Robust Land–Ocean Contrasts in Energy and Water Cycle Feedbacks*

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ABSTRACT

Building on recent observational evidence showing disproportionate increases in temperature and aridity over land in a warming climate, this study examines simulated land–ocean contrasts in fully coupled projections from the Third Coupled Model Intercomparison Project (CMIP3) archive. In addition to the projection of disproportionate changes in temperature and moisture over land, the analysis reveals contrasts in clouds and radiative fluxes that play a key role in the eventual equilibration of the planetary energy budget in response to forcing. Despite differences in magnitude, the nature of the feedbacks governing the land–ocean contrast are largely robust across models, notwithstanding the large intermodel differences in cloud parameterizations, and suggest the involvement of fundamental constraints.

The model responses are consistent with previously proposed ideas maintaining that relative humidity (RH) over land decreases with warming because precipitation and the hydrological cycle are governed primarily by transports of moisture from the oceans, where increases in lower-tropospheric temperature and saturated humidity fail to keep pace with those over land. Here, it is argued additionally that constraints on RH imply systematic changes in the cloud distribution and radiative feedbacks over land, as decreased RH raises the lifting condensation level, even as moist instability increases, and suppresses convective clouds. This effect is shown to be particularly strong at low latitudes where the dynamical influence of competing sources of maritime deep convection may further suppress convection. It is found that as a result of the coincidence between strong warming and a muted net greenhouse feedback associated with decreases in RH and clouds, the mean increase in outgoing longwave radiation (OLR) over land ($1.0 \text{ W m}^{-2} \text{ K}^{-1}$) in transient simulations at 2200 is almost double that over the ocean ($0.6 \text{ W m}^{-2} \text{ K}^{-1}$), and a strong negative net top-of-atmosphere (TOA) radiative perturbation emerges as the simulations approach and attain equilibrium. However, over the oceans a positive radiative imbalance persists and the increase in water vapor and other greenhouse gases does not allow a local TOA equilibration to occur. The contrast results in an increase in the transport of energy from ocean to land relative to the twentieth century that is accompanied by lasting increases in both OLR and absorbed shortwave radiation globally.

A conceptual model to describe the simulated variability is proposed that involves the following: 1) the differing albedos and lower-tropospheric lapse rates over land and ocean, 2) the nonlinearity of the saturated lapse rate in a warming environment, and 3) the disproportionate response in temperature, moisture, clouds, and radiation over land versus ocean. It is noted that while the land–ocean contrast plays a key role in achieving global radiative equilibrium, it entails disproportionate increases in temperature and aridity over land and therefore is likely to be associated with substantial environmental impacts.

1. Introduction

Identifying and understanding the character of a climate system's response to externally imposed forcing remains an objective of chief importance, particularly given the wide disparity that exists among coupled model projections of anthropogenic climate change. Perhaps chief among the responses that are robust across models is the differential warming exhibited by land and ocean

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regions, a feature that is ubiquitous in both early and modern coupled general circulation models (Manabe et al. 1990, 1991; Joshi et al. 2008). Fundamental constraints governing the contrast have been suggested, including the nonlinearity of the moist-adiabatic lapse rate in a warming environment and its disproportionate influence over the ocean (Joshi et al. 2008), as well as the partitioning of the surface energy budget (Sutton et al. 2007). It has yet to be established, however, whether land–ocean contrasts play an important role in the energy budget on a global scale or what role, if any, they play in planetary equilibration. In this paper, the role of land–ocean contrasts is highlighted as a key aspect of the changes in the energy flow through the climate system and the equilibration of the planetary energy balance as the radiative forcing and climate stabilize.

While individual processes can be scrutinized in evaluating their potential impacts on climate, it is useful to evaluate the system holistically, using the flow of energy as a framework for establishing a hierarchy of various feedbacks. The movement of energy through the climate system is complex and is associated with a variety of mechanisms involved in its absorption, transport, storage, and emission (Trenberth et al. 2009). Energy enters the system as solar radiation, approximately 70% of which is absorbed by the atmosphere or surface. The sun–earth geometry, partly modified by clouds and other factors such as surface albedo and aerosols, contributes to large latitudinal gradients in absorption. As a result, energy is stored within, and transported and released by the atmosphere, oceans, and, secondarily, the land surface, so as to reduce temperature gradients generally, thus contributing to a more uniform emission of outgoing longwave radiation (OLR) than would otherwise occur. Large land–ocean contrasts in this energy flow also exist and the net annual mean transport from ocean to land regions exceeds 2 PW, though it is much greater in northern winter (Fasullo and Trenberth 2008a). As the net terrestrial storage tendency is small on time scales greater than 1 yr, the land–ocean exchange is balanced approximately by the net radiative flux at TOA (R_T) over land. Summaries of these flows have recently been provided for the globe (Trenberth et al. 2009), for global land and ocean domains (Fasullo and Trenberth 2008a), for the zonal mean (Fasullo and Trenberth 2008b), and for the ocean (Trenberth and Fasullo 2008).

But how do these flows change in response to forcing and what role do land–ocean contrasts play? As greenhouse gases increase in the atmosphere, the initial response is a decrease in OLR (Randall et al. 2007; Levinson 2008; Murphy et al. 2009), which in turn drives warming, a moistening of the atmosphere, and an enhanced hydrologic cycle (e.g., Trenberth et al. 2005). Among the most

robust model feedbacks, are those dealing with water vapor; and under the assumption of constant relative humidity (RH) (Wetherald and Manabe 1988; Cess et al. 1990), the feedback roughly doubles the warming response (Randall et al. 2007; Murphy et al. 2009). Yet, the water vapor feedback is also likely to be central to changes in the transport of energy within the system as well and, in particular, between ocean and land. Due to land–ocean contrasts in albedo, the diurnal cycle, and temperature, the cloud feedback is also likely to be central and strong reductions in cloud amount are endemic to the models (Trenberth and Fasullo 2009).

