Simulation of Present-Day and Twenty-First-Century Energy Budgets of the Southern Oceans

KEVIN E. TRENBERTH AND JOHN T. FASULLO

National Center for Atmospheric Research,* Boulder, Colorado

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ABSTRACT

The energy budget of the modern-day Southern Hemisphere is poorly simulated in both state-of-the-art reanalyses and coupled global climate models. The ocean-dominated Southern Hemisphere has low surface reflectivity and therefore its albedo is particularly sensitive to cloud cover. In modern-day climates, mainly because of systematic deficiencies in cloud and albedo at mid- and high latitudes, too much solar radiation enters the ocean. Along with too little radiation absorbed at lower latitudes because of clouds that are too bright, unrealistically weak poleward transports of energy by both the ocean and atmosphere are generally simulated in the Southern Hemisphere. This implies too little baroclinic eddy development and deficient activity in storm tracks. However, projections into the future by coupled climate models indicate that the Southern Ocean features a robust and unique increase in albedo, related to clouds, in association with an intensification and poleward shift in storm tracks that is not observed at any other latitude. Such an increase in cloud may be untenable in nature, as it is likely precluded by the present-day ubiquitous cloud cover that models fail to capture. There is also a remarkably strong relationship between the projected changes in clouds and the simulated current-day cloud errors. The model equilibrium climate sensitivity is also significantly negatively correlated with the Southern Hemisphere energy errors, and only the more sensitive models are in the range of observations. As a result, questions loom large about how the Southern Hemisphere will actually change as global warming progresses, and a better simulation of the modern-day climate is an essential first step.

1. Introduction

Major recent advances in understanding the energy budget have been provided by satellite data and globally gridded reanalyses. A synthesis of the top-of-atmosphere (TOA) global, land, and ocean domain energy budgets and their annual cycle (Fasullo and Trenberth 2008a, henceforth FT08), meridional structure (Fasullo and Trenberth 2008b), and the role of the ocean (Trenberth and Fasullo 2008) combined with surface constraints has provided a new set of observational information on the global energy budget for the 2000–05 period (Trenberth et al. 2009). This paper therefore uses these observations to evaluate models, starting with state-of-the-art atmospheric models used for reanalysis and then progressing to climate models. The focus is on the energy balance and its projected changes over the southern oceans in a global context. A reason for this focus is the unusual nature of the model projected changes over the southern oceans in the twenty-first century whereby absorbed solar radiation (ASR) increases at all latitudes except between about 50° and 60°S, and this turns out to be robust across models, whether run at low or high resolution. As a result there is a decrease in net radiation into the ocean in this band. Furthermore, the current energy budgets of the southern oceans are poorly simulated in all models examined, including those used for reanalyses, and the unique trends in this region are probably only possible because of the nature of these errors and thus are unlikely to be realized. We therefore describe and critique model performance for the current climate and projections of energy-related quantities into the future.

Atmospheric reanalyses from the European Centre for Medium-Range Weather Forecasts (ECMWF) known
as the 40-yr ECMWF Re-Analysis (ERA-40) (Trenberth and Smith 2009) and the Japanese 25-yr Reanalysis (JRA-25) (Trenberth and Smith 2008) have been evaluated with a focus on energy budgets using observational constraints to complement earlier evaluations of the National Centers for Environmental Prediction—National Center for Atmospheric Research (NCEP–NCAR) reanalyses (NRA) and other flux-based products (Trenberth et al. 2001). We further extend the evaluations to climate model simulations of the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4). We assemble these results for the 1990s, which is the last decade available from climate model simulations of the past. These simulations play important roles both in interpreting climate variability in the recent past and present, and in projecting the range of anticipated climate change under various forcing scenarios associated with the evolving planetary energy imbalance. The latter arises mainly from the continuing increases in greenhouse gases, including carbon dioxide, from human activities, as well as changes in aerosols. The simulations thus contribute not only to scientific understanding, but also to future planning, adaptation, and policy guidance.

The Third Coupled Model Intercomparison Project (CMIP3; Meehl et al. 2007a) compiled the combined contributions of 18 modeling groups running 24 coupled climate models. The project represents an unprecedented coordinated effort to provide the broader scientific community access to a full range of simulations of past and present climate under an approximately uniform set of forcing scenarios. The model data were made available through the U.S. Department of Energy’s Program for Climate Model Diagnosis and Intercomparison (PCMDI). Understanding the feedbacks that shape the simulated climates of the CMIP3 database remains an important science objective (e.g., Randall et al. 2007), and comprehensive analysis of the energy budget is fundamental to this pursuit. We report here on the A1B moderate emissions scenario, but we have also examined the A2 scenario, with similar results.

