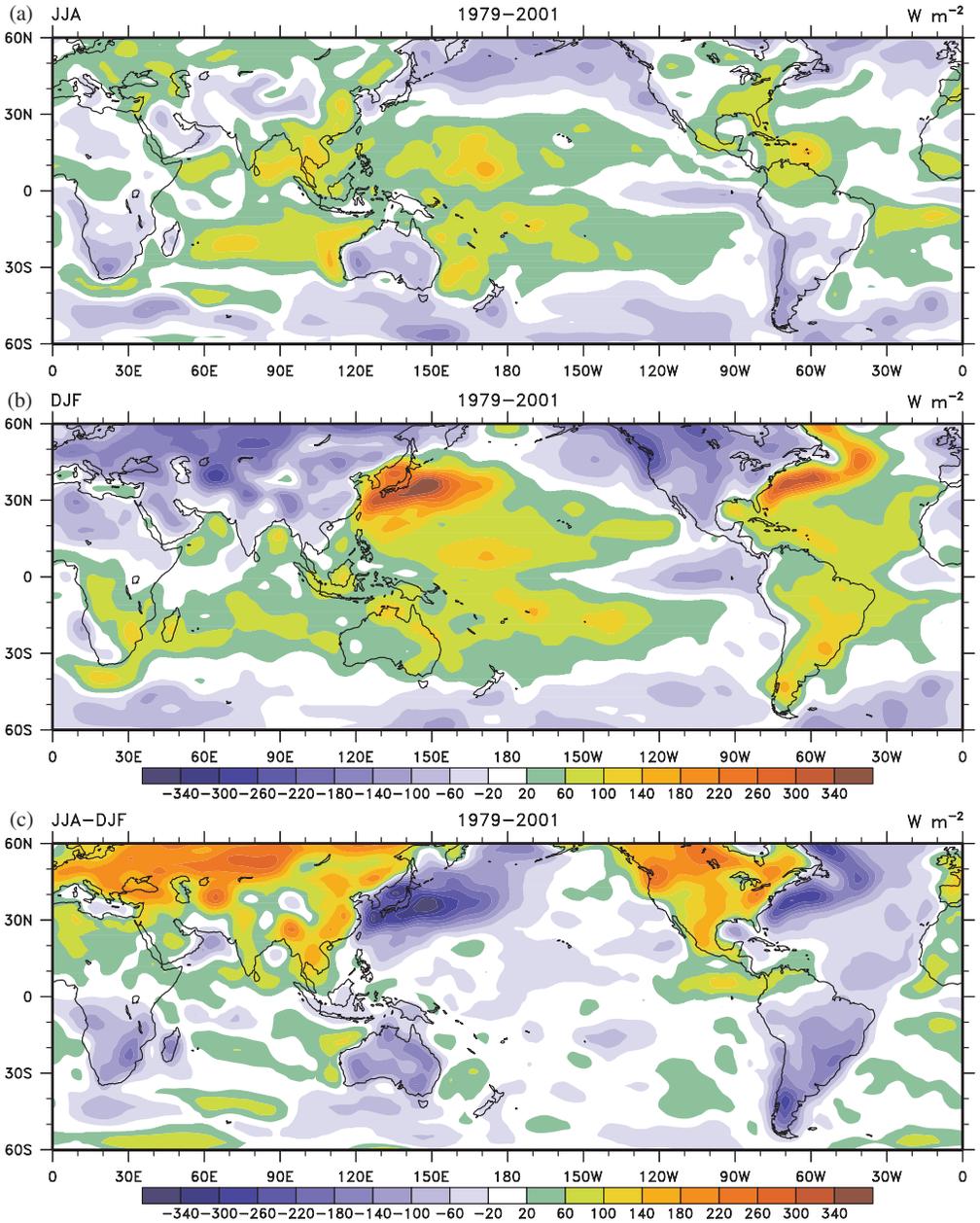


Figure 9. Vertically integrated atmospheric heating, $Q_1 - Q_2$, for 1979–2001 averaged over: (a) June, July and August (JJA), (b) December, January and February (DJF), and (c) their difference JJA–DJF. The contour interval is 40 W m^{-2} . See text for details.