While the gross features of present-day land–ocean transports are reasonably simulated in many models (Shin et al. 2006), trends in the flow have yet to be widely assessed and the role of land–ocean contrasts in the global energy budget generally is largely uncertain. Of particular interest in this study is their role as the system approaches equilibrium. It is known that the Planck response to warming, whereby longwave emissions increase in proportion to T^4 , exerts a strong stabilizing influence and is fundamental to equilibration. Yet, its TOA manifestation is determined both by the magnitude of the warming of a given component of the climate system, weighted by that component's effective ability to radiate to space. Many constituents of the climate system, and particularly the land and ocean surfaces, are obscured from TOA by intermediaries such as clouds and water vapor, and thus while increases in emissions with warming within the system are large, their impacts on the planetary balance are often highly muted. Therefore, higher temperatures must coincide with higher transmission through the atmosphere in order to enable a net cooling from the Planck response.

In this study, it will be shown that strong regional contrasts in this coincidence between warming and transmission exist generally, and that such contrasts are particularly significant between land and ocean. Lower land heat capacity, higher temperatures, and other mechanisms that persist though equilibrium, combined with a dependence on oceans as a moisture source, are shown to result in muted water vapor increases and reduced cloud over land, thus providing such a mix. Simplified slab-ocean equilibrium model runs are used along with fully coupled multicentury simulations of the Third Coupled Model Intercomparison Project (CMIP3; Meehl et al. 2007) archive, using the energy budget as a basis for diagnosing the role of land–ocean contrasts. Section 2 describes the models and methods, while section 3 examines the temporal evolution of the global fields. Section 4 documents the regional patterns of the contrasts and explores the association of water vapor and clouds, through diagnosis of their spatial and temporal relationships. In

section 5, a physical basis for the observed variability is proposed while further discussion and conclusions are given in section 6.

2. Methods and model projections

To help gauge the net radiative effects of land–ocean contrasts, the simplified equilibrium (control and $2 \times \text{CO}_2$) and fully coupled transient simulations [climate of the twentieth century, and Intergovernmental Panel on Climate Change (IPCC) Special Report on Emissions Scenarios (SRES) scenario A1b] of the CMIP3 archive are examined (summarized in Table 1). The control simulation incorporates preindustrial CO_2 concentrations of 280 ppmv within a modeling framework that includes a single-layer ocean model in order to expedite equilibrium. The $2 \times \text{CO}_2$ scenario simulates the control configuration's response to an instantaneous doubling of CO_2 from preindustrial concentrations and is initialized with the control run's equilibrium conditions. Both simulations are performed long enough to produce stable statistics over their last 50 yr, although as not all simulations provide a full 50-yr record to the archive, only the last 30 yr of the simulations are used here. While expediting the equilibrium, the slab-ocean component represents a significant simplification of the climate system's coupled response and limits the realism of ocean variability and its associated air–sea interactions. Cloud feedbacks are particularly sensitive to the use of a slab model (Yokohata et al. 2008), and, regionally, these issues are pronounced where the role of ocean dynamics is large (Danabasoglu and Gent 2009). The use of carbon dioxide as a single greenhouse gas forcing in these runs is also highly idealized.

To address these shortcomings, a more complete modeling framework and set of forcing agents is used in the twentieth-century and SRES A1b (A1b) simulations, which include historical and idealized emission scenarios of multiple greenhouse gases [CO_2 , CH_4 , N_2O , and chlorofluorocarbons (CFCs)] and sulfate aerosols. The estimated variability in solar and twentieth-century volcanic forcings is also incorporated within many of the simulations; however, given the wide range of individual forcings imposed by the various modeling centers (Meehl et al. 2007), it is nontrivial to discern the intermodel contrasts in radiative feedbacks from that in the forcings. The twentieth-century and emission scenario simulations are coupled to a full-ocean model and therefore the nature of their coupled response is considerably more complete than for the equilibrium simulations. For the A1b scenario, several simulations extend to 2300 with concentrations fixed at 2100 levels (750 ppmv), which is considerably larger than for the $2 \times \text{CO}_2$ runs (560 ppmv). Despite extending simulations under fixed forcing after

2100 for several centuries, the A1b simulations do not demonstrate stable statistics by 2300, an indication of the long equilibration time scale inherent in the fully coupled system (Meehl 2005).

For purposes of compositing and comparison, all fields are regridded from their native grid onto a T-63 grid (192×96) using bilinear interpolation. Substantial differences exist in the duration of simulations submitted to the archive and, thus, of the 24 simulations included in the general archive, only 12 continue to 2300 and are therefore considered in this analysis. Twelve models are also included in the control and $2 \times \text{CO}_2$ experiment archive but not all of these submitted extended A1b simulations. To consider as wide a range of models as is possible, additional model screening is not employed in the present analysis, though concerns exist regarding the use of energy flux corrections, missing and erroneous fields, and the absence of ozone effects in some simulations (Trenberth and Fasullo 2009, 2010). To remove systematic biases, anomalies in the energy budget relative to the early half of the twentieth century, during which time the external forcing is small (Trenberth and Fasullo 2009), are considered in gauging feedbacks in the A1b simulations. Further highlighting the challenge of achieving an adequate sample size, several of the models (Table 1) share a common lineage and physical parameterizations. Thus, despite including 12 transient and equilibrium simulations, the number of independent realizations is probably not more than 10.