The energy budget is a fundamental determinant of the mean state of the climate, as the flow of energy through the climate system is intrinsically tied to its spatial and temporal structure, and the variance of temperature, humidity, winds, rainfall, and clouds. The systematic flow of energy required by the basic sun–earth geometry involves all components of the climate system, including the atmosphere, ocean (at all depths), cryosphere, and the land surface and its hydrology. Variability in any one component affects others, but often with different response times. Similarly, biases in any one component can undermine the simulation of the entire climate system, and an improved understanding of the energy budget in any single aspect can improve the simulation of all of the system’s components. Hence the interaction between the energy budget and feedbacks in the climate system will be considered here.

In the Southern Hemisphere (SH), several previous studies have highlighted the large and important changes over the southern oceans in association with the southern annular mode (SAM) (e.g., Gillett and Thompson 2003; Trenberth et al. 2007; Cai and Cowan 2007; Sen Gupta et al. 2009). Gillett et al. (2003, 2005) formally detected and attributed the changes to human influences and ozone depletion in particular. Thus, in the present analysis, it is expected that results may be sensitive to the nature of forcings used in any given simulation (in particular whether ozone depletion is included) and we focus on models with more realistic ozone depletion.

In section 2 we outline the models and observations and methods used. Section 3 describes the main results relevant to the current observed climate for reanalyses and the climate models. Section 4 then delves into changes in energy-related quantities since 1950, relative to a base of 1900–50, and extending through the twenty-first century. Section 5 discusses the results and the conclusions are given in section 6. The focus here is on aspects of the simulated budgets that are resilient both to the effects of problems in the models and archives, as discussed in section 2.

2. Methods and data

Evaluation of the model energy and water budgets is nontrivial and is hampered by several issues. These include the adequacy of observations for validation, the natural variability in the models, and, for coupled models, the observed climate (such as El Niños) not being synchronized, inconsistent forcings used in the models (especially for aerosols and ozone), and errors in the model archives and the lack of closure in the models themselves.

A foremost issue is the lack of adequate absolute accuracy in continuous global energy observations over several years. Only in recent decades have global-scale observations been possible through satellite observations, and much of the data record is plagued by biases that change over time as well as temporal discontinuities, often associated with changes in satellites and instruments, that preclude meaningful trend analyses (Trenberth et al. 2007). Moreover, present-day observing systems are unable to resolve the global energy imbalance relevant to climate change (e.g., FT08; Loeb et al. 2009), which is small (order 1 W m$^{-2}$) particularly in relation to the large spatial and temporal variance in the individual fluxes. However, this paper builds upon
the comprehensive assessment we have made of these aspects (FT08; Fasullo and Trenberth 2008b; Trenberth and Fasullo 2008; Trenberth et al. 2009).

As the model simulations are coupled, the phase of internal modes of variability—including, for example, El Niño–Southern Oscillation (ENSO)—is not generally in phase with that of nature. Thus divergence of the simulated energy budgets from observations related to such internal variability can become conflated with model bias, and the dissection of the two is not trivial. In addition, the nature of external forcings (e.g., volcanic aerosols) employed by the various modeling groups in CMIP3 is generally not uniform and represents a simplification of the actual forcings in nature. Some models (8) do not include forcing associated with observed ozone depletion (see Cai and Cowan 2007; Son et al. 2008). Moreover, the treatment of the indirect effects associated with some forcings, such as aerosols, is nonuniform among modeling centers and its incompleteness is a reflection of limitations in our understandings of these interactions. Some models include flux adjustments, which invoke spurious sources and sinks of energy to achieve a more reasonable simulated climate. Thus spatial and temporal variability in the energy budget can simply result from differences in the modeling framework.

The framework for the quantities examined in this paper is given in Fig. 1. The TOA net radiation down (\(R_T\)) is examined, along with its two components, the ASR and outgoing longwave radiation (OLR); \(R_T = \text{ASR} - \text{OLR}\). The vertically integrated energy transports in the atmosphere and ocean are also used. Globally, by definition, the divergence of the energy transports is zero for the atmosphere and the global ocean. The surface flux \(F_s\) is directed out of the ocean by convention.

### a. Observations and reanalyses

Present-day energy budget estimates following FT08 and Fasullo and Trenberth (2008b) are used as an observational baseline. At the TOA, adjusted radiances from the Clouds and the Earth’s Radiant Energy System (CERES) surface averages (SRB AVG) Edition 2D Rev1 are used but updated to cover March 2000 to October 2005.

In the atmosphere, mass-corrected budgets from reanalyses from the NRA (Kalnay et al. 1996), ERA-40 (Uppala et al. 2005), and JRA (Onogi et al. 2007; Trenberth and Smith 2008) are used as they contain the most comprehensive estimates of global atmospheric temperature and moisture fields. Vertically integrated quantities for atmospheric energy components, the tendencies, transports, and divergences have been computed and regridded to a T63 (192 × 96) grid. Biases in the energy budget at TOA result principally from the cumulative effect of errors in temperature, moisture, cloud, and aerosol fields and also from errors in the model radiation codes and other model components that may fail to conserve energy. The spread among the reanalyses in these fields contributes to differences in the vertically integrated budgets that are considerably larger than the uncertainty in CERES TOA fluxes (given by Trenberth and Fasullo, 2008). As a result, the TOA fluxes serve as an important constraint for evaluating the fidelity of the reanalysis fields. Of interest is whether the biases found in the reanalyses are systematic, therefore suggesting that they result from gaps in our understanding of data assimilation or modeling. This hypothesis is explored in greater detail in subsequent sections.