Note that the zonal average atmospheric poleward energy transport is seamless, a point we return to later.

Differences between Figs. 7 and 8 reveal the heating over extratropical land in summer and the land–sea contrasts; these can be seen especially well in Fig. 9 and particularly in Fig. 9(c) that highlights the seasonal variations. Note how values over the continental land masses change sign across the equator. Meanwhile values over the

ocean tend to have the opposite sign to values over the land and also reverse across the equator, indicating the role of heat storage discussed in section 4(b).

The atmospheric circulation naturally responds to the diabatic forcings and, at the same time, transports heat, energy and moisture. The low-frequency part of the flow plays a dominant role throughout the tropics and can be depicted by monthly means, while transients, depicted by within-month variations, are prominent in the extratropics. The divergent component of the transports and the total energy divergence are given in Fig. 10 for the quasi-stationary component and Fig. 11 for the transient component (cf. Trenberth and Stepaniak 2003b). The former dominates in the tropics, and is more directly associated with the monsoons and the Hadley and Walker circulations. But note how the pattern of diabatic heating associated with latent heating has largely disappeared in Figs. 9(a) and (b) and instead the moisture sources are emphasized. In this depiction, the energy transport by the overturning circulations (Fig. 10) is fairly modest, although the Walker circulation transports are particularly evident in DJF. The overturning transports are seen more clearly in the *DSE* than in the *MSE*, because of large compensation that occurs in the latter as moisture fluxes are converted into sensible heat through release of latent heat (Trenberth and Stepaniak 2003b). In the Asian summer monsoon (Fig. 10) the divergence of energy from the upward branch is transported mostly to 30°S in the Indian Ocean, from where it is further transported polewards by the transient eddies (Fig. 11). In the Pacific and Atlantic the overturning transports energy northwards to about 30°N (Fig. 10), from where it is transported farther north in the ocean storm tracks (Fig. 11). In these areas, there is also a component transported into the SH.

In DJF and in the inter-seasonal differences (JJA–DJF) the strong NH stationary waves dominate (Fig. 10) although the storm tracks over the ocean also play a major role (Fig. 11). Note how the storm track activity shifts equatorwards and becomes stronger in winter, and with divergences of order 200 W m^{-2} near 30°N and over 100 W m^{-2} near 30°S over parts of the oceans. The difference also reveals the huge switch over the northern oceans from being a source of heat in winter to a sink in summer, while the land acts in the reverse sense.

6. DISCUSSION

From the figures in the previous section, it is clear that, from an energy budget standpoint, latent heating drives the upward branch of the monsoonal overturning circulation. There is a strong balance between adiabatic cooling, associated with rising air, and latent heating. Also, the latent heating is a response to the circulation itself and not a fundamental cause. The original source of the latent heating is the moisture convergence, and thus ultimately evaporation from the surface primarily over the subtropical oceans. Hence the latent heating is linked to the absorbed solar radiation. However, the latter is taken up and stored within the ocean, so that maximum sea surface temperatures tend to occur at the end of the summer season in the extratropics. Moreover, some aspects of the evaporation of moisture into the atmosphere are linked to its transport and convergence into precipitation zones associated either with the ITCZ, SPCZ, SACZ or monsoon rains. This is because the evaporation itself is also partly a result of the circulation and the turbulent fluxes into the atmosphere. In order for all this to take place, there has to be an overturning circulation in the atmosphere.

