Understanding Land–Sea Warming Contrast in Response to Increasing Greenhouse Gases. Part I: Transient Adjustment

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ABSTRACT

Climate model simulations consistently show that surface temperature over land increases more rapidly than over sea in response to greenhouse gas forcing. The enhanced warming over land is not simply a transient effect caused by the land–sea contrast in heat capacities, since it is also present in equilibrium conditions. This paper elucidates the transient adjustment processes over time scales of days to weeks of the surface and tropospheric climate in response to a doubling of CO2 and to changes in sea surface temperature (SST), imposed separately and together, using ensembles of experiments with an atmospheric general circulation model. These adjustment processes can be grouped into three stages: immediate response of the troposphere and surface processes (day 1), fast adjustment of surface processes (days 2–5), and adjustment of the whole troposphere (days 6–20).

Some land surface warming in response to doubled CO2 (with unchanged SSTs) occurs immediately because of increased downward longwave radiation. Increased CO2 also leads to reduced plant stomatal resistance and hence restricted evaporation, which increases land surface warming in the first day. Rapid reductions in cloud amount lead in the next few days to increased downward shortwave radiation and further warming, which spreads upward from the surface, and by day 5 the surface and tropospheric response is statistically consistent with the equilibrium value. Land surface warming in response to imposed SST change (with unchanged CO2) is slower. Tropospheric warming is advected inland from the sea, and over land it occurs at all levels together rather than spreading upward from the surface. The atmospheric response to prescribed SST change in about 20 days is statistically consistent with the equilibrium value, and the warming is largest in the upper troposphere over both land and sea. The land surface warming involves reduction of cloud cover and increased downward shortwave radiation, as in the experiment with CO2 change, but in this case it is due to the restriction of moisture supply to the land (indicated by reduced soil moisture), whereas in the CO2 forcing experiment it is due to restricted evaporation despite increased moisture supply (indicated by increased soil moisture). The warming over land in response to SST change is greater than over the sea and is the dominant contribution to the land–sea warming contrast under enhanced CO2 forcing.

1. Introduction

Coupled atmosphere–ocean general circulation models (CGCMs) are among the most powerful tools both to

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in the IPCC Fourth Assessment Report (IPCC AR4; Meehl et al. 2007). The CGCMs show some important similar features in their projected surface air temperature changes. In particular, climate models consistently show that warming is greater over land than over sea (Manabe et al. 1991; Stouffer and Manabe 1999; Murphy and Mitchell 1995; Cubasch et al. 2001; Lambert and Chiang 2007; Meehl et al. 2007; Sutton et al. 2007; Joshi et al. 2008; Compo and Sardeshmukh 2008; Dommenger 2009).

Using IPCC AR4 model integrations, Sutton et al. (2007) investigated the tendency for greater warming over land than over sea in response to greenhouse gas forcing. In all the 20 CGCMs examined, warming over land exceeds warming over sea; that is, the land–sea warming ratio is greater than 1. They have further illustrated that for any given model, the warming ratio in the presence of increasing radiative forcing is fairly constant in time, implying that the land–sea temperature difference increases with time. Furthermore, the enhanced warming over land is not principally a transient effect caused by the greater heat capacity of the ocean; although this is a contributory factor, the contrast is present to nearly the same extent if CO$_2$ is held constant at twice its initial value in CGCM integrations and in equilibrium conditions for double CO$_2$ in models with mixed-layer (“slab”) oceans, consistent with early studies using atmosphere–mixed layer ocean models (e.g., Manabe and Stouffer 1980; Hansen et al. 1988).

Joshi et al. (2008) reached a similar conclusion by using an atmosphere GCM (AGCM) forced with prescribed sea surface temperatures (SSTs). They found that if SST is raised uniformly by 4 K, with no externally imposed radiative forcing, the land warms by more than 4 K, and the land–sea warming contrast is similar to that found in CO$_2$-forced experiments. They showed that the land–sea warming contrast is associated with different changes in lapse rate over land and sea, restriction of moisture transport from sea to land, and local feedbacks associated with the hydrological cycle and cloud changes over land. The nonlinear dependence of saturation specific humidity on temperature plays a critical role in the changes in specific humidity and lapse rate.

Understanding the land–sea warming contrast is the subject of this study. What especially distinguishes land from sea is the supply of moisture for evaporation. The limited capacity for water storage on land means that a much wider range of surface moisture conditions is realized over land in proceeding from arid to moist climatic zones, and this has a strong effect on the nature of variations in surface energy budget, cloud, precipitation, atmospheric temperature, and humidity. Other aspects of the role of changes in the hydrological cycle in climate change have been studied by various authors (e.g., Yang et al. 2003; Held and Soden 2006; Wang and Lau 2006).

Sutton et al. (2007) and Joshi et al. (2008) studied equilibrium climate change or transient climate change on annual time scales. However, the use of annual time averages obscures the processes through which the anomalies come about. We thus adopt a different approach in this paper by examining the transient adjustment processes, focusing on the daily evolution of climate anomalies over land and sea in the first month following a sudden doubling of CO$_2$ and/or imposition of SST change. The same method of specifying both the SST forcing and atmospheric radiative forcing in an atmosphere-only model has been employed to understand the relative roles of SSTs and greenhouse gases on observed surface temperature changes (Sexton et al. 2003; Compo and Sardeshmukh 2008), on regional climate sensitivity (Cash et al. 2005), and on observed atmospheric circulation trends (Deser and Phillips 2009).