### b. CMIP3 fields

Fields from the CMIP3 archive (Meehl et al. 2007a) are based on the climate of the twentieth century (20c3m) and twenty-first century [Special Report on Emissions Scenarios (SRES)-A1b] simulations. CMIP3 fields are regridded to the T63 grid to facilitate comparison. One member (run1) of each model’s simulation ensemble (which differ in size) is chosen to compute the annual mean and mean annual cycle from 1990 to 1999 for the...
evaluation against observations. Because the TOA observations are for post-2000 and none exist for the 1990s while the 20c3m runs stop in 1999, some small regional discrepancies arise from the difference in time periods. Given the evaluation of model performance, projections of the twenty-first century are then assessed.

A screening of simulations is employed to narrow the models considered from the full CMIP3 archive. Although many diagnostics were computed with all models, it was possible to use only a subset for complete energy balances. Table 1 summarizes the model names, acronyms, and horizontal and vertical resolutions of models used in the current analysis. The model screening imposed excludes those simulations with energy budgets known to be unphysical, such as those with energy flux corrections, and those that fail to include the effects of ozone, which exerts a key forcing in the SH. For some terms in the energy budget, the archive is incomplete and does not allow for computation of all fields. Projections of the twenty-first century from the same set of models from 2000 to 2099 are examined and gaps in the early twenty-first century are interpolated linearly, where such gaps exist, to provide continuity from 1900 through 2100.

In the course of evaluating the model energy budgets, it became apparent that not all of the archive values are necessarily correct. For instance, in the case of the Community Climate System Model, version 3 (CCSM3) from NCAR, there were errors in the reported values archived that arose from confusion over top-of-model versus TOA fluxes. In fact there are layers of the atmosphere above the top of model that may absorb radiation and can radiate downward and also alter the OLR value that would be seen by satellite. The difference can be of the order of 1 W m$^{-2}$. It is therefore not possible to retrieve the correct OLR from the CCSM3 CMIP archive. Similarly, errors in the derivation of surface latent heat flux were found and this precluded a derivation of the net surface energy flux. It is not known the extent to which these kinds of problems exist elsewhere. Notwithstanding similar errors, violations of energy are also known to occur within the NCAR model, for example in relation to the treatment of glacial calving, resulting in a spurious energy sink that, when averaged globally, is of order 0.3 W m$^{-2}$. It is expected that other models exhibit violations of energy conservation similar to those in the CCSM3. Yet another problem is that some small terms are not available and some are not accounted for. In a model, is the sensible heat of precipitation accounted for when it hits the ground or goes into the ocean? Does the melting of snow to produce rain fully account for the latent heat released? Is water conserved? Is the heat associated with friction accounted for? Is the latent heat of snow landing on ocean accounted for? For a huge reservoir of heat

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such as the ocean or ice sheets (that do not grow or decay anyway), heat and water may not be fully accounted for and sea level does not vary in any model. In all of these cases, there is effectively a “flux adjustment” that is built into the tuning of many models and the accounting of energy is inexact.

Yet another problem uncovered in the CCSM3 is that, as for all models, a tuning is performed to achieve a balanced TOA radiation. Normally this would be done for the period 1870–80 or a similar period when the global mean TOA net radiation is very small, so that the 1990s would be appropriately out of balance because of anthropogenic forcing. However, for CCSM3, the tuning was performed for the 1990s, creating an unrealistic zero imbalance for that decade but a negative imbalance for the late 1800s. Presumably the changes over time are not greatly affected, but there is an offset that needs to be accounted for.

Hence, to conduct a meaningful comparison and allow for outliers, rather than take the absolute energy budget values, because of likely errors in the archive, we remove the bias and examine differences relative to the base period 1900–50.

3. The mean global energy budget depicted by models

a. Reanalyses

Atmospheric models run in so-called Atmospheric Model Intercomparison Project (AMIP) mode, where sea surface temperatures (SSTs) are specified as observed, are not required to conserve energy even if the same atmospheric model has been designed to do so in a coupled model (e.g., Trenberth et al. 2009). This also applies to models used in reanalysis that are constrained by assimilated observations. Moreover, biases can be quite large as model fields are nudged toward observations at each analysis time with an increment, and thus physical constraints are not satisfied. As observed energy fluxes are not explicitly assimilated by reanalyses and energy terms are a product of the reanalysis framework, comparison with observed energy fluxes provides an important diagnostic of the reanalysis framework, including both its assimilation scheme and forecast model and their collective biases.