3. The perturbed global energy budget: 2000–2300

The evolution of surface temperature anomalies from 2000 to 2300 relative to the 1900 to 1950 mean for land, ocean, and their difference, is shown in Fig. 1. During the twenty-first century, as forcing increases, land regions warm by 4.7 ± 0.8 K while the oceans warm by only 3.1 ± 0.5 K and both the mean warming and its spread among simulations are greater for land than for ocean. The land–ocean warming contrast in 2100 is 1.5 ± 0.4 K and while the contrast increases dramatically during the twenty-first century, it is nearly constant afterward, despite continued warming over both land and ocean. Thus, the growth in the contrast coincides most directly with the forcing rather than with adjustments associated with the warming itself. Similarly, when the contrast is expressed as a ratio (e.g., Sutton et al. 2007; Lambert and Chiang 2007; Joshi et al. 2008), rather than an absolute difference, its largest values occur in the twenty-first century in most models. For all models, the positive land–ocean differential is robust and in no simulation does ocean warming exceed that over land. In equilibrium runs (both Fig. 1 and Table 2), warming and the contrast are similar in character but

TABLE 1. List of models considered in this analysis. Simulation “run1” was used unless otherwise indicated. See Randall et al. (2007) for details. The model data are available online (<http://www-pcmdi.llnl.gov/>).

Model	Agency and country	Atmospheric levels	Atmospheric resolution	Ozone?/ Volcano?
Coupled General Circulation Model, version 3.1 [CGCM3.1(T47)] ^{a,b,c}	Canadian Centre for Climate Modeling and Analysis (CCCma), Canada	31	T47	No/no
CCCma T63 ^{a,b,c}	CCCma	31	T63	No/no
Coupled Global Climate Model, version 3 (CM3) ^c	Centre National de Recherches Météorologiques (CNRM), France	45	T63	Yes/no
Mark version 3.0 (Mk3.0) ^b	Commonwealth Scientific and Industrial Research Organisation (CSIRO), Australia	18	T63	Yes/no
CSIRO Mk3.5 ^c	CSIRO	18	T63	Yes/no
Climate Model version 2.0 (CM2.0) ^c	National Oceanic and Atmospheric Administration/Geophysical Fluid Dynamics Laboratory (NOAA/GFDL), United States	24	2° × 2.5°	Yes/yes
CM2.1 ^c	NOAA/GFDL	24	2° × 2.5°	Yes/yes
Model E-R (GISS-ER) ^{b,c}	National Aeronautics and Space Administration Goddard Institute for Space Studies (NASA GISS), United States	20	4° × 5°	Yes/yes
Coupled Model, version 3.0 (INM-CM3.0) ^b	Institute for Numerical Mathematics (INM), Russia	21	4° × 5°	No/yes
Model for Interdisciplinary Research on Climate 3.2, high-resolution version [MIROC3.2(hires)] ^b	Center for Climate System Research, University of Tokyo; the National Institute for Environmental Studies; and Frontier Research Center for Global Change (JAMSTEC), Japan	56	T106	Yes/yes
MIROC(medres) ^{b,c}	JAMSTEC	20	T42	Yes/yes
ECHAM and the global Hamburg Ocean Primitive Equation (ECHO-G) ^{a,c}	Meteorological Institute of the University of Bonn, Germany, and Meteorological Research Institute of the Korea Meteorological Administration (KMA), and the Model and Data Group, Korea	30	T30	No/no
ECHAM5 ^{b,c,d}	Max Planck Institute for Meteorology (MPI), Germany	31	T63	Yes/no
Coupled General Circulation Model, version 2 (CGCM2) ^{a,b,c}	Meteorological Research Institute, Japan	30	T42	No/no
Parallel Climate Model, version 1 (PCM1) ^{c,d,f}	National Center for Atmospheric Research (NCAR), United States	18	T42	Yes/yes
Community Climate System Model, version 3 (CCSM3) ^{b,c,d,e}	NCAR	26	T85	Yes/yes
Hadley Centre Global Environmental Model version 1 (HADGEM1) ^b	Met Office, United Kingdom	38	1.3° × 1.9°	Yes/no

^a Energy flux corrections are imposed.

^b Equilibrium control and 2 × CO₂ runs are used here.

^c Climates of the twentieth-century and SRES-A1b runs are used here.

^d Run2 of the archive was used due to insufficient data for run1.

^e Net TOA flux is not available in the A1b archive.

^f OLR is not available to 2300 from the archive.

smaller in magnitude than that achieved by 2300 in the A1b simulations, as may be expected from the weaker imposed forcing in the equilibrium runs. The contrast in warming has been studied extensively elsewhere (Manabe

et al. 1991; Sutton et al. 2007; Joshi et al. 2008) and involves basic constraints on the surface energy budget, and the disproportionate and nonlinear effects of the moist-adiabatic lapse rate over ocean in a warming climate.

FIG. 1. Mean changes in land and ocean temperatures, and their contrast, from 2000 to 2300 for the 12 runs in the CMIP3 archive extending to at least 2300. Shading depicts the interquartile range in the simulations. Also shown at specified intervals are the mean and standard deviation of the temperature distribution (Table 1) and equilibrium values (right column) where lines span the interquartile range and circles are centered on the median. A 10-yr running smoothing has been applied.

Through the twenty-second and twenty-third centuries, warming persists and climate adjustment continues (Meehl et al. 2006), although the rate of warming is not large. Thus, despite continued transient conditions, these simulations may also offer important insights into the eventual equilibration of the fully coupled system that are not resolved by a slab-ocean configuration.