Our key aim in this paper is to discover the mechanisms causing the initial development of the land–sea warming contrast under the two kinds of forcing. In section 2, the model and experiments are described briefly. The time-dependent global mean changes in surface air temperature in response to CO$_2$ change and imposed SST change are described and contrasted in section 3. Analysis of the transient adjustment in response to CO$_2$ change and to imposed SST change is presented in sections 4 and 5, respectively. A summary and conclusions appear in section 6. In a future paper we will investigate in more detail the equilibrium response to doubled CO$_2$ and prescribed SST change.

2. Model and experiments

The numerical model used is the UK Met Office Hadley Centre atmosphere general circulation model HadAM3 (Hadley Centre Atmospheric Model version 3). (Pope et al. 2000) at a horizontal grid resolution of $2.5^\circ \times 3.75^\circ$ with 19 levels in the vertical. The model simulates the diurnal cycle and calculates radiation fluxes every 3 h using the radiation scheme developed by Edwards and Slingo (1996) and modified by Cusack et al. (1999). The model includes Met Office Surface Exchange Scheme (MOSES), the land surface scheme developed by Cox et al. (1999), which includes the dependence of stomatal resistance, and therefore evapotranspiration, on CO$_2$.

The experiments performed are summarized in Table 1. They are designed to quantify the separate influences of SST change and CO$_2$ change on enhanced land warming. The direct response to the CO$_2$ change has two components, one associated with purely radiative effects and
the other associated with the reduction in stomatal resistance. The reduction in stomatal resistance inhibits transpiration and thus surface evaporative cooling. To assess the relative role of these two components on land surface warming in response to the CO$_2$ change, an additional experiment has been performed in which any change in stomatal resistance is prevented.

For the control experiment, we use climatological monthly mean SST and sea ice concentrations derived from Hadley Centre Sea Ice and Sea Surface Temperature data set (HadISST) observations (Rayner et al. 2003) for 1961–90 and CO$_2$ at the average concentration for 1961–90. In perturbed CO$_2$ experiments, the CO$_2$ concentration is doubled. For the perturbed SST experiment, monthly mean SST anomalies are added to the climatological fields used for the control experiment. The SST anomalies were derived from an experiment with the Hadley Centre Coupled Model, version 3 (HadCM3), which comprises HadAM3 coupled to an ocean GCM, in which CO$_2$ increased at 1% year$^{-1}$ from the preindustrial CO$_2$ concentration, reaching 4 times its preindustrial value at year 140. To derive anomalies appropriate to a doubling of CO$_2$, the difference between two 30-yr means: (i) the mean for the period with CO$_2$ values near present day and (ii) the mean for period with CO$_2$ values near 2 $\times$ present-day CO$_2$. The sea ice edge is taken from the latter 30-yr mean.

To separate externally forced variability from internal variability, an ensemble of six integrations—each 1 month long, starting from 1 December—was performed for each experiment. For each experiment, the six integrations differed only in their initial conditions. The same six sets of initial conditions were used for each experiment and were taken from the end of a spinup integration, which was forced with climatological SST. The response to the forcing anomalies is estimated as the difference between the ensemble means of a pair of experiments. Using the ensemble standard deviations, we have done $t$ tests for the statistical significance of the changes and will describe only those changes that are significant at the 95% level; others changes apparent in the figures are not significant.

The figures and numerical results of this paper are mostly based on the December experiments. For comparison, we have also carried out a similar ensemble of integrations with initial conditions for 1 June. Although quantitatively the results are somewhat seasonally dependent, they are qualitatively the same. We point out in the text the few cases of substantial differences.

In this paper, we focus on the transient adjustment processes during the first month after the forcings are introduced. We compare the responses during the first month with the “equilibrium” responses for the same forcings, by which we mean the results of allowing the atmosphere and surface to reach equilibrium with the various forcings. We evaluate the equilibrium response to each forcing using an ensemble of two integrations, each of 25 yr, from which a mean over years 6–25 is evaluated.

### 3. Transient response of global mean surface air temperature

Figure 1a shows daily surface air temperature changes over land in response to CO$_2$ and SST change for December initial conditions. In response to doubling CO$_2$ without change in stomatal resistance, land surface air temperature warms by about 0.1°C in the first day and then rises more slowly to reach 0.4°C at day 10, which is statistically consistent with the December equilibrium response (shown by the symbols at the far right $y$ axis of Fig. 1), which will be further discussed in a future study. The time evolution of the transient response in
the experiment with change in stomatal resistance is qualitatively similar but has a larger magnitude of warming (Fig. 1a). The warming is about 0.2°C after day 1 and 0.6°C at day 10, again statistically consistent with the December equilibrium response. As time goes on, the ensemble members diverge, as shown by the increasing spread within the control ensemble (Fig. 1a). The signal of climate change is the difference between pairs of parallel experiments; it therefore has an uncertainty which is larger by a factor of $\sqrt{2}$ once the internal variability has become decorrelated. Compared with this uncertainty, the fluctuations that appear toward the end of the month are not statistically significant.