To illustrate the nature of energy problems in reanalysis models, JRA TOA fields are compared against adjusted CERES retrievals for the 1990s (Fig. 2). As found in Trenberth and Smith (2008), in JRA net downward TOA flux ($R_T$) is deficient across the tropics, particularly over ocean. At higher latitudes, and particularly in the southern oceans, $R_T$ is excessive. However, on a global basis $R_T$ biases are dominated by low latitudes, with significant contributions from both ASR and OLR (Trenberth et al. 2009). Over ocean throughout the low latitudes, ASR is systematically deficient because of excessive albedo and biases are likely to be due to excessive shortwave cloud forcing. At high latitudes, ASR is generally too large, particularly for the southern oceans, and this compensates somewhat for low-latitude biases in the global mean, although the global ASR bias is still strongly negative (Trenberth et al. 2009). Over land ASR is generally too large, with exceptions in Australia and some mountainous domains. Biases in OLR are mostly positive and are particularly pronounced over low-latitude land areas, suggestive of deficient cloud, and over the intertropical convergence zone (ITCZ) and warm tropical oceans. There is therefore a substantial (19 W m$^{-2}$) bias in the global mean.

The principal biases in JRA have also been identified in other reanalyses, in NRA (Trenberth et al. 2001) and ERA-40 (Trenberth and Smith 2009), and are summarized in Trenberth et al. (2009). Walsh et al. (2009) document problems in radiative fluxes associated with
cloud errors in the Arctic. For the 1990s, the zonal mean biases in ASR (Fig. 3) are strongest in the summer as they scale with solar insolation. For the reanalyses, we can also check these biases for the same period as the CERES data exist, and the results are robust. Negative biases in NRA are large across much of the oceans because of an incorrect surface albedo that is much too high (see Trenberth et al. 2009). This masks the effects of cloud biases in the NRA. In ERA-40 the negative bias in SH low latitudes persists throughout the year and exceeds 30 W m$^{-2}$; analogous biases are present in Northern Hemisphere (NH) oceans. In high southern latitudes the excess of ASR owing to deficient clouds is endemic to all reanalyses and hence SH problems are pronounced. It is suggestive therefore that the biases reflect fundamental shortcomings in the ability to simulate key aspects of the SH energy budget.

b. CMIP3

Unlike the reanalysis models, coupled models operate in a framework in which energy and moisture is conserved to order 1 W m$^{-2}$ and the climate state is not nudged by observations. Thus biases are often controlled by mitigating feedbacks that lessen their overall magnitude from those produced in reanalyses. Despite this, the dominant systematic mean CMIP3 biases (Fig. 4) closely resemble those of the JRA. The annual mean departure of the mean CMIP3 model $R_T$ from observations features negative biases over much of the tropical oceans and positive biases poleward of 35° and, particularly, over the southern oceans and off the west coasts of the Americas.

The corresponding cloud amount relative to International Satellite Cloud Climatology Project (ISCCP) D2 retrievals (Rossow and Schiffer 1999) (Fig. 5) shows a low bias over most areas. ISCCP data combine geostationary satellite observations, which have the advantage of high-frequency sampling, with polar orbiter observations. Over the Indian Ocean sector, where the geostationary satellite observations are missing, there is a spurious wedge in the ISCCP data that also shows up in the differences because of the lack of data from that region’s geostationary satellite. Observed cloud amount depends greatly on the methodologies and sensors used and therefore differences are large between ISCCP and High Resolution Infrared Radiation Sounder (HIRS) (Wylie et al. 2005), while new observations from CloudSat and Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) (Chepfer et al. 2008) reveal other sources of bias. Perhaps more importantly, modeled and observed clouds are not strict analogs. In instances, models report the existence of clouds with minute liquid
and ice water paths that are not radiatively important while overall models simulate fewer clouds particularly in the midlatitude middle troposphere, than are observed. There are also several issues that preclude direct comparison of observed cloud amount and modeled radiative fluxes (Zhang et al. 2005). For instances, cloud amount averages all times of day and year equally, whereas shortwave radiation quantities effectively weight cloud effects by the amount of incoming radiation, which is stronger by day and in summer. Nonetheless, the ISCCP and ground-based observations are sufficiently similar compared with model results to reveal substantial biases in models (e.g., Dai and Trenberth 2004). However, cloud information is used only to help interpret the radiative diagnostics and the comparison in Fig. 5 provides qualitative insights into the energy budget.

Overall the greatest biases in clouds occur over midlatitude ocean domains, while over land the zonal mean bias is less. The most marked biases are evident over the Southern Ocean where zonal mean deficiencies exceed 20% in most models. Important regional biases are also evident, with deficient cloud in the stratus deck regions of the eastern ocean basins and excess cloud in the sub-sidence of subtropical ocean domains in the equatorial Pacific and Atlantic Oceans. A well-known error common to coupled climate models is the tendency to have a double-ITCZ structure for annual mean conditions (Lin 2007), resulting from a spurious migration of the ocean ITCZs into the SH. This in turn produces too much cloud in the southern tropics over the Pacific and Atlantic Oceans and too little in the normal ITCZ location, and these biases are reflected in $R_T$. Too much cloud over Indonesia can arise from problems with El Niño, which is also common in models.