In the tropics, where the earth's rotational effects are weak, much of the movement of energy occurs through large-scale overturning of the atmosphere, which implies that

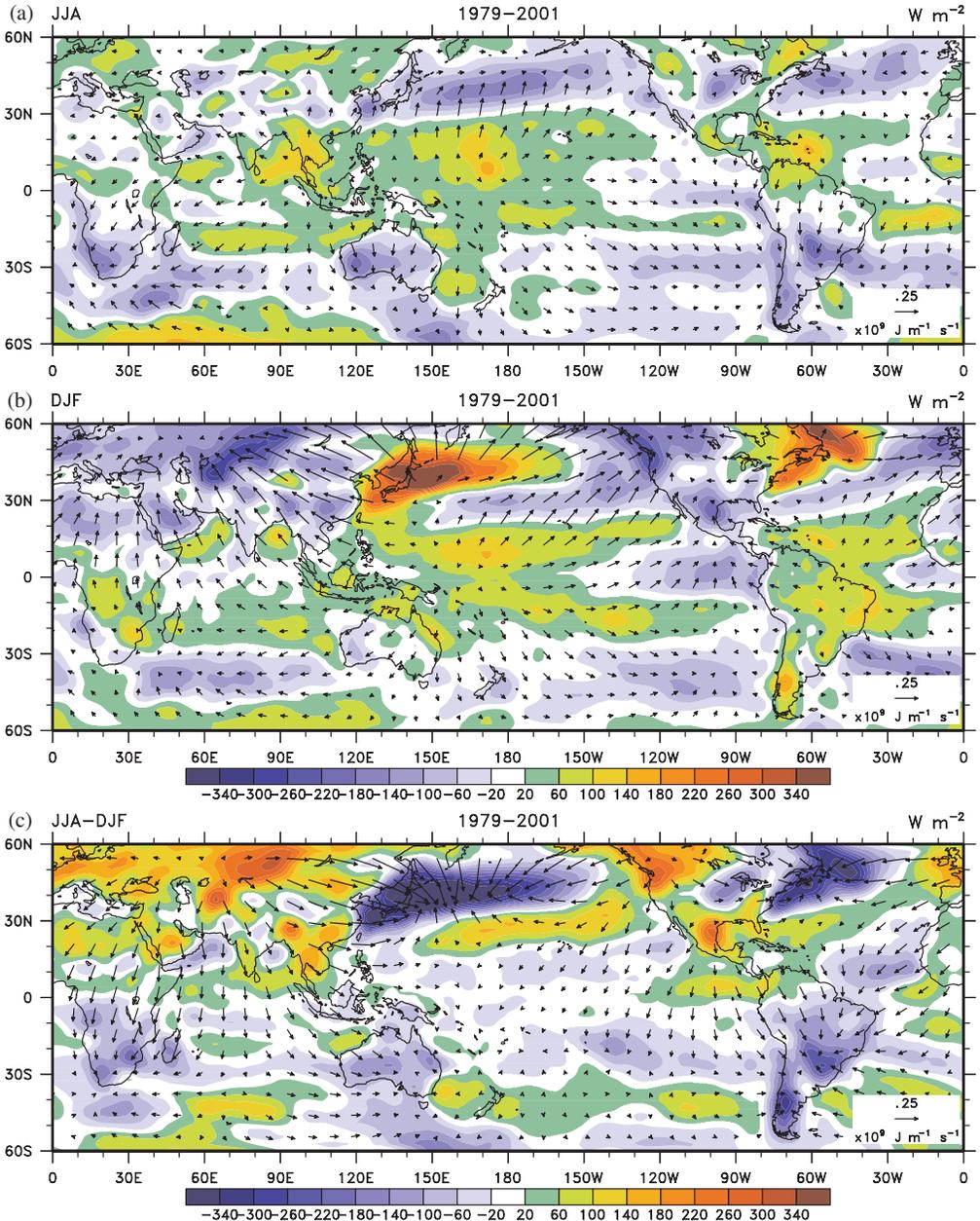


Figure 10. Divergence of, and divergent component of, vectors of the vertically-integrated total atmospheric energy transports for 1979–2001 averaged over: (a) June, July and August (JJA), (b) December, January and February (DJF), and (c) their difference JJA–DJF for the component from the quasi-stationary flow (monthly means and longer). The contour interval is 40 W m^{-2} and a reference vector is shown in each lower right-hand corner.