In response to the imposed SST change, the rising trend in land surface air temperature is initially slower than in response to CO$_2$ forcing but is sustained for longer; the temperature rises approximately linearly with time for the first 10 days. After about 15 days, it

FIG. 1. The ensemble mean changes in global mean surface air temperature (°C) (a) over land and (b) over sea, and (c) the land–sea warming difference in the sensitivity experiments relative to the control in December. The experiment acronyms are explained in Table 1. The thin lines in (a) and (b) show the $N^{-1/2}$ of internal standard deviation in the control experiment with the number of integrations $N = 6$. The symbols at the far right y axis indicate global mean surface air temperature changes in December for the equilibrium integrations. (d)–(f) The results in June.
exceeds the (imposed) warming over sea (1.8°C). By day 20, it reaches a magnitude of 2.3°C, being statistically consistent with the equilibrium response. As in the CO2 experiments, the decrease of warming after day 22 is due to internal atmospheric variability, predominantly associated with large fluctuations of surface air temperature over mid and high latitudes in both hemispheres (not shown). The difference between the CO2 and SST experiments is large in the latter part of the month and clearly statistically significant.

The land surface air temperature response to CO2 and SST changes together during the first 15 days (Fig. 1a) is nearly identical to the sum of the separate responses to CO2 and SST changes, indicating that the influences of SST and CO2 changes add linearly during this period. After day 16, some differences arise, which are associated with the internal atmospheric variability and are not statistically significant. The equilibrium warming under the combined forcing (3.1°C) is close to the sum of the separate responses (2.4°C + 0.6°C) and is also very close to the warming diagnosed from the coupled simulation (3.2°C). These results indicate that enhanced warming over land in CO2-forced experiments in the coupled model is mostly due to SST change (80%) rather than to the direct response to local CO2 forcing (Joshi et al. 2008; Compo and Sardeshmukh 2008; Dommenget 2009).

Surface air temperature changes over sea in December for the various experiments are illustrated in Fig. 1b. In response to CO2 change only, sea surface air temperature change is small as expected because of the constraint of no SST change. In response to SST change, Fig. 1b shows that surface air temperature change over sea reaches the equilibrium magnitude (1.8°C) in less than 5 days, indicating strong coupling between the sea surface and the atmospheric boundary layer.

Time evolutions of land–sea warming differences in December for the various experiments are illustrated in Fig. 1c. The land–sea warming differences in response to doubling CO2 with or without change in stomatal resistance match the land surface warming due to the constraint of no SST change. The land–sea warming difference in response to SST change is negative initially, increases with time, and becomes positive after day 16 when land surface warming exceeds the warming over sea. SST forcing only thus gives a land–sea warming ratio of greater than 1 after day 16 and a warming ratio of 1.3 in the equilibrium response.

Figures 1d–f show surface air temperature changes from the ensemble with June initial conditions, corresponding to Figs. 1a–c, respectively. The results are very similar over sea, and over land they agree that the warming is initially slower but eventually larger for SST forcing than for CO2 forcing. However, there are two main differences. First, the warming over land in response to doubling CO2 with a change in stomatal resistance is about 50% stronger than in December, reaching 0.6 K by the end of the month, still slightly below its equilibrium value. Second, the warming over land in response to SST change is slower in June than in December. By the end of the month, it has just equaled the warming over ocean, about three-quarters of its equilibrium value, which itself is a little smaller than in December. The slow response might be associated with weaker atmospheric circulation in the Northern Hemisphere summer than in winter. The equilibrium warming under the combined forcing (3.3°C) is equal to the sum of the separate responses (2.1°C + 1.1°C) and is very close to the warming diagnosed from the coupled simulation (3.3°C), with the SST forcing explaining 64% of the land warming signal. The seasonality of the equilibrium land surface warming in response to different forcings will be investigated in detail in a future study.

4. Understanding the transient adjustment in response to CO2 change

What processes are responsible for the transient evolution of surface air temperature changes over land and sea in response to different forcings, as seen in the previous section? In this section, we examine the transient evolution in response to CO2 change of the land and sea surface energy budget (Fig. 2), cloud feedback (Fig. 3), hydrological cycle (Fig. 4), vertical profiles of temperature and relative humidity (Figs. 5 and 6), and the geographical patterns of changes in surface air temperature, 850-hPa temperature, and surface evaporation (Figs. 7 and 8). The major features of these changes and the time scales associated with them are summarized in Table 2.

a. Response over land to double CO2 without change in stomatal resistance

In the double CO2 experiment without change in stomatal resistance, the dominant influence is an increase in net downward surface longwave radiation (∼1.0 W m⁻²; Fig. 2a and Table 2), due to the fact that the increased CO2 traps more outgoing longwave radiation (i.e., the greenhouse effect). The change in cloud cover over land is generally small during the first 15 days (Fig. 3a). It is, therefore, the change in clear-sky longwave radiation that is responsible for the immediate onset and plays the leading role in the land surface warming. Both upwelling and downwelling longwave radiation increase as the surface and atmosphere warm up (not shown), so the net remains roughly constant.
FIG. 2. The time evolutions of the global averaged surface energy anomalies (W m$^{-2}$) between the sensitivity and control experiments (left) over land and (right) over sea. Positive values indicate warming the surface.
From the outset, warming develops near the surface and cooling in the stratosphere (Fig. 5a). The latter effect is the well-known stratospheric adjustment via increased infrared emission (e.g., Shine et al. 2003); the stratosphere is not tightly coupled to the troposphere and its temperature evolution is not discussed further. As time passes, the warming amplifies and extends upward but remains greatest near the surface, corresponding to an

**Fig. 3.** The time evolutions of the global averaged total cloud cover (fraction) and CRF anomalies (W m$^{-2}$) between the sensitivity and control experiments (left) over land and (right) over sea. Positive values of CRF indicate warming at the surface.
increase of lapse rate throughout the troposphere. The relative humidity change is small (Fig. 6a).