Nonetheless, Fig. 5 is suggestive that cloud biases are the root basis of many of the energy budget shortcomings. The $R_T$ bias suggests a widespread and systematic negative bias in net downward radiation from about $35^\circ$S to $35^\circ$N over most of the oceans (Fig. 4) that is less evident in the cloud amount bias. In this region, it appears as though the bias arises more from cloud optical thickness and cloud-top heights that are too large in regions of convection and large-scale ascent (e.g., Webb et al. 2001; Norris and Weaver 2001; Lin and Zhang 2004; Field et al. 2008). This is also true for the ERA-40 reanalysis model (Trenberth and Smith 2009; Trenberth et al. 2009). The errors in the subtropics were commented on by Webb et al. (2001) for three models and they found that the source of the errors included excessive high cloud, cloud amount, and optical thickness of boundary cloud. Bony and Dufresne (2005) suggest that the simulation of the sensitivity of marine boundary layer clouds to changing environmental conditions constitutes the main source of uncertainty in tropical cloud feedbacks simulated by general circulation models. Bony et al. (2004; 2006) and Dufresne and Bony (2008) conclude that the main source of model differences remains cloud feedback, and especially from low-level clouds and in regions of subsidence.

Other aspects of clouds are more clearly deciphered by separately examining the ASR and OLR fields, for, if the amount is correct but its vertical distribution is skewed, ASR and OLR can experience disproportionate effects (e.g., see Webb et al. 2001) and indeed this possibility is strongly suggested in Fig. 6. In the subtropics
and eastern Indian Ocean, biases in ASR and OLR suggest too much cloud and associated excesses in albedo with ASR far less than observed. Biases between 30° and 60°S suggest the opposite effect, as evident in JRA (Fig. 2), while biases over the Saharan region may relate to absence of dust or the appropriate surface albedo in models.

The OLR fields (Fig. 6b) suggest compensation of biases in many regions, such as the eastern Indian Ocean, western Pacific warm pool, central Pacific Ocean, and over tropical land as biases in OLR mirror those in ASR and thus cancel in RT. Over the southern oceans, however, because of the cooler surface temperatures and reduced role of cloud in determining OLR, biases in RT largely mirror those in ASR.

c. The annual cycle

The zonal mean annual cycle for the TOA radiation biases from Figs. 4 and 6 are present most of the year although with some modulation (Fig. 7). ASR biases dominate the bias in RT generally, although at high latitudes OLR biases tend to augment those in RT, while at low latitudes they tend to lessen them. Biggest deviations are poleward of about 50°S latitude where zonal mean excesses of ASR and RT exceed 15 W m⁻² in the summer half year, especially from October through March. Similar but weaker biases extend from April to September, when incident solar flux at the SH TOA is much reduced from summer. As well as cloud, sea ice errors could contribute near the fringe of Antarctica, at about 65°S, where largest biases in ASR occur, and large negative biases in low latitudes are also strongest in summer in each hemisphere. In the NH, there is a negative bias from June to September north of 60°N and a small positive bias for the rest of the year, particularly April through June, suggesting problems with Arctic sea ice and cloud. In the longwave, the greatest errors are related to seasonality in the ITCZ from January through June (related to the spurious migration of the ITCZ across the equator); however, systematic biases are also evident throughout the winter season for the respective hemispheres, and particularly the NH.

Averaged globally, the annual cycle of the energy budget (Fig. 7d) reveals excessive absorption during boreal winter, coinciding with excessive SH ASR, but negative values in March–April and July–August, stemming mostly from lower-latitude biases.

d. Meridional transports

The meridional transport of energy by the atmosphere and ocean from the equator to pole represents the dynamic compensation for the gradient in ASR, driven largely by the sun–earth geometry. As the multiyear tendency term in storage is small, biased meridional transports are a direct consequence of errors in the gradient in RT. Observational uncertainty in total transport is substantially less than for atmospheric and oceanic components, which rely additionally on estimates of
the divergence of atmospheric energy transports $\mathbf{V} \cdot \mathbf{F}_A$ (Fig. 1). A thorough analysis of errors in the observations is provided in Trenberth and Fasullo (2008).

When the radiative imbalance at the TOA is converted to the required northward energy transport (Fasullo and Trenberth 2008b; Trenberth and Fasullo 2008) and compared with the CMIP3 model values, systematic biases in the total and both atmosphere and ocean are evident (Fig. 8). In the NH, peak total transport of slightly less than 6 PW is simulated reasonably well by most models, but in the SH more than 75% of the models are well outside the bounds of uncertainty and all models are deficient. The bias in total transport is the direct consequence of too much solar radiation entering the system in the SH and too little entering at low latitudes (Fig. 4). At low latitudes in the SH, where the ocean contributes most significantly to total transport, the simulated ocean transports are universally too weak, by about 0.4 PW between 5°S and 10°S, indicative of excessive net upward surface flux that is associated with deficient surface incident solar radiation (not shown) and deficient ASR (Fig. 6). At high southern latitudes, biases in surface ASR are moderated by latent heat and longwave fluxes, and errors in ASR do not project significantly onto the oceanic transport. Nonetheless, in the atmosphere, mean biases in transport of 0.9 PW are evident between 35° and 45°S.

