Contrasting Regional Responses to Increasing Leaf-Level Atmospheric Carbon Dioxide over Australia

FAYE T. CRUZ AND ANDREW J. PITMAN
Climate Change Research Centre, University of New South Wales, Sydney, New South Wales, Australia

JOHN L. MCGREGOR
Centre for Australian Weather and Climate Research, and CSIRO Marine and Atmospheric Research, Aspendale, Victoria, Australia

JASON P. EVANS
Climate Change Research Centre, University of New South Wales, Sydney, New South Wales, Australia

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ABSTRACT

Using a coupled atmosphere–land surface model, simulations were conducted to characterize the regional climate changes that result from the response of stomates to increases in leaf-level carbon dioxide (CO₂) under differing conditions of moisture availability over Australia. Multiple realizations for multiple Januarys corresponding to dry and wet years were run, where only the leaf-level CO₂ was varied at 280, 375, 500, 650, 840, and 1000 ppmv and the atmospheric CO₂ was fixed at 375 ppmv. The results show the clear effect of increasing leaf-level CO₂ on the transpiration via the stomatal response, particularly when sufficient moisture is available. Statistically significant reductions in transpiration generally lead to a significantly warmer land surface with decreases in rainfall. Increases in CO₂ lead to increases in the magnitude and areal extent of the statistically significant mean changes in the surface climate. However, the results also show that the availability of moisture substantially affects the effect of increases in the leaf-level CO₂, particularly for a moisture-limited region. The physiological feedback can indirectly lead to more rainfall via changes in the low-level moisture convergence and vertical velocity, which result in a cooling simulated over Western Australia. The significant changes in the surface climate presented in the results suggest that it is still important to incorporate these feedbacks in future climate assessments and projections for Australia. The influence of moisture availability also indicates that the capacity of the physiological feedback to affect the future climate may be affected by uncertainties in rainfall projections, particularly for water-stressed regions such as Australia.

1. Introduction

Carbon dioxide (CO₂) directly affects the physiology of plants (Field et al. 1995; Ainsworth and Long 2005). Stomates, which regulate the fluxes of water vapor and CO₂ at the leaf surface, reduce their opening under an enriched CO₂ environment because they are able to take up CO₂ more efficiently. As a consequence of the reduced stomatal conductance, plants become more water use efficient by transpiring less, which in turn affects the surface energy and water balance. Modeling studies have examined the contribution of this physiological feedback in projections of the global climate. Simulations in which the stomatal resistance is doubled show that the suppression of transpiration leads to a global mean increase in temperature with effects on rainfall and soil moisture (Pollard and Thompson 1995; Henderson-Sellers et al. 1995; Martin et al. 1999). Inclusion of the physiological forcing also offsets the enhanced evapotranspiration that would have resulted if only the radiative effect of CO₂ was considered (Bounoua et al. 1999; Levis et al. 2000). Recent studies indicate an ongoing discussion on whether the physiological feedback to increased CO₂ leads to increases or decreases in runoff (Martin et al. 1999; Gedney et al. 2006; Piao et al. 2007; Betts et al. 2007; Huntington 2008).
Although the changes in the surface climate arising from the physiological response of vegetation are small compared to the radiative effect of doubled atmospheric CO$_2$ at the global scale, they are significant at regional scales, including the tropics (Sellers et al. 1996) and boreal forests (Pollard and Thompson 1995; Henderson-Sellers et al. 1995; Martin et al. 1999). There have been few studies that have explored the effect of the physiological feedback on the Australian climate, which is significant because in contrast to boreal and tropical systems, evaporation in Australia is typically water limited. It is possible that a large sensitivity would be found where moisture was available but limited. Australia provides an environment where this is regionally common under specific climate regimes. For example, below-average rainfall is associated with El Niño events, particularly over north and northwest Australia, rainfall declines during austral autumn are found to be associated with El Niño Modoki events (Taschetto and England 2009). Over north and northwest Australia, rainfall declines during austral autumn are found to be associated with El Niño Modoki events (Taschetto and England 2009).

Some studies of CO$_2$-stomate–hydrometeorological interactions have already been done. Aston (1984) predicted an increase in streamflow for a subcatchment in Australia as a consequence of doubled stomatal resistance. Narisma and Pitman (2004) also examined the effect of biospheric feedbacks on the changes in the Australian January climate due to land cover change. However, a systematic assessment of the effect on Australia of increased CO$_2$ on plant–soil–climate interactions in a climate model has yet to be reported.