As time passes, evaporation increases somewhat in response to temperature increase, but precipitation increases by slightly more than evaporation (not shown) as a result of weakly enhanced moisture transport from sea to land (not shown). This gives a positive change in precipitation minus evaporation during days 5–15 (Fig. 4a) and an increase of soil moisture (Fig. 4c).

b. Response over land to double CO₂ with change in stomatal resistance

In the double CO₂ experiment with changing stomatal resistance, the heat flux changes are rather different. There is a positive anomaly in net surface downward longwave on day 1, but it is much smaller than in the case with fixed resistance (Fig. 2c and Table 2). After day 1, the upwelling surface longwave radiation increases more rapidly than the downwelling (not shown separately), leading to a net longwave cooling anomaly during days 2–10. In this experiment, from day 2 onward a substantial part of the net heating comes from the increased net downward surface shortwave radiation (Figs. 2c and 2i; ~1 W m⁻²). This is explained as an increase in shortwave cloud radiative forcing (CRF; Fig. 3c), related to a cloud cover decrease of ~1.0% from days 2 to 15 (Fig. 3a), mainly associated with low cloud. The opposing decrease in longwave CRF (Fig. 3e) explains why downwelling surface longwave does not increase so much in this experiment as in the one with fixed stomatal resistance, in which the CO₂ greenhouse effect is the same but the longwave CRF change is less negative.

The changes in shortwave radiation and cloud are linked to reduced evaporation (Figs. 8i–k) and a decrease of relative humidity in the boundary layer (Fig. 6c). In contrast to the small increase in upward latent heat flux (evaporative cooling) with fixed stomatal resistance, from day 1 there is a large decrease of upward latent heat flux (~2 W m⁻²; Fig. 2e), that is, an anomalous latent heat flux that acts to warm the surface. The reduction of evaporation is caused by the reduced stomatal resistance in response to higher CO₂; this leads to a large increase in the Bowen ratio, meaning that evaporation becomes a less effective way of removing the heat from the land surface (Manabe and Wetherald 1975). The decrease in evaporation leads also to a larger increase of precipitation minus evaporation (Fig. 4a) than in the double CO₂ experiment without change in stomatal resistance and hence to a larger increase of soil moisture (Fig. 4c) too.

The increased shortwave heating and reduction in upward latent heat flux are opposed by an enhanced upward sensible heat flux (Fig. 2g). Nonetheless, the net downward heat flux is somewhat greater than in the double CO₂ experiment without change in stomatal resistance, so the warming is greater both at the surface and in the troposphere but has a similar time evolution.
FIG. 5. The time evolutions of the ensemble mean global temperature change (°C) with height between the sensitivity and control experiments (left) over land and (right) over sea. Note the corresponding equilibrium responses in December are also plotted in dark red lines (EQ). Note that the scales of x axis are not the same for all experiments.
FIG. 6. The time evolutions of the ensemble mean global relative humidity change (%) with height between the sensitivity and control experiments (left) over land and (right) over sea. Note the corresponding equilibrium responses in December are also plotted in dark red lines (EQ).
The surface air temperature declines during the last 10 days of the month (Fig. 1a), despite the anomalous net heat flux being positive throughout the month; the net heat input is going to the lower soil layers, which continue to warm (not shown).

The greater warming in June (Fig. 1d) than in December (Fig. 1a) is associated with a larger anomalous net heat flux in June. In June the reduction in evaporation and latent heat flux are twice as large (4 W m\(^{-2}\)) as in December, presumably because a large fraction of the global land and therefore its vegetation are located in the Northern Hemisphere. Both leaf area and evaporation are greater in summer, so stomatal closure is more influential. The reduction in cloud is consequently greater in June, and because there is also greater insolation, the increases in shortwave CRF and net downward shortwave are 4 times larger than in December.

c. Response over sea to double CO\(_2\)

Over sea, surface warming is prevented in the CO\(_2\) experiments, and change in stomatal resistance has no direct impact, so the evolution of the surface energy budget is essentially the same in the two cases. Greenhouse warming in the longwave flux is present from day 1 (~1.0 W m\(^{-2}\); Fig. 2b and Table 2), and the net downward heat flux is enhanced by reductions in upward latent heat flux associated with reduced evaporation (see next paragraph) as time passes (Figs. 2d,f). The troposphere near the surface warms little because of the surface constraint, but greenhouse warming occurs above the surface, with a maximum around 800 hPa, producing a decrease of lapse rate in lower levels (i.e., an increase in stability; Fig. 5), which presumably suppresses convective mixing. There is a decrease in cloud cover at all
levels (not shown), giving a negative change in longwave CRF and a positive in shortwave CRF (Figs. 3f,d), but the change in net surface shortwave radiation is smaller (Fig. 2d), presumably because of increased absorption by CO₂ and water vapor in the boundary layer. Unlike over land, the surface energy budget perturbation does not tend to return to zero but fluctuates at around $1.3 \pm 0.2$ W m$^{-2}$ (Fig. 2j). Heat is not being conserved because SST is fixed, and the ocean cannot warm up in response to the net surface heat flux. CO₂ change leads to a decrease of evaporation (because the boundary layer becomes warmer and moister but the SST does not change) and hence to a small decrease of relative humidity in the troposphere (Fig. 6b), but not at the surface, where the air is always saturated.