The main poleward transport of energy in the SH occurs from transient baroclinic storms (Trenberth and Stepaniak 2003) that are active over the southern oceans, and consequently the deficit in poleward transport is a direct indicator of insufficient storminess in the SH in models, which no doubt contributes to well-known errors in the strength of the SH westerlies (e.g., Hurrell et al. 2006 for the CCSM). The deficit in transport is therefore also likely to be related to deficits in convection and cloud in present-day simulations (Fig. 5).

4. Projections for the twenty-first century

a. Changes in ASR and cloud

The changes in ASR relative to 1900–50 in the CMIP3 archive (Fig. 9) show that, uniquely among all zones, the Southern Ocean from about 48° to 62°S experiences a decrease in ASR after about 2000. The zonal mean
time section shows that the linear trend for the twenty-first century is a good depiction of the spatial structure over the southern oceans. In the twentieth century, however, ASR evolves rather differently as changes in aerosols play a key role in the A1B scenario. Increases in ASR in the twenty-first century occur in almost all models. Trends in other model energy quantities (RT and OLR) are given in Trenberth and Fasullo (2009) for a somewhat different set of models. Results are robust to model selection for all quantities and they do not seem to be very sensitive to whether ozone was treated realistically or whether volcanoes are included.

The unique zone in between these two regions occurs between 50° and 60°S over the open ocean. Presumably it could also occur in the NH but the large landmasses preclude that behavior. OLR tends to decrease in this zone but RT follows ASR and exhibits a slight decrease (Trenberth and Fasullo 2009) (see Fig. 12 presented later). The high-latitude SH, in particular the zone from 50° to 60°S and poleward of about 75°S, exists as the only region for which the shortwave feedback is negative in models. While increases in cloud amount over ocean in winter do not affect ASR greatly, the summer increase is significant, and to more fully account for how cloud alters ASR it is essential to examine the seasonal variations (section 4c).

An outstanding and remarkable result is the very strong resemblance between the current-day errors in cloud amount (Fig. 5), as simulated by the CMIP3 models, and the twenty-first-century trend (Fig. 10 lower panels). The resemblance extends to the positive values in the tropical eastern Pacific, the strong negative values in midlatitudes, and the positive values over both polar regions. The models therefore have a strong tendency to make the original errors bigger relative to today’s climate, which raises questions about the likelihood of the trends being realized. The 50°–60°S zone is one exception in that the trend is the opposite of the error.

b. Changes in sea level pressure and atmospheric circulation

Broad aspects of the cloud changes appear to have a dynamical basis stemming in large part from changes in the atmospheric circulation, as indicated by changes in sea level pressure (Fig. 11). Overall, there is an increase in the mass of the atmosphere because of increased water vapor, and this is reflected in a small increase in the global mean surface and sea level pressure fields. Higher pressures over the Indonesian region reveal a local increase in convection and cloud, such that OLR decreases, canceled to some extent by decreases in ASR, but with an overall increase in $R_T$ (Trenberth and Fasullo 2009). The Pacific trends are offset elsewhere in the tropics where clouds decrease and ASR, OLR, and $R_T$ all increase. Poleward of about 60° in both hemispheres, reductions in sea and land ice and snow cover reduce the planetary albedo (not shown) and lead to increases in ASR even as substantial increases in cloud cover (Fig. 10) offset the influence of the snow and ice albedo reduction. Part of this response stems from increased open water and higher SSTs and local evaporation, with increases in water vapor. Over Antarctica, however, warming increases OLR (Trenberth and Fasullo 2009) such that that the change in $R_T$ is negative.

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are part of the trend to more El Niño–like conditions. Elsewhere, the sea level pressure trends suggest higher values in the subtropics and especially midlatitudes, and with lower pressures at high latitudes. Accordingly, the gradient in sea level pressure increases substantially over the Southern Ocean, and this is associated with stronger westerlies and a poleward shift in storm-track activity (Yin 2005) (see discussion below).

c. Seasonality of the changes

The annual cycle of the changes that occur from 2090 to 2100 for \( R_T \), ASR, OLR, and cloud amount (Fig. 12) reveal strongest changes in ASR in the summer season, which is an obvious consequence of that being the time of maximum solar radiation. For the zonal means (Fig. 12), the changes persist year-round. Over the southern oceans, the exceptional zone from about 50° to 60°S is present in all months but is significant only from November to March, when it also extends to about 65°S. In this zone, the ASR decrease corresponds to an increase in cloud amount (Fig. 12). In the spring months, when sea ice is at a maximum and sea ice changes are large, the increase in ASR over Antarctica extends slightly north of 60°S, offsetting the decrease during summer south of 60°S.