The warming of the climate system is associated with a perturbed TOA energy budget, and the integrated anomalies of these fluxes provide broad insights into the system's response (Fig. 2). The integrated net TOA radiative flux anomaly ($\int R_T'$) is characterized by a gradual increase from 2000 through 2300 between 4 and 5 YJ (1 yottajoule = 10^{24} J), approximately equivalent to a 1 W m^{-2} mean imbalance over the period, which is substantially smaller than for anomalies in shortwave ($\int \text{ASR}'$) and longwave ($\int \text{OLR}'$) fluxes, which each attain magnitudes approaching 10–15 YJ on average by 2300. The compensation of ASR and OLR anomalies is a characteristic of the interspectral compensation that is common among the climate system's dominant feedbacks, such as those involving clouds (Cess et al. 1990). Similarly, the intermodel scatter in $\int R_T'$ is also narrow, and as for surface temperature (Fig. 1), $\int R_T'$ increases for several centuries following the imposed forcing. However, by 2300 the rate of change in $\int R_T'$ is small relative to that during the twenty-first century and it

continues to decrease, suggesting that planetary equilibration in the energy budget, while not reached, is being approached.

The TOA imbalance in the twenty-first century is driven principally by shortwave feedbacks associated with clouds and secondary effects related principally to surface snow and ice extent (Soden and Held 2006; Trenberth and Fasullo 2009). Increases in water vapor, which augments ASR and reduces OLR, are also likely to be important. Offsetting these positive feedbacks are the emission (i.e., Planck) and lapse rate negative feedbacks. In contrast to $\int R_T'$, both $\int \text{ASR}'$ and $\int \text{OLR}'$ continue to increase substantially through the twenty-third century. Their evolution is suggestive of a lasting shift in the partitioning of ASR and OLR from early twentieth-century values. Moreover, as both albedo and temperature anomalies increase through 2300 and drive the moistening of the atmosphere and the melting of snow and ice, a physical basis exists for the expectation that the ASR and OLR

TABLE 2. Global mean equilibrium TOA (W m^{-2}) differences between $2 \times \text{CO}_2$ and slab control experiments.

	25%	Median	75%
R_T	<0.1	0.1	0.2
OLR	1.6	3.2	4.2
ASR	1.8	2.9	5.4

FIG. 2. The mean evolution patterns of the integrated anomalies in absorbed solar, net, and OLR are shown for simulations extending to 2300 along with their interquartile distributions. The sign convention employed is such that increases in flux anomalies correspond to a net warming of the system.

anomalies are lasting. Initially, in the late twentieth and early twenty-first centuries, the enhanced greenhouse effect overrides the Planck response and anomalies in OLR are negative such that they augment ASR imbalances. However, in conjunction with substantial and continuing increases in temperature and OLR in the twenty-second and twenty-third centuries, a balance in R_T is approached. The forced equilibrium responses of ASR and OLR in slab-ocean runs (Table 2) exhibit similar fundamental characteristics as the transient simulations, with strong equilibrium increases in ASR and OLR that are robust across the simulations.

4. Regional and zonal-mean structures

Trends in OLR in the twenty-first century exhibit increases in the global mean that are ubiquitous across the simulations; however, their spatial structure is complex and demonstrates fundamental contrasts between land and ocean (Fig. 3). Over ocean, OLR trends are modest generally ($0\text{--}4.5 \text{ W m}^{-2} \text{ century}^{-1}$) and in some regions, such as in the deep tropics and at high latitudes over the southern oceans (Trenberth and Fasullo 2010), they are negative ($<-4.5 \text{ W m}^{-2} \text{ century}^{-1}$). In stark contrast, OLR increases over land are strongly positive ($4.5\text{--}9 \text{ W m}^{-2} \text{ century}^{-1}$) and are apparent over all continents, including Antarctica. In the zonal mean, OLR increases are systematically larger over land than over ocean, except north of 70°N where little land exists, and the contrast between land and ocean is pronounced between 30°S and 30°N , a zone where the areal extent of land is substantial and therefore significantly influences

the global budget. The land–ocean contrast in the OLR response is significantly greater than the model spread at low latitudes and thus there is robust model consensus that a substantial land–ocean contrast exists in the OLR feedbacks. Equilibrium simulations, basic aspects of which are summarized in Table 3, demonstrate a similar land–ocean OLR contrast as increases in OLR over land exceed those over ocean from 1.2 to 2.5 W m^{-2} . Thus, the negative feedbacks associated with increased OLR occur disproportionately over low-latitude land regions, a contrast that is common to both the transient and equilibrium simulations.

Variability in the upwelling flux (Fig. 4a) is primarily determined by surface temperature, which increases substantially over the course of the twenty-first century (Fig. 1). As longwave emissions are proportional to T^4 , their increase per degree of warming is substantially weighted toward warmer regions (i.e., low latitudes). Thus, the contribution of the greater total warming at higher latitudes to the upwelling flux is largely offset by this effect such that trends in the upwelling longwave flux, particularly over land, are not a strong function of latitude, with the exception of southern South America where maritime influences moderate the rate of warming over land considerably. Significant increases in upwelling flux are also apparent over ocean regions that trend toward ice-free conditions in summer and fall, such as in the Arctic; however, these regions have little influence on the global TOA budget as their areal extent is small and R_T anomalies are lessened substantially by compensating increases in ASR (Trenberth and Fasullo 2009). Thus, increases in the upwelling surface flux,

FIG. 3. Composite mean twenty-first century trends in OLR with median zonal-mean trends (dots) for ocean (blue), land (red), and all surface types (black). Lines about the zonal means indicate model interquartile ranges, and stippling and hatching indicate regions where the sign of the mean response is shared by at least $3/4$ of the models.

particularly as they relate to the global integral, are substantially biased toward land regions.