This paper characterizes the changes in the surface climate over Australia that result from the response of the stomates to increases in leaf-level CO$_2$ under differing conditions of moisture availability. Using a coupled atmosphere–land model, we examine the changes in the means and mean diurnal cycles that result from the interaction between the limiting factors on the stomatal conductance in the land surface model. Consistent with the results from the literature, reductions in transpiration leading to a significantly warmer land surface are simulated in response to the increased leaf-level CO$_2$ during summer. However, strong cooling over Western Australia (WA) was also simulated when there was more moisture available in the environment, even though transpiration has decreased. This paper identifies and then examines the mechanisms involved in this unexpected response.

Our work focuses on the effect of changes in the CO$_2$ only at the leaf level. This isolates any additional contribution from the changes in radiative forcing due to atmospheric CO$_2$ changes. We explore the effect of increases in leaf-level CO$_2$ at various concentrations to determine the extent and magnitude of changes at each level of increase. We have not considered any changes to the structural feedbacks—such as changes in leaf area index (LAI), canopy height, and root depth—which have been shown to have an opposing effect to the physiological response of stomates to the increasing CO$_2$ (Betts et al. 1997; Levis et al. 2000; Piao et al. 2007; Betts et al. 2007; Calvet et al. 2008). We have also conducted simulations only for the month of January. Although this does not allow us to explore the effect of changing trends in the soil moisture that may occur in a longer simulation period, we were able to conduct multiple realizations to obtain statistically robust results. Our methodology is described in section 2. Section 3 presents the changes in the mean over Australia and the effect of the elevated CO$_2$ on the mean diurnal cycles over the Murray–Darling Basin (MDB) and southwest Western Australia (SWWA). Section 4 examines the anomalous cooling simulated over Western Australia in one of our experiments, followed by our discussion in section 5. Conclusions are presented in section 6.

2. Methodology

a. The global climate model

The Conformal-Cubic Atmospheric Model (CCAM) is the global climate model used in this study. Developed at the Commonwealth Scientific and Industrial Research Organisation (CSIRO), CCAM is a two-time-level semi-implicit hydrostatic variable-resolution climate model, which is described in detail in McGregor and Dix (2001, 2008). The dynamics in the model include a total variation diminishing (TVD) vertical advection, semi-Lagrangian horizontal advection with bicubic horizontal interpolation, an unstaggered grid with a reversibly staggering scheme for winds (McGregor 1993, 1996; McGregor and Dix 2001; McGregor et al. 2002; McGregor 2005). The model's physical parameterization schemes include McGregor (2003) cumulus convection, Geophysical Fluid Dynamics Laboratory (GFDL) parameterization for shortwave and longwave radiation (Schwarzkopf and Fels 1991), gravity wave drag, Louis (1979) stability-dependent PBL scheme with nonlocal vertical mixing (Holtslag and Boville 1993), and a soil–snow scheme (Kowalczyk et al. 1994).

CCAM has been extensively tested, evaluated, and applied for a variety of global and regional studies (McGregor et al. 2002; Hope et al. 2004; Nunez and McGregor 2007; Lal et al. 2008) and trace gas modeling (Law et al. 2006). This model has participated in model intercomparison projects, such as the Regional Climate Model Intercomparison Project (RMIP) for Asia (Fu et al. 2005), the Stretched-Grid Model Intercomparison Project (SGMIP) for Asia (Vaughan et al. 2007), and the Regional Modelling and Analysis (RMET) project for Australia (Cruz et al. 2007).

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Project (SGMIP; Fox-Rabinovitz et al. 2006), and the Global Land–Atmosphere Coupling Experiment (GLACE; Koster et al. 2006; Guo et al. 2006).

b. The land surface model

The CSIRO Atmosphere Biosphere Land Exchange (CABLE) land surface scheme is used to represent the terrestrial surface in CCAM (Kowalczyk and Wang 2004; Kowalczyk et al. 2006). It includes a representation of the canopy, soil and snow, carbon pool dynamics, and soil respiration (Kowalczyk et al. 2006). Both traditional and innovative measures of model performance have been used to test CABLE (Abramowitz 2005; Wang et al. 2007; Abramowitz et al. 2008). The coupled CCAM and CABLE model has also been employed following the protocol of phase I of the Coupled Carbon Cycle Climate Model Intercomparison Project (C4MIP; Law et al. 2006).