d. Relationships with climate sensitivity

In evaluating the processes that both drive and moderate climate change, an understanding of feedbacks in the system, particularly as they relate to the energy budget, is crucial (Randall et al. 2007). The equilibrium climate sensitivity \( S \), defined as the equilibrium change in global surface temperature for a doubling of carbon dioxide, is a measure of the strength of the positive and negative feedbacks. The importance of the energy budget of the SH is underscored by the fact that a strong relationship exists between present-day simulations of the net radiation at TOA and simulated climate sensitivity \( S \) (Fig. 13) with a highly significant negative correlation of \( -0.73 \). The observed value of \( R_T \) in the SH is 1.4 W m\(^{-2}\) from FT08, but other observational values place it at \( -0.1 \) W m\(^{-2}\) (from Loeb et al. 2009). Accordingly, most models have a value much too high, which is related to the excess ASR and deficient OLR over the southern oceans. The models that fall within the observational uncertainty are at the high end of \( S \). Models that, in general, have a substantial net incoming flux in the SH, such as the NCAR Parallel Climate Model (PCM1) and both of the Goddard Institute for Space Studies (GISS) models, are less sensitive, generally, than other models. Models with the highest sensitivity generally have a small net SH flux, including the Model for Interdisciplinary Research on Climate (MIROC) medium- and high-resolution models. A notable exception is the Met Office (UKMO) Hadley Centre Global Environmental Model version 1 (HadGEM1) for which the net SH flux is relatively large (3 W m\(^{-2}\)) but its sensitivity is high (4.4).

5. Discussion

Observations from 1970 to 2000 reveal increases in westerly winds south of 50°S associated with the southern annular mode (Gillett and Thompson 2003; Trenberth et al. 2007; Cai and Cowan 2007) accompanied by increases in ocean heat content (Gille 2008). This trend is projected to continue into the twenty-first century. As shown by Yin (2005), the storm track, as indicated by transient eddy kinetic energy (EKE), is currently centered on about 45°S in southern summer but increases in magnitude and shifts poleward and upward, leading to an increase in activity from 45° to 60°S and decreases farther north in the latter part of the twenty-first century. The maximum increase in eddy activity is between 50° and 60°S in association with increased poleward temperature gradients and baroclinicity, and increased precipitation occurs between about 50° and 70°S (Yin 2005). In winter, the changes are similar but extend over
a broader latitudinal range. The change in sea level pressure (SLP) (Fig. 11) is consistent with this poleward shift in storm track, and changes in cloud accompany both the storm-track activity and the generally lower pressures over Antarctica.

Higher sea level pressures indicate more anticyclonic conditions and settled weather, while cloudier areas over the polar regions correspond to more cyclonic conditions, lower surface pressures, and enhanced storminess. However, the change in storm tracks and the westerlies shifts the cloud pattern equatorward relative to that suggested by sea level pressure alone. Accordingly, the change in cloud is consistent with dynamical changes in circulation.

The pervasive decrease in cloud cover outside of the polar regions in both hemispheres leads to an increase in ASR that dominates the planetary budget and is analyzed in detail in Trenberth and Fasullo (2009). ASR also increases in polar regions but for different reasons, as cloud amount tends to increase. All models simulate reductions in sea ice during the twenty-first century in both polar regions, and in the SH these reductions occur mainly south of 60°S. There is also a strong tendency for the models to simulate a corresponding increase in evaporation and cloud, as the warmer ice-free ocean allows more heat and moisture into the atmosphere. Observations suggest that this occurs notably in spring (e.g., Walsh et al. 2009). It results in an offset in the albedo change, which is not then as large as the sea ice loss alone might suggest, and there is an increase in surface radiation into the polar oceans.

However, in the SH, the increase in cloud extends farther north than the former sea ice domain and leads to a decrease in net surface radiation in the 50° to 60°S zone that is unique. It is closely associated with a decrease in surface pressure south of about 60°S and an increase in the strength of the westerlies from about 45° to 65°S that continues recent trends (Yin 2005). The projected changes in the jet stream are consistent with the southward shift in transient baroclinic activity that is responsible for some of the cloud changes; see also Cai and Cowan (2007) and Cai et al. (2010). The latter study shows that the biggest warming of the oceans takes place
from 35° to 50°S but stems from fluxes into the ocean south of 50°S, as there are related impacts on the currents in the southern oceans and changes in the ocean transports (see also Sen Gupta et al. 2009) that are related to the changes in storm tracks and associated surface winds.