The surface net upward longwave flux is positive generally, with limited exceptions where a warm lower troposphere overlies a cool surface, such as can occur seasonally at high latitudes or in tropical oceanic upwelling zones. Trends in the twenty-first century net flux (Fig. 4b) are influenced by substantial increases in downwelling emissions associated with a warming and moistening atmosphere that exceed generally those in upwelling (Fig. 4a). Over the tropical ocean, the net effect is to decrease the net cooling (i.e., warm the surface) from 3 to $10 \text{ W m}^{-2} \text{ century}^{-1}$. To first order,

additional surface heating is balanced by increased evaporation and is a key source of energy for an enhanced hydrologic cycle (Held and Soden 2006). However over land, though the increases in downwelling are larger than those in upwelling, the trend in the net flux is systematically smaller than over ocean. This occurs in part because there is limited moisture available for evaporation (Sutton et al. 2007), there are enhanced seasonal and diurnal cycles of surface temperature, and these effects project nonlinearly onto emissions and result in substantial increases in net upwelling. The contrast is pronounced between 30°N and 30°S , where the range of simulated trends is considerably less than the land–ocean

TABLE 3. Summary of equilibrium changes in energy, clouds, and moisture over land, ocean, and their contrast between the control and $2 \times \text{CO}_2$ simulations [25%, 50% (median, bold), 75%].

Parameter global	Δ_{land}	Δ_{ocean}	$\Delta (\text{land} - \text{ocean})$
T_{sfc} (K)	3.4, 3.6 , 4.9	2.6, 2.9 , 3.7	0.6, 0.7 , 1.2
R_T (W m^{-2})	−1.6, −1.1 , −0.4	0.3, 0.6 , 0.8	−2.5, −1.8 , −0.6
OLR (W m^{-2})	2.6, 4.2 , 6.3	1.4, 2.8 , 3.3	1.2, 1.5 , 2.4
ASR (W m^{-2})	1.8, 2.9 , 5.4	1.7, 3.6 , 4.8	−0.8, 0.1 , 1.6
$\text{LW}\uparrow_{\text{sfc}}$ (W m^{-2})	18.2, 18.7 , 25.9	15.0, 15.4 , 19.9	3.3, 3.7 , 5.9
$\text{LW}\downarrow_{\text{sfc}}$ (W m^{-2})	21.0, 22.4 , 28.8	17.7, 20.0 , 24.2	2.0, 3.3 , 4.4
LW_{net} (W m^{-2})	−3.1, −2.7 , −2.2	−4.2, −3.9 , −3.8	0.7, 1.3 , 2.3
RH 950 mb	−0.6, −0.2 , 0.4	−0.2, 0.1 , 0.2	−0.4, −0.2 , 0.2
Total cloud (%)	−1.5, −0.9 , 0.5	−2.0, −1.4 , −0.6	−1.0, 0.5 , 1.8
PW (mm)	4.0, 4.3 , 6.0	5.6, 5.8 , 7.3	−1.6, −1.4 , −1.1

FIG. 4. Composite mean twenty-first century trends in (a) upwelling and (b) net surface longwave fluxes. Zonal means are indicated as in Fig. 3.

contrast itself (Fig. 4b, right). As for OLR, land-ocean contrasts in surface longwave fluxes between control and doubled CO_2 equilibrium runs (Table 3) exhibit the main features of transient simulations with systematically greater increases in upwelling flux over land and large decreases in the net flux over ocean. The land-ocean contrast in the net longwave flux in the equilibrium simulations increases by approximately $1\text{--}2 \text{ W m}^{-2}$.

Mechanisms that determine the contrast in surface longwave fluxes can be inferred from twenty-first century trends in 950-mb RH, total cloud amount, and precipitable water (PW; Fig. 5). In computing RH, an average across the 1000–900-mb layer has been computed and missing values associated with orography have been

screened. Trends in RH (Fig. 5a) are substantial and negative over mid- and low-latitude land regions and some inland seas, such as the Mediterranean Sea, and decreases exceeding $1\% \text{ century}^{-1}$ prevail over most land regions. A similar spatial structure in humidity reductions, albeit of reduced magnitude, is found across the lower troposphere in most models. A strong spatial coherence exists between decreases in RH and cloud amount (Fig. 5b), and cloud loss is pervasive across land regions equatorward of 60°N and 45°S . Moreover, land-ocean contrasts in cloud and humidity trends are considerable in the tropics and collectively suggest a coupling with dynamics whereby, as major maritime convective zones strengthen and tropical conditions become more El-Niño like (e.g.,

FIG. 5. Composite mean twenty-first century trends in (a) mean relative humidity between 1000 and 900 mb, (b) total cloud amount, and (c) PW. Zonal means are indicated as in Fig. 3. The humidity trends have been computed at levels where available data exist and have been screened to include only cases where the mean surface pressure is below 900 mb in at least $\frac{3}{4}$ of the models.

FIG. 6. Composite mean land-ocean contrasts in twenty-first century trends for (a) cloud amount, (b) RH, and (c) temperature as a function of latitude and height. Hatching and stippling denote regions where at least $3/4$ of the models agree on the sign of the contrast.