d. Geographical patterns of response to double CO₂

In the experiment without change in stomatal resistance (Fig. 7), the surface warming first develops over North Africa and central Asia where soils are dry, and over Antarctic and Greenland where the atmosphere is very stable, because in these regions the atmosphere transmits more longwave radiation from the surface, so the enhancement in the clear-sky greenhouse effect at the surface is stronger, as illustrated by large increase in surface longwave radiation in these regions, while the change in latent heat flux is weak initially (not shown). As time passes, these warming anomalies amplify and propagate upward, with 850-hPa temperature anomalies being smaller than those at the surface. Meanwhile, the warming in the lower troposphere extends to adjacent seas through advection. As the land surface warms, anomalous low pressure develops over land and high pressure develops over sea (not shown).

In the experiment with change in stomatal resistance (Fig. 8), the surface warming initially develops over land regions (Fig. 8a) where evaporation decreases (Fig. 8i) with a pattern correlation of $-0.64$ between the two
<table>
<thead>
<tr>
<th></th>
<th>Over land: CO₂ only with fixed SR</th>
<th>Over land: CO₂ only with variable SR</th>
<th>Over sea: CO₂ only</th>
<th>Over land: SSTA</th>
<th>Over sea: SSTA</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Surface air temperature (°C)</strong></td>
<td>+0.1 (immediate)</td>
<td>+0.2 (immediate)</td>
<td>~0 (2–3 days)</td>
<td>Increases 2.5 (10–20 days)</td>
<td>+1.5 (2–3 days)</td>
</tr>
<tr>
<td><strong>Net longwave (down: W m⁻²)</strong></td>
<td>+1 (immediate) roughly constant</td>
<td>Small − (2–3 days)</td>
<td>+1 (immediate)</td>
<td>Small − (2–3 days)</td>
<td>−5 (immediate)</td>
</tr>
<tr>
<td><strong>Net shortwave (down: W m⁻²)</strong></td>
<td>Small − (2–3 days)</td>
<td>+1 (2–3 days)</td>
<td>Small + (2–3 days)</td>
<td>+2.5 (2–3 days)</td>
<td>+3 (2–3 days)</td>
</tr>
<tr>
<td><strong>Latent heat (down: W m⁻²)</strong></td>
<td>Small − (2–3 days)</td>
<td>+2 (immediate)</td>
<td>Small + (2–3 days)</td>
<td>Small − (2–3 days)</td>
<td>−30 (immediate) declines</td>
</tr>
<tr>
<td><strong>Sensible heat (down: W m⁻²)</strong></td>
<td>Small − (2–3 days)</td>
<td>−2 (immediate)</td>
<td>Small + (2–3 days)</td>
<td>Small − (2–3 days)</td>
<td>−8 (immediate) declines</td>
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<td><strong>Cloud (%)</strong></td>
<td>−0 (2–3 days)</td>
<td>−1 mainly low (2–3 days)</td>
<td>Small − (2–3 days)</td>
<td>−1.5 (2–3 days)</td>
<td>Small−low small + high (2–3 days)</td>
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<td><strong>Shortwave CRF (W m⁻²)</strong></td>
<td>Small + (2–3 days)</td>
<td>+1 (2–3 days)</td>
<td>+1 (2–3 days)</td>
<td>+2.5 (2–3 days)</td>
<td>+2.5 (2–3 days)</td>
</tr>
<tr>
<td><strong>Longwave CRF (W m⁻²)</strong></td>
<td>−0.5 (2–3 days)</td>
<td>−1 (2–3 days)</td>
<td>−1 (2–3 days)</td>
<td>+0.5 (2–3 days)</td>
<td>−1.0 (2–3 days)</td>
</tr>
<tr>
<td><strong>Precipitation (mm day⁻¹)</strong></td>
<td>~0</td>
<td>~0</td>
<td>Small − (2–3 days)</td>
<td>−0.2 (2–3 days)</td>
<td>+0.8 (immediate) declines</td>
</tr>
<tr>
<td><strong>Evaporation (mm day⁻¹)</strong></td>
<td>~0</td>
<td>−0.1 (immediate)</td>
<td>Small − (2–3 days)</td>
<td>Small + (2–3 days)</td>
<td>+1.4 (immediate) declines</td>
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<tr>
<td><strong>P − E (mm day⁻¹)</strong></td>
<td>Small + (10–20 days)</td>
<td>Small + (2–3 days)</td>
<td>Small − (2–3 days)</td>
<td>−0.2 (2–3 days)</td>
<td>−0.6 (immediate) declines rapidly</td>
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<tr>
<td><strong>Soil moisture content (kg m⁻²)</strong></td>
<td>Increases slowly</td>
<td>Increases slowly</td>
<td>NA</td>
<td>Decrease slowly</td>
<td>NA</td>
</tr>
<tr>
<td><strong>Temperature (as a function of height)</strong></td>
<td>Warming starts at surface, spreads upward, most at surface</td>
<td>Warming starts at surface, spreads upward, most at surface</td>
<td>Warming in mid troposphere</td>
<td>Similar warming at all levels, more near surface and tropopause</td>
<td>Initially most at surface, later most at tropopause</td>
</tr>
<tr>
<td><strong>Relative humidity (%)</strong></td>
<td>~0 (immediate)</td>
<td>~1 near surface</td>
<td>~0 near surface, small − above (10–20 days)</td>
<td>Small − near surface (immediate) declines</td>
<td></td>
</tr>
</tbody>
</table>
figures, as well as over the dry regions where it appears first in the previous experiment. As in that experiment, with the passage of time the anomalies amplify, propagate upward, and are advected over the sea. Because of the large local feedbacks associated with the reduction in evaporation and changes in hydrology and cloud, the warming over land surface is greater than in the experiment without change in stomatal resistance.