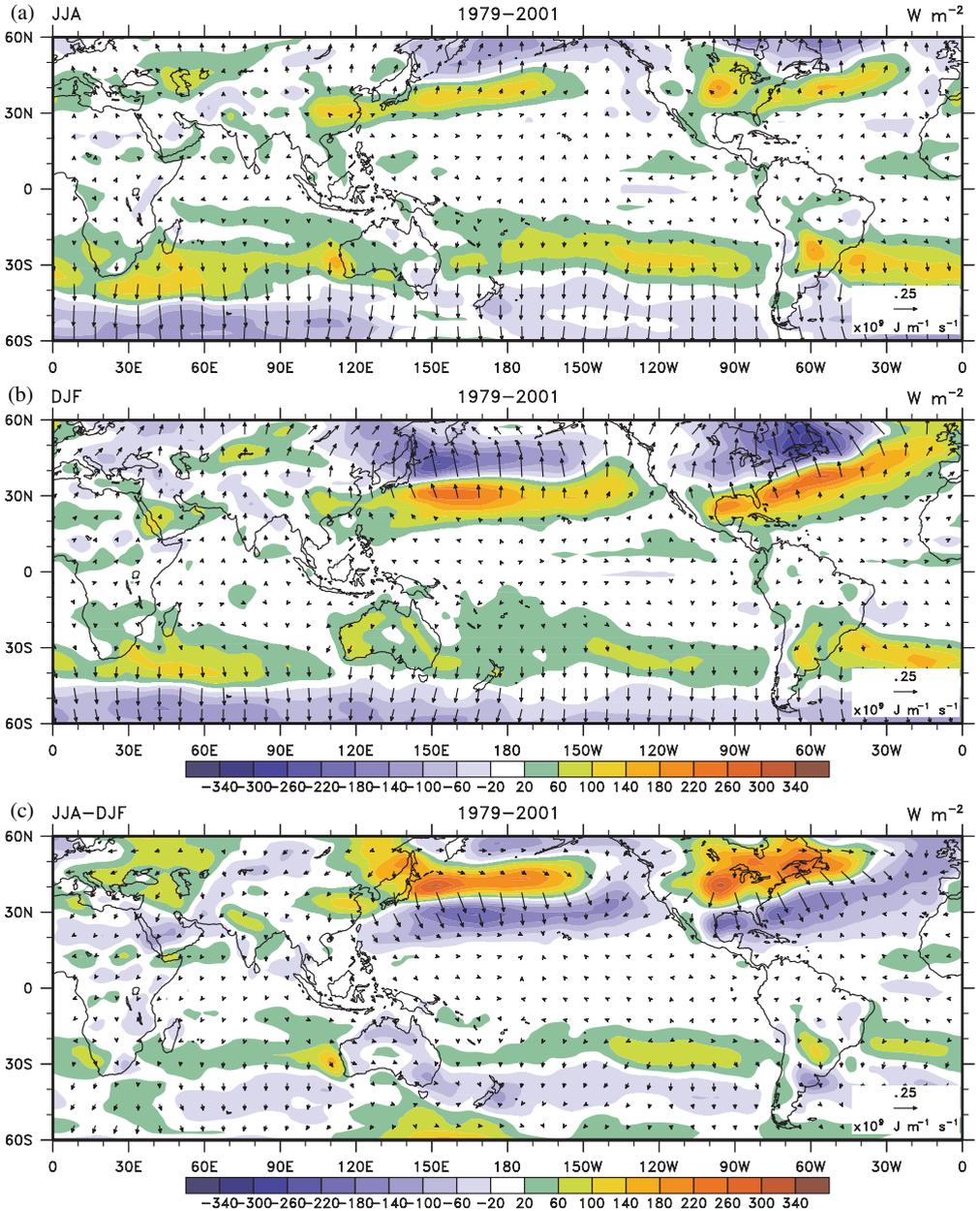


Figure 11. Divergence of, and divergent component of, vectors of the vertically-integrated total atmospheric energy transports for 1979-2001 averaged over: (a) June, July and August (JJA), (b) December, January and February (DJF), and (c) their difference JJA-DJF for the transient flow contributions. The contour interval is $40 W m^{-2}$ and a reference vector is shown in each lower right-hand corner.