Meehl and Washington 1996), land regions become more subsident and thus drier. Increased subsidence over land has been confirmed in vertical velocity fields (see Fig. A1 in the supplemental material for this paper, <http://dx.doi.org/10.1175/2010JCLI3451.s1>), and will also be investigated further within the context of the vertical structure of these trends below. The inability of the moisture supply over land to keep pace with the substantial terrestrial warming that occurs in the twenty-first century and an associated exponential increase in saturation specific humidity (q_s) associated with the Clausius-Clapeyron relation is also apparent in the PW (Fig. 5c). Muted increases in PW are evident in regions of substantial cloud and RH reductions, such as over equatorial South America and both northern and southern Africa. These trends are in stark contrast to our expectations of constant RH (Wetherald and Manabe 1988; Cess et al. 1990) whereby increases over land would be markedly greater than over ocean due to the land's

disproportionate warming. Instead, at all latitudes, zonal mean increases in PW over land are actually smaller than over ocean, despite greater terrestrial warming.

The vertical profile of land-ocean contrasts exhibits significant and consistent structures of variability across the CMIP3 projections, with disproportionate negative trends across the twenty-first century in clouds and RH extending throughout the troposphere at low latitudes (Fig. 6). Strong decreases coincide through the depth of the troposphere at 45°S and between 20°S and 10°N , suggesting a physical linkage between the two fields, and the intermodel agreement is robust generally, with a similar sign of trends in at least $3/4$ of the simulations in the domains of strongest land-ocean contrast. In other regions, such as the Northern Hemisphere middle troposphere, both the anomalies and intermodel agreement are weaker and may be obscured by either internal model variability or the lack of a physical connection between the fields. It is noteworthy that the land-ocean

contrasts in clouds and RH extend deep into the troposphere, particularly equatorward of 40° , where the contrast in temperature trends between land and ocean is itself small (Fig. 6c). Thus, the broad vertical extent of RH and cloud anomalies suggests that processes beyond the immediate influence of boundary layer moisture associated with land–ocean exchanges are responsible for the contrasts. Rather, the vertical structure of the contrasts is suggestive additionally of dynamical influences, likely in association with variability in both local and remote convection, and their alteration of both subsidence and deep convection at low latitudes. While further attribution of causality in these trends would be provided from the computation of atmospheric moisture, mass, and energy budgets in the models, such computations are not possible given the limited data provided in the CMIP3 archive.

The equilibrium response of moisture and clouds (Table 3) in some respects bolsters the relevance of the transient simulations to equilibrium conditions, with larger decreases in RH and cloud amount over land than over ocean, and a disproportionately small increase in PW over land. However, a lack of consistency also exists in the land–ocean contrast in the RH and cloud feedbacks, as unlike in the transient simulations, greater decreases in cloud amount and RH are simulated over ocean than over land in some simulations. Inspection of the regional structure of the equilibrium cloud and RH responses (Figs. A2 and A3, <http://dx.doi.org/10.1175/2010JCLI3451.s1>), reveals that the main differences between transient and equilibrium sensitivities occurs primarily over ocean, as low-latitude anticyclonic domains experience a greater loss of stratocumulus clouds in the slab-ocean configuration per degree warming than they do in fully coupled runs. However, the severity of cloud loss in the equilibrium simulations in these regions is likely to be overstated, as a consequence of large SST increases arising from a simplified slab-ocean configuration that imposes a noninteractive Q-flux term in parameterizing the mixing at the base of the slab layer. The slab models thus do not permit realistic variability in equatorial and coastal upwelling and, in turn, the simulations are likely to misrepresent a key aspect of the air–sea coupling in the stratocumulus domains, and contribute to exaggerated total surface warming estimates. Associated SST biases are likely to preclude a realistic coupling to clouds as they distort the maintenance of the lower-tropospheric stability that sustains the subtropical stratocumulus decks. Moreover, in comparing the fully coupled and slab-model simulations, it can be inferred that by preserving low cloud over ocean, air–sea feedbacks actively bolster the land–ocean contrast in temperature, humidity, and cloudiness.

The evolution of land–ocean contrasts in OLR from 2000 to 2300 (Fig. 7a) reveals a TOA response that is disproportionately strong over land. Anomalies in OLR over land are $4.4 \pm 1.2 \text{ W m}^{-2}$ by 2100 and $5.3 \pm 1.4 \text{ W m}^{-2}$ by 2200, and as with temperature (Fig. 1) the increases are strong during the twenty-first century and are gradual thereafter. Over ocean, OLR increases are substantially smaller than over land, at $1.2 \pm 1.7 \text{ W m}^{-2}$ by 2100 and $2.1 \pm 1.3 \text{ W m}^{-2}$ by 2200. The discrepancy between land and over ocean cannot be explained merely as a proportionate scaling associated with differences in surface warming, as by 2200, the response over ocean ($0.6 \text{ W m}^{-2} \text{ K}^{-1}$) is approximately half of that over land ($1.0 \text{ W m}^{-2} \text{ K}^{-1}$). Also, as has been explicitly shown, greenhouse feedbacks associated with the water vapor and clouds are systematically weaker over land than ocean. The contrast in OLR between land and ocean is $3.0 \pm 1.0 \text{ W m}^{-2}$ in both 2100 and 2200. As in temperature, the OLR contrast between land and ocean increases in conjunction with the forcing (2000–2100) while remaining approximately constant thereafter. The interquartile range of the model spread in the contrast, which differs little from the full model range, is also considerably less than the contrast itself, demonstrating its robustness to the parameterizations of any individual model. Finally, the significance of the land–ocean contrast in radiation is underscored by comparing it with that in temperature (Fig. 1). Unlike the contrast in temperature, the radiative contrast between land and ocean exceeds the ocean anomaly itself, whereas for temperature it is significantly less than the warming over either land or ocean. Equilibrium simulations (Fig. 7a, Table 3) exhibit many of the characteristics of the transient simulations, including a larger increase in OLR over land than over ocean, and a considerable land–ocean differential, the spread of which is much smaller than the mean contrast itself. The equilibrium response in the OLR contrast varies from 1.2 to 2.4 W m^{-2} .