5. Understanding the transient adjustment in response to imposed SST change

In this section, we examine in a similar way the transient evolution over land and sea (Figs. 2–6) in response to imposed SST change, together with its geographical patterns (Figs. 9 and 10), as summarized in Table 2.

5a. Response over sea to imposed SST change

In the experiment with imposed SST change, from day 1 the change in net surface heat flux over the sea is dominated by a very large increase in evaporation (1.4 mm day$^{-1}$; Fig. 10c and Table 2) and hence latent heat flux (over 30 W m$^{-2}$; Fig. 2f). There are also substantial increases in net upward longwave flux (~5.0 W m$^{-2}$; Fig. 2d) and upward sensible heat flux (~8.0 W m$^{-2}$; Fig. 2h). On day 1 the net downward surface shortwave flux is reduced as well (by ~1.0 W m$^{-2}$; Fig. 2d), but this anomaly becomes positive on day 2 (reaching 3.0 W m$^{-2}$ on day 3).

Through these upward fluxes, the imposed SST change leads to a rapid warming of the boundary layer (Fig. 5f); the warming near the surface at day 5 is already very close to the equilibrium response. The warming of
boundary layer leads to a decrease of tropospheric stability; this and the enhanced moisture availability enhance deep convection, leading to an increase of high cloud and a decrease in low and medium cloud. The changes of shortwave and longwave fluxes are closely related to these cloud changes (Fig. 3). On day 1, low cloud cover decreases by 1.5% while high cloud cover increases by 1% (not shown), and the changes are greater on days 2 and 3. The shortwave CRF anomaly is generally positive and is responsible for the net shortwave anomaly but has large fluctuations. The longwave CRF anomaly is fairly constant and negative.

The immediate large increase in evaporation leads to an increase in precipitation (0.8 mm day$^{-1}$) on day 1; the magnitude of both anomalies gradually decreases (not shown). The initial increase of evaporation is greater than the initial increase of precipitation (Fig. 4b), consistent with the increasing moisture content of the warming atmosphere (not shown).

Relative humidity is initially reduced by about 3% near the surface (Fig. 6f) as a direct effect of a warming without a commensurate increase in specific humidity, but subsequently it mostly recovers because of the enhanced evaporation. Warming spreads upward from the surface (Fig. 5f) as a new moist adiabatic profile is established, with a reduced lapse rate and the greatest warming eventually emerging in the upper troposphere. The magnitude of the anomalous upward sensible heat flux reduces gradually. The net surface longwave anomaly changes sign after 10 days as the troposphere gets warmer and builds up to a substantial net downward flux (3 W m$^{-2}$; Fig. 2b). The net downward shortwave flux anomaly declines due increasing cloud and absorption by water vapor and changes sign after about 15 days.

FIG. 10. The time evolutions of precipitation (mm day$^{-1}$), evaporation (mm day$^{-1}$), and sea level pressure (hPa) anomalies in response to SST change. (bottom row) The corresponding equilibrium responses in December. Note that contour intervals are different from Figs. 7 and 8.
The net upward heat flux anomaly is initially very large and reduces with time, but not to zero because the fixed SST does not permit the surface to cool, and heat is not conserved.

b. Response over land to imposed SST change

In contrast to the results in the double CO₂ experiments, there are no large changes on day 1; there is no immediate response because there is no forcing directly over land. In this case, the warming in the troposphere does not originate at the surface, being nearly uniform in the vertical during the first few days (Fig. 5e). From day 10 onward, warming is more pronounced in the upper troposphere and at the surface. The former means a reduction in lapse rate in the mid to upper troposphere, which is weaker than the reduction in lapse rate over the sea (cf. Figs. 5e and 5f at equilibrium; Joshi et al. 2008).

The pronounced warming near the surface suggests an influence of land surface feedbacks. By day 2–3, there is a negative anomaly of about 0.2 mm day⁻¹ in precipitation minus evaporation (Fig. 4a), associated with reduced moisture transport from sea to land (not shown) and weakly enhanced evaporation (not shown), consistent with a decrease in soil moisture content (Fig. 4c). The weak change in evaporation is consistent with a decrease in total cloud cover (by 1.5%; Fig. 3a) and reductions in relative humidity (Fig. 6e) at all levels, which are amplified in the upper troposphere as time passes and are greater than over the sea.