energy is being transported to cooler regions (Fig. 10). Therefore, there also has to be a heat sink (cooling) in the downward branch of the circulation to balance the warming by subsidence. The classic examples are the monsoon circulations and the Hadley circulation (Webster 1994; Trenberth and Stepaniak 2003a). At low levels, the moisture in the atmosphere is transported towards areas where the air is forced to rise and hence cool, resulting in strong latent heating that drives the upward branch of the overturning cells and generates the monsoon rains or the ITCZ. In the subtropics, the downward branch of the circulation suppresses clouds and is very dry as the 'freeze-dried' air from aloft slowly descends, and the air warms as it descends and is compressed. Conventional wisdom suggests that radiative cooling to space, enabled because there is not much interference from clouds and water vapour greenhouse effects, compensates for this adiabatic warming. This is partly the case, but once again the relatively clear skies are a consequence of the circulation and thus are a feedback and not a fundamental cause (Hoskins 1996). This is mostly what happens over the Sahara (see especially DJF in Fig. 10).

However, another key part of the cooling in the subtropics that drives the Hadley cell is the link to midlatitude weather systems (Fig. 11). The latter transport both sensible and latent heat polewards in the cyclones and anticyclones and their associated cold and warm fronts, which arise from the equator-to-pole temperature differences and distribution of heating. Hence they act to cool the subtropics, while warming the higher latitudes. The seamless nature of the total transports, therefore, indicates a fundamental link between the Hadley cell transports and those of the baroclinic midlatitude weather systems and quasi-stationary waves (Trenberth and Stepaniak 2003a,b). Rodwell and Hoskins (1996) point out that theoretically, and in agreement with observations, the subtropics are where both vertical and horizontal advection are of comparable importance in balancing diabatic heating. Consequently it has been established that the major part of the cooling in the subtropics over the oceans comes from divergence of energy (cooling) by the baroclinic waves, which in turn are often organized into storm tracks (Fig. 11). The preferred storm tracks are slightly polewards of the main jet streams, which are a consequence of the convergence of momentum transports both by the overturning and the baroclinic eddies. Effectively, the baroclinic eddies carry energy to higher latitudes where it is eventually radiated to space as OLR. Consequently the cooling in the downward branch of the overturning circulations is linked to OLR diabatic forcing through radiation to space in all subtropical and extratropical regions.

Normally there are two Hadley cells in the atmosphere, one in each hemisphere. In the equinoctial seasons, both can be seen with roughly comparable strength (see Trenberth *et al.* 2000); but in the solstitial seasons the summer hemisphere Hadley cell is weak and the cell with subsidence into the winter subtropics becomes dominant. At the TOA, the net radiation (Fig. 4) provides the main reason why this is so, and the primary change is in ASR not OLR. However, both the ASR and the OLR have major contributions at the earth's surface, and to properly interpret what is happening within the atmosphere it is also necessary to examine the surface energy budget and the key role of evaporation in providing latent energy, which in turn involves the oceans and their uptake and release of stored heat (Trenberth and Stepaniak 2003a,b).

7. FUTURE CHALLENGES

The pattern of net radiation is largely externally imposed by sun–earth geometry; other influences, such as the surface albedo and cloud effects, are of secondary

importance. Hence the global heat budget provides a strong constraint on the climate system (Stone 1978; Trenberth and Stepaniak 2003a) and is a primary driver of the atmospheric circulation, mandating a poleward heat transport by the fluid components of the climate system: the atmosphere and ocean.

In the above we have presented a partial view of energetics of the climate system with emphasis on the role of energy storage in the oceans, and transports both in the oceans and especially in the atmosphere. The many different forms of energy in the atmosphere are associated with the rich meteorological behaviour of dry and moist dynamics that, nevertheless, are necessarily constrained to act in systematic ways to satisfy climate constraints: balancing the mass, moisture and energy budgets have been emphasized here.