Associated with contrasts in OLR, land–ocean contrasts in R_T (Fig. 7b) are also substantial and demonstrate the central role played by land in driving the energy budget toward equilibrium in response to an imposed forcing. Initially in the twentieth and early twenty-first centuries, R_T increases over both land and ocean, in part due to changes in clouds but also due to melting of surface snow and ice and increases in water vapor and other greenhouse gases (Trenberth and Fasullo 2009; Murphy et al. 2009). Globally, R_T reaches a maximum over both land and ocean during the later half of the twenty-first century, subsequent to which a strong reduction in R_T occurs over land through the twenty-second and twenty-third centuries. By the end of the twenty-third century, R_T over land is negative for about $\frac{3}{4}$ of the simulations

FIG. 7. Evolution of global-land and global-ocean mean (a) OLR and (b) net radiation from 2000 to 2300 UTC. Zonal means are indicated as in Fig. 3.

FIG. 8. Seasonality of the land–ocean contrast in twenty-first century OLR trends is shown for the globe for low latitudes in the Southern (0° – 30° S) and Northern Hemispheres (0° – 30° N) and for the extratropics (for 30° – 90° in both hemispheres). Gray bars indicate the model interquartile range and seasons in the extratropics are examined for winter (DJF), spring (MAM), summer (JJA), and fall (SON); and in addition all months (AN).

and exhibits a negative trend in all projections. The flux contrasts greatly with that over ocean, which while lessening somewhat by 2300 remains robust and positive in all simulations through at least 2300. The contrast in R_T between land and ocean is therefore also significant and is $-1.5 \pm 1.6 \text{ W m}^{-2}$ in 2100, $-1.7 \pm 1.8 \text{ W m}^{-2}$ in 2200, and continues to increase through 2300. In equilibrium simulations (Table 3), the mean contrast is of similar magnitude (-1.7 W m^{-2}) and ranges from -2.5 to -0.6 W m^{-2} . Despite the issues with air–sea interactions related to the slab model clouds already discussed, equilibrium responses largely bolster the interpretation of the energy budget from the transient runs, with land–ocean contrasts in R_T of similar magnitude to the transient simulations, despite being subject to weaker forcing. In addition, despite the limited spatial extent of land relative to ocean, area-integrated R_T anomalies over land and ocean mutually compensate, with each amounting to approximately 0.2 PW. The land–ocean contrast thus exerts a key influence in the evolution toward planetary equilibration. Moreover, in conjunction with changes in

the partitioning of R_T between ASR and OLR (Fig. 2), a robust and lasting partitioning of R_T between the land and ocean regions is evident. This finding is both consistent with and complementary to the documented role of land–ocean exchanges in regulating the temperature contrasts in a changing climate shown by Lambert and Chiang (2007).

As land regions experience significantly greater seasonal variability in temperature and some aspects of the energy budget than the oceans (Fasullo and Trenberth 2008a), it is anticipated that seasonal contrasts in trends may also exist, particularly at low latitudes where interactions with the monsoons are likely. To explore this hypothesis, seasonal variability, defined as the standard deviation of seasonal means, is quantified for the land–ocean contrast in twenty-first century OLR trends across various latitudinal bands aimed at delineating both the tropical monsoon domains and the extratropics (Fig. 8). In the global mean, trends in the OLR contrast are not large generally and the annual mean ($2.6 \text{ W m}^{-2} \text{ century}^{-1}$) exhibits modest seasonal variability ($\pm 0.2 \text{ W m}^{-2} \text{ century}^{-1}$).

- 10.1175/2010JCLI3451.s1) and the energy required to trigger convection. The vertical extent of the contrasts in RH and clouds extends well beyond that in T (Fig. 6), suggesting a key role for dynamic–convective couplings, both locally and remotely.
- 7) The cloud amount decreases (Figs. 5b and 6a) in response to reduced RH, the elevated LCL (Fig. A7, <http://dx.doi.org/10.1175/2010JCLI3451.s1>), and associated stomatal feedbacks (Doutriaux-Boucher et al. 2009), and the distribution of convection shifts toward fewer and more intense events. Related runoff effects at the surface may further augment aridity.
 - 8) As a result of greater warming and a weaker net greenhouse feedback, land regions exhibit an OLR sensitivity that is almost twice as large as that over the ocean (Fig. 7a) and drives lasting land–ocean contrasts in OLR and R_T .

While some aspects of the conceptual model have been proposed in previous works, several aspects of the conceptual model are original and many remain speculative and require further validation. For example, the fractional influence of the dry and moist lapse rates over land and ocean (k) is not constant and may change as the climate changes. Indications from the CMIP3 simulations are that the decrease in k is likely to be largest over land, where the mean state is driest (e.g., Fig. A5, <http://dx.doi.org/10.1175/2010JCLI3451.s1>) and decreases in RH are largest, and this effect will thus augment further the surface warming differential between land and ocean. However, as the moisture content of the boundary layer over land also relies on precipitation and a variety of surface processes including evaporation, transpiration, and runoff, the trends of which contain significant uncertainty, the robust projection across models of substantial RH decreases over low-latitude land regions nonetheless contains substantial uncertainty. Recent analysis of the stomatal response supports the conceptual model (Doutriaux-Boucher et al. 2009), though field studies are as of yet unclear regarding the ability of a 10%–20% change in stomatal conductance to regulate RH and cloud amount on a broad scale (e.g., Bonan 2008). Moreover, the potential influence of the stomatal effect is clearly limited in regions where vegetation is not widespread, such as over Australia and the deserts of northern Africa and southwestern Eurasia—regions where large reductions in RH and cloud amount are nonetheless simulated. Meanwhile, observations show that RH reductions may already be playing an important role on a broad scale (Simmons et al. 2010), thus bolstering confidence in the realism of the CMIP3 simulations. Nonetheless, observational confirmation

of the full vertical extent of the contrast in RH and clouds suggested by models may be elusive, due to significant biases, drift, and discontinuities in the radiosonde record. Continued analyses of the processes involved in the interaction between warming, humidity reductions, and cloud loss over land will improve our ability to attribute radiative feedbacks to specific processes and will further increase confidence in the processes that drive land–ocean gradients and their energy budget impacts.