The reduction in cloud leads to a positive shortwave CRF anomaly (Fig. 3c), and it is anomalous net shortwave radiation that initiates the warming (Fig. 2a) at the surface. It increases over days 1–5 (reaching ~2.5 W m⁻²) and then remains roughly constant. The surface shortwave anomaly is partly offset by upward longwave, sensible, and latent heat flux anomalies (Figs. 2a,e,f). As in the CO₂ case, the net heat input to the land continues throughout the month, with the lower soil layers warming even after the surface air temperature has attained its maximum anomaly.

c. Response to both CO₂ and SST changes

In response to SST and CO₂ changes together, the time evolution of the surface energy budget, hydrology cycle, cloud feedback, and vertical profiles of temperature and relative humidity is very similar to the time evolution of the sum of the separate responses, as was the case for the global land surface air temperature responses seen in Fig. 1, indicating that the responses combine linearly to the imposed CO₂ change and SST change. Consequently, the evolution of temperature over land from days 1 to 3 is dominated by direct CO₂ change (rapid warming in boundary layer and surface), whereas from day 5 onward it is the SST change that dominates both over land and sea.

Over the sea, as over the land, the net heat flux anomaly returns to zero when both CO₂ and SST forcings are included, although it does not with either separately. This is not directly because of energy conservation, since SST is prescribed with both forcings, but rather because the SST anomalies being used have themselves been diagnosed from the coupled experiment with CO₂ forcing, in which heat was conserved.

d. Geographical patterns of response to imposed SST change

Surface air temperature over the sea quickly approaches its equilibrium warming (Fig. 9). In the lower troposphere (850 hPa in Fig. 9), the pattern of warming is similar to the surface, indicating a local thermodynamic coupling. Surface and lower troposphere warming is advected inland from the coasts, first in South America and central Africa, and a decrease in precipitation (Fig. 10) is associated with the warming. The pattern of upper tropospheric temperature anomalies (300 hPa in Fig. 9) is not as strongly tied to pattern of the surface temperature anomalies. Instead, the warming initially develops over the tropics (Fig. 9), which can be attributed to enhanced deep convection, and is advected more quickly over land than in the lower troposphere because the tropical atmosphere cannot maintain strong temperature gradients, so the large-scale dynamics act to advect the heating across the entire tropical belt (Sobel et al. 2002; Chiang and Lintner 2005; Joshi et al. 2008). Another distinct feature of equilibrium land surface warming in response to imposed SST change is the warming over all land areas while there is weak regional cooling in some land areas in response to CO₂ change both without and with change in stomatal resistance (Figs. 7d and 8d).

Increased evaporation occurs immediately over almost the entire ocean (Fig. 10). At time passes, the magnitude of the evaporation anomalies decreases over the sea because the specific humidity increases (not shown). However, precipitation anomalies exhibit both enhancement and reduction over sea; the local response tends to depend not only on the SST anomaly but also on the control SST, since precipitation tends to be organized in regions of the highest absolute SST in the tropics.

In the first few days, anomalous low pressure develops over sea and high pressure anomalies develop over land (Fig. 10) because the sea is warmer. This large-scale circulation change is consistent with reduced moisture transport from sea to land (not shown).
6. Summary and conclusions

During the last 20–25 yr, observed sea surface temperature and land air temperature have shown a warming trend, with the warming over land being greater than over sea (Kumar et al. 2004; Sutton et al. 2007; Compo and Sardeshmukh 2008). Sutton et al. (2007) showed that enhanced warming over land is a robust feature of climate model responses to increasing CO₂ using the World Climate Research Programme’s Coupled Model Intercomparison Project phase 3 (CMIP3) results. In this paper, we separate the contributions to the enhanced warming over land from CO₂ change and imposed SST change. Even without a change in CO₂, the warming over land is greater than over sea when SST is raised (Joshi et al. 2008; Dommenget 2009), and our experiments show that the land–sea warming contrast in response to elevated CO₂ is due mainly to this SST-forced contribution, with the SST forcing explaining 80% and 64% of the land warming signal in December and June, respectively. Land surface warming in response to imposed SST change is the warming over all land areas, whereas there is weak regional cooling in some land areas in response to CO₂ change. In a future study, we will analyze the equilibrium response in detail by focusing on understanding the spatial pattern of the response and its seasonal evolution. Here, we investigate the transient adjustment processes of the surface and troposphere when these forcings are turned on separately and suddenly. The adjustment processes can be grouped into three stages: immediate response of the troposphere and surface fluxes (day 1), fast adjustment of surface processes (days 2–5), and adjustment of the whole troposphere (days 6–20).

1) Immediate response of the troposphere and surface fluxes: Owing to the greenhouse effect, increase in CO₂ with or without change in stomatal resistance leads to the increase of downward longwave radiation at the land surface, which causes land surface warming. With change in stomatal resistance, the increase in CO₂ also leads to a decrease in land surface evaporation, which enhances the land surface warming. (Consequently, the enhanced upward longwave radiation offsets the surface longwave greenhouse effect). The troposphere over the sea (above the surface) shows a weak greenhouse warming. In response to imposed SST change, the warming over land on day 1 is small. Over the sea, there are very large immediate changes in upward heat fluxes, enhanced evaporation and precipitation, and very rapid lower boundary layer warming.