The seamless character of poleward energy transports by the atmosphere, involving several different mechanisms, means therefore that there are fundamental links between them that have not been adequately appreciated. We have suggested that the cooling by transient eddies in the subtropics is a fundamental driver of the observed Hadley circulation, and realizes the seamless transport from tropics to extratropics on perhaps monthly and certainly seasonal time-scales. Atmospheric adjustment times for quasi-stationary waves and their associated storm tracks are of order one to two weeks (e.g. see Hoskins and Karoly 1981) making monthly fields somewhat noisy. Sea surface temperature patterns over the tropical oceans are important as they determine where the upward branch of the Hadley circulation is located. The relatively clear skies in the subtropics further provide for ample absorption of solar radiation at the surface where it feeds strong evaporation; this exceeds precipitation, and supplies the equatorward flow of latent energy into the upward branch of the Hadley circulation as well as the poleward transports into midlatitude storm tracks. The evaporation is sufficiently strong for it also to be compensated by a subsurface ocean heat transport that in turn is driven by the Hadley circulation surface winds (Held 2001).

Humans are interfering with the global flows of energy by changing the atmospheric composition and creating 'global warming'. The current radiative imbalance at the TOA is 0.5 to 1 W m⁻² owing to increases of greenhouse gases (Hansen *et al.* 2002). Some heat melts glaciers and ice, contributing to a rise in sea-level (Levitus *et al.* 2001). However, the main candidate for a heat sink is the oceans, leading to thermal expansion and further sea-level rise. Levitus *et al.* (2000) have estimated that the heat content of the oceans has increased on average by about 0.3 W m⁻² over the past few decades, and this accounts for most of the observed sea level rise of about 1.8 mm year⁻¹, which appears to have accelerated in the 1990s when accurate global measurements of sea level from the ocean TOPography EXperiment (TOPEX/POSEIDON) and JASON altimetry became available. Recent estimates of sea level rise are 3.1 mm year⁻¹ for 1993–2002, of which 0.3 mm year⁻¹ is from isostatic rebound, and with the suggestion that most of this is thermosteric (Cabanes *et al.* 2001; Cazenave and Nerem 2004). Other estimates place only 1.6 mm year⁻¹ of this total as being the contribution from thermosteric rise, corresponding to 0.86 ± 0.12 W m⁻² into the ocean (Willis *et al.* 2004), or about 0.6 W m⁻² globally. Pinning down these values both globally and regionally are key challenges for climate change science and the climate observing system.

Hence a major challenge in climate is to better determine the heat budget at the surface of the earth on a continuing basis. Provision needs to be made for changes in heat storage of oceans and associated changes in sea surface temperatures, glacier and ice sheet melt, changes over land in soil moisture and vegetation, and associated changes in atmospheric circulation, some aspects of which should be predictable on decadal time-scales. Several models, given the global sea surface temperatures, can

now simulate major changes such as: the sub-Saharan African drought beginning in the 1960s (e.g. Giannini *et al.* 2003); the 'Dust Bowl' era in North America in the 1930s (Schubert *et al.* 2004); and changes in the North Atlantic Oscillation (Hurrell *et al.* 2004). Coupled models cannot yet predict these evolutions, although there is hope that they will improve. In any case, because the climate is changing systematically from human influences, models should show some predictive skill simply based on the current state, although the latter is not yet sufficiently well known to allow full attribution of why the current state is the way it is. Such assessments require full consideration of changes in heat storage in the various reservoirs, changes in forcings, clouds and aerosols, fluxes among different climate system components, and transports by the atmosphere and oceans, all of which can be integrated into a climate system model framework. The synthesis of all components and fluxes, along with their uncertainties, provides for a holistic view that should sharpen knowledge of the climate system and highlight the outstanding issues. For instance, it would enable weaknesses to be reduced by combining components whose fluxes are poorly known. Such an information system would be extremely valuable.

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