6. Discussion and conclusions

The land–ocean contrast in various feedbacks in response to anthropogenic greenhouse forcing has been assessed and documented in equilibrium and transient runs from the CMIP3 archive. The contrast involves differential surface warming and related responses in the water and energy cycles that act collectively such that the net TOA radiative flux anomaly over land is generally negative by 2300, while over ocean, it remains substantial and positive. Despite the simplified slab-ocean framework employed, equilibrium simulations replicate many features of the fully coupled transient runs, though shortcomings in the air–sea interaction of the slab-ocean simulations are also highlighted. Discrepancies between the fully coupled and slab runs suggest further that ocean dynamics, particularly in low-latitude anticyclonic regions, play an important role in driving and sustaining aspects of the land–ocean contrast. Associated contrasts in TOA feedbacks are pronounced at low latitudes, despite the region’s muted surface warming, due to the nonlinear relationships exhibited between temperature and both longwave emissions and saturation specific humidity. Moreover, a lasting shift in the net TOA flux partitioning between land and ocean is demonstrated, implying an increase in the transport of energy from ocean to land. Despite the reliance of these feedbacks on variability in the hydrologic cycle and clouds, their general character is robust across fully coupled models in the CMIP3 archive.

The mechanisms that collectively drive this response have therefore been investigated. It is known that under climate change conditions a disproportionate warming occurs over land, largely as a result of contrasts in aridity and lower-tropospheric lapse rates over ocean and land. While the total moisture in the boundary layer increases with warmer temperatures, the increase over land fails to keep pace with the exponential rise in q_s and RH thus decreases. Consistent with earlier studies, this is interpreted as being a likely result of constraints on the moisture transport from ocean associated with its muted warming but is also likely to involve interactions with

vegetation. Recently reported RH trends and proposed stomatal feedbacks bolster support for the model projections of increased aridity, soil drying, and cloud loss, though caveats are discussed. A dynamic suppression of convection over land is also suggested by the deep vertical structure of the land–ocean contrast in clouds and RH at low latitudes that coincide with enhanced tropical deep convection over ocean. Thus, despite increases in moist static energy and instability, the level of free convection over land is raised and both the triggering and maintenance of moist convection is inhibited. This simulated response is consistent with expectations of convective variability over land that demonstrate a shift toward more intense and less frequent events as has been hypothesized (Trenberth et al. 2003), modeled (Cubasch et al. 2001), and observed (Goswami et al. 2006; Zhai et al. 2005). The shift is also associated with a reduced frequency of frontal precipitation and increased occurrence of deep convective systems, which are associated with strong updrafts and broad regions of descent, thus compounding the cloud loss in the middle and lower troposphere. The net result of the increase in convective inhibition and the shift in the convective distribution is thus a disproportionate loss of cloud relative to that over ocean. This in turn has immediate radiative consequences, especially in regard to the thermal radiation that is allowed to escape to space given the large diurnal and seasonal cycles of temperature and upwelling longwave radiation that are present over land. Feedbacks related to the partitioning of moisture convergence over land into runoff may also increase aridity, though these are not a central requirement of the proposed conceptual model.

The related impacts of the land surface on the climate system's approach to equilibrium are shown to be significant. A net cooling radiative flux anomaly over land is identified in both transient and equilibrium simulations that offsets positive feedbacks over ocean and hastens the climate system's stabilization in response to forcing. It is demonstrated that land regions may exert an important influence in determining climate sensitivity, suggesting fundamental limitations in the suitability of aquaplanet configurations for exploring such issues. The dependence of these feedbacks on latitude suggests further that both land's areal extent and its shifting latitudinal distribution over paleoclimatic time scales can alter both the time scale and efficacy of the climate system's feedbacks, presenting a physical basis for associated changes in climate sensitivity over time.

The relevance of the proposed set of feedbacks to the current climate variability may also be large. Recent work aimed at understanding the oceans' role in influencing variability over land identifies strong asymmetries in the interaction (e.g., Compo and Sardeshmukh

2009; Dommenget 2009). The question is thus raised as to whether the feedbacks proposed here are also relevant to the interpretation of present-day variability. Also of interest is whether observations of such variability can be used to evaluate simulated model feedbacks. A cursory survey performed here of simulated interannual variability in cloud amount over land during the twentieth century finds a strong negative correlation (-0.58) with equilibrium climate sensitivity in the CMIP3 models, whereas the relationship for cloud amount over ocean is very weak (-0.02). While being both compelling and consistent with the perspective that feedbacks in clouds over land can exert a disproportionate stabilizing affect on the energy budget, this finding is put forth here as motivation for future study in applying observable fields to models and interpreting their role in determining the simulated sensitivity.

Finally, a main caveat of the findings herein relates to the existence of systematic biases across models that influence the land–ocean contrast, particularly as the biases relate to inadequacies in the treatment of clouds, the diurnal cycle, and processes that are omitted from the current generation of models. Improving our understanding of these effects as they relate to both our current changing climate and the historical record will bolster the results presented here and continues to be a science goal of considerable importance.

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