2) Fast adjustment of surface processes, days 2–5: In the double CO₂ experiment with change in stomatal resistance, rapid surface warming is enhanced by a net increase in surface shortwave radiation, associated with reduction of cloud, evaporation, and lower tropospheric relative humidity. In both CO₂ experiments, the warming over land spreads upward from the surface, and by day 5 the surface and tropospheric temperature response is statistically consistent with the equilibrium value. Likewise, there is rapid warming and moistening in the troposphere over the sea, spreading upward from the surface in response to imposed SST change. By contrast, land surface warming in response to imposed SST change is slower, and tropospheric warming occurs at all levels together rather than spreading upward from the surface. This warming is associated with a reduction in precipitation, relative humidity, and cloud cover, and hence with enhanced surface shortwave radiation.

3) Adjustment of whole troposphere, days 6–20: Further adjustments over land and ocean in both double CO₂ experiments from day 6 onwards are quite small. By contrast, in response to imposed SST change, amplification of warming continues in the troposphere over both land and ocean, with a maximum warming developing in the upper troposphere. This upper tropospheric warming is the result of the reduction of the saturated adiabatic lapse rate as temperature rises (Joshi et al. 2008) and is always observed in GCM global warming experiments (see, e.g., Meehl et al. 2007). The whole tropospheric response in about 20 days is statistically consistent with the equilibrium response. This time scale is presumably the time required for the global humidity distribution to equilibrate following the change in forcing.

Our CO₂ experiments confirm the findings of recent authors (Gregory and Webb 2008; Andrews and Forster 2008; Lambert and Webb 2008; Andrews et al. 2009) that there are significant tropospheric adjustments to CO₂ increase that involve changes in clouds and perturbations to surface fluxes (see Fig. 5 of Gregory and Webb), and their short time scale of only a few days supports their inclusion in the effective radiative forcing. They are analogous to radiative forcing by the aerosol indirect and semidirect effects. In our experiments, the adjustment of the troposphere to SST forcing is also rapid, but in the real world, or a climate model including the ocean, the SST change is not itself a forcing; rather, SST change is part of the response to forcing of the climate system, which develops on decadal time scales at a rate determined by ocean heat uptake. In the real world, therefore, we would expect
the CO2-driven and SST-driven parts of the land–sea warming contrast to emerge on different time scales; the former is practically simultaneous with CO2 increase, whereas the latter will lag behind by years. Although our separation of the effects is idealized, it has practical relevance to time-dependent climate change.

Both CO2 and SST forcing contribute to the enhanced warming over land. First, with CO2 forcing, there is warming over land (which is obviously more than over the sea if SSTs do not change) due to the greenhouse effect. Second, the CO2-induced warming in the case with change in stomatal resistance is amplified both by reduced evaporation and by increased net downward shortwave due to cloud changes. Third, with SST forcing, the surface enhancement of warming again involves increased net downward shortwave due to cloud changes. The second and third cases are thus similar with regard to the role of clouds and shortwave radiation, but the causes are different. In the second case, the direct response of stomatal resistance to CO2 reduces the evaporation, and the reduced evaporation means that soil moisture increases (Fig. 4c). In the third case, evaporation increases rather than decreases; however, Joshi et al. (2008) argue that this increase is limited by moisture availability, as indicated by the reduction in soil moisture (Fig. 4c). In both cases, restricted evaporation is associated with reduced relative humidity (Figs. 6c,e).

With CO2 forcing, there is a decrease of precipitation over the ocean and in global mean precipitation, which is dominated by the sea. This is consistent with the increased stability and suppressed convection caused by CO2 warming in the troposphere over a fixed surface temperature (Figs. 5b,d). That is a process-based explanation, and the result can also be understood on energetic grounds (Allen and Ingram 2002; Yang et al. 2003; Lambert and Webb 2008; Andrews et al. 2009). Increased CO2 traps more outgoing longwave radiation and enhances the tropospheric radiative heating. The troposphere adjusts to a new state by reducing precipitation and hence condensational heating to compensate.

By contrast, higher SSTs cause precipitation to intensify (Yang et al. 2003), as is seen in our experiments with imposed SST change, in which there is an increase of precipitation over ocean and in global mean precipitation. Stronger condensational heating associated with precipitation warms the troposphere, which adjusts to a new state by the increase in radiative cooling. The extra radiated heat in this case is being supplied at the ocean surface, where the SSTs are prescribed, implying an unrestricted heat source.

The results in this study indicate that the reduction of cloud cover enhances land surface warming in response to both CO2 change and imposed SST change through its effect on surface shortwave radiation. The change in cloud, in turn, is closely related to the change in the hydrological cycle. Joshi et al. (2008) similarly show that the land–sea warming contrast is very sensitive to the way in which GCMs treat clouds. These sensitivities strengthen the conclusion that realistic representation of cloud cover and its radiative effect and the hydrological cycle are essential to the ability of numerical models to predict responses to anthropogenic climate forcing.

The results obtained in this study are obtained by using one model. It is likely that the time evolution of transient adjustment processes, as well as its amplitude and pattern, may vary to some extent with the model. Similar analysis of other models would be helpful to test whether the findings identified in this paper are robust features of the atmospheric response to double CO2 and SST changes.

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