There is reason to be skeptical about the reality of these simulated changes as the biases in models over the southern oceans are large and reduce poleward energy transports by the ocean and atmosphere relative to modern-day observations. These biases are endemic to atmospheric models and occur in both very high-resolution numerical weather prediction models, such as those used for reanalysis at ECMWF in ERA-40 and the Japan Meteorological Agency (JMA) in JRA-25 (Trenberth and Smith 2008, 2009) and in reanalyses in the Arctic (Walsh et al. 2009) and lower-resolution climate models. The main sources of discrepancy between climate models and satellite observations of TOA radiation stem from problems in simulating cloud distribution, cloud type, cloud-top height, and radiative properties. The general tendency is for a deficit in ASR in the tropical and subtropical oceans that is more related to cloud type and optical properties but leads to insufficient energy that should be available for transport to higher latitudes. Insufficient cloud over the southern oceans is endemic. In this very cloudy region, models tend to project increases in cloud, yet it is physically impossible to increase cloud above 100%. In the real world, cloud cover of order 90% does not allow much scope for increases to occur, while model simulated values of 80% or less do. It is probably only the low bias in current simulations that enables a positive trend to occur, casting doubt on the projections. The combination of too little absorption in the tropics and too much over the southern oceans indeed reduces baroclinicity associated with the storm tracks and thus the southward energy transports by the atmosphere and the ocean in models compared with observations.

The relationship between model climate sensitivity and net SH radiation in Fig. 13 is highly suggestive of a link with the SH energy budget, but it also raises many questions. It is likely that the relationship (Fig. 13) is merely symptomatic of other things going on in the model and is not the explanation of either the sensitivity or the errors in the SH, but it is strongly suggestive that the larger the errors in the SH, the smaller the sensitivity, and this in turn relates to the negative biases in cloud amount. However, what are the mechanisms that drive sensitivity and how is present-day SH climate tied to these mechanisms? Is the relationship in Fig. 13 even indicative of the processes that determine $S$, or is its relationship ancillary? The scope of these questions extends beyond the current investigation but will be addressed in subsequent work.

6. Conclusions

Following earlier observational analysis of the energy budget, we have evaluated models as to their performance. High-resolution state-of-the-art atmospheric models used in atmospheric reanalysis reveal large biases in $R_T$, ASR, and OLR that are linked to clouds. For these model-based analyses, involving assimilation of observations, the evaluations can be done month by month. CMIP3 climate models used for AR4 projections have coarser-resolution atmospheric models but exhibit many of the same biases. In the NH, the biases are complicated by the land–ocean distribution and the role of aerosols, and hence the picture is simpler in the SH.

In the SH a negative bias in ASR dominates the tropical oceans because of clouds that are too bright, and for the reanalyses OLR positive biases tend to add to the error, giving a deficit in the $R_T$ received. For the climate models, OLR biases compensate to some degree but the deficit in $R_T$ follows the error in ASR. Over the southern oceans, however, cloud amount is underestimated and too much ASR occurs in all models. The result is that there are significant underestimates in poleward energy transport by both the atmosphere and ocean in the SH, with the ocean transports too low in lower latitudes and atmospheric transport too low in midlatitudes. This affects the present-day simulated strength of the westerlies in the SH and storm activity. Ocean model deficiencies, especially through lack of eddy-resolving resolution, may also contribute to the problems (Cai et al. 2010).

Projections for the twenty-first century of the CMIP3 models show increases in ASR at all latitudes except about 50°–60°S. In polar regions this is caused by decreases in albedo associated with decreases in snow and ice but compensated to some extent by increases in cloud amount. In low and midlatitudes, albedo decreases because of decreases in cloud amount. Only in the 50°–60°S zone does cloud amount increase in summer and result in a decrease in ASR.

The errors in climate simulations of cloud amount bear a strong resemblance to its projected changes for the twenty-first century. An exception is over the southern oceans, where cloud amount is simulated to be too small and is projected to increase. With observed zonal mean cloud amount approaching 90% coverage near 50°S but simulated amounts of order 80%, it seems likely that this increase is only possible because of the negative bias simulated in present-day climate. The very high amount of cloud is an indication that it is cloudy not only in cyclonic conditions, where models do well, but also in anticyclonic conditions, where low-level stratiform cloud is prevalent over the oceans in reality, but which is not well simulated (Walsh et al. 2009). Note that observations
suggest that stronger lower-tropospheric stability associated with inversions produces an increase in low cloud (Klein and Hartmann 1993). A more complete discussion is in Trenberth and Fasullo (2009).

Therefore in this study a basis for resolving issues related to simulation of the SH energy budget is presented. This basis consists of four major findings including 1) the disproportionately large biases that exist in both reanalysis and global coupled models in the energy budget of the SH that is directly linked to simulation of clouds, 2) the unusual nature of the future projected changes whereby ASR increases at all latitudes other than between about 50° and 60°S, 3) the remarkably strong relationship between the projected changes in clouds and the simulated current-day cloud errors, and 4) the strong relationship that exists between SH simulated radiation and model climate sensitivity. It follows that models errors in the SH constitute a fundamental limitation to our ability to simulate even global metrics related to climate change and are deserving of more attention.

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REFERENCES