CABLE uses a one-layer two-big-leaf canopy model to solve the leaf temperature, photosynthesis, stomatal conductance, CO₂, and energy fluxes for sunlit and shaded leaves (Wang and Leuning 1998). It uses a radiation scheme that determines the photosynthetically active radiation, near-infrared, and thermal radiation absorbed by the sunlit and shaded leaves (Wang and Leuning 1998). CABLE includes a coupled model of the photosynthesis and stomatal conductance, where stomates are allowed to respond to changes in the atmospheric and soil environments (Leuning 1995; Wang and Leuning 1998). The stomatal conductance \( G_s \) is

\[
G_s = G_0 + \frac{a_1 f_w A_c}{(C_s - 1)[1 + (D_s/D_0)]}, \tag{1}
\]

where \( G_0 \) is the stomatal conductance when the net photosynthesis \( A_c \) is zero, \( C_s \) is the CO₂ concentration at the leaf surface, and \( D_s \) is the water vapor pressure deficit (VPD) at the leaf surface. The following are empirical parameters: \( a_1 \) is related to the intercellular and leaf-level CO₂ concentration, \( f_w \) describes the sensitivity of stomates to soil water availability, \( \Gamma \) is the CO₂ compensation point of leaf photosynthesis, and \( D_0 \) describes the sensitivity to VPD [Eq. (5) in Wang and Leuning 1998].

CABLE represents vegetation above the ground, which allows full radiative and aerodynamic interaction between the atmosphere, vegetation, and ground (Finkele et al. 2003; Kowalczyk et al. 2006). A plant turbulence model developed by Raupach et al. (1997) is used to solve the air temperature and humidity within the canopy. Soil temperature and soil moisture for six layers are calculated in a soil module, using the heat conduction equation and Richards equation for soil moisture (Kowalczyk et al. 2006). A simple carbon pool model developed by Dickinson et al. (1998) determines the allocation of carbon to leaves, wood, and roots; and fast and slow carbon pools. However, it was not used in this study because the monthly LAI was prescribed in our simulations.

c. Model configuration

For this study, CCAM has been configured to provide a quasi-uniform spatial resolution of about 210 km, with 18 sigma levels in the vertical. Maps of the vegetation type and cover, LAI, soil albedo and texture are obtained from the International Satellite Land Surface Climatology Project, Initiative II (ISLSCP II) data archive (Hall et al. 2005). However, some data were adapted to the classification used in CABLE and then interpolated to the model grid.

The boundary conditions prescribed in the model were based on the 1° resolution monthly sea surface temperature (SST) and sea ice concentration (SIC) observational dataset from the Atmospheric Model Intercomparison Project Phase II (AMIP II; Fiorino 2007). The SST and SIC boundary condition datasets were derived following Taylor et al. (2000). The midmonthly means used as boundary conditions were calculated to ensure that the observed monthly means are preserved when the monthly mean is determined from the daily values in the model output. These datasets are then interpolated to the CCAM grid.

d. Design of the numerical experiments

This study aims to isolate the changes in the land surface that result only from the stomatal response to CO₂ fertilization. Thus, only the level of CO₂ concentration at the leaf surface is changed in the parallel simulations using the coupled CCAM–CABLE model. In all simulations, a fixed atmospheric CO₂ concentration of 375 ppmv is specified in CCAM to remove any direct effect of the changes in CO₂ on the net radiation at the surface.

Six concentration levels of CO₂—280, 375, 500, 650, 840, and 1000 ppmv—have been specified in the land surface model for the experiment. These conditions are fixed during the entire simulation period. The control value, 280 ppmv, represents the preindustrial, whereas 375 ppmv represents the present day conditions. A CO₂ level of 500 ppmv may be reached by ~2050 under the emission scenarios used by the Intergovernmental Panel on Climate Change (IPCC; Meehl et al. 2007). However, the higher levels do not occur under a low future-emission scenario, such as B1. On the other hand,
650 ppmv is reached in about 2060, 840 ppmv in about 2080, and 1000 ppmv in about 2100, under a high emission future, such as A1F1. Stabilization scenarios up to 1000 ppmv have also been explored by the IPCC in the Fourth Assessment Report. The proposed emissions for the Fifth Assessment Report are framed as “representative concentration pathways” (RCPs; Moss et al. 2008). The RCP 4.5 reaches ~650 ppmv of CO₂-equivalent postequilibrium after 2100. In contrast, RCP 8.5 equilibrates above 1370 ppmv CO₂ equivalent around 2100. These suggest that our 1000 ppmv is midrange in terms of projected concentrations.

A global-scale AMIP II–type simulation was conducted earlier for the years 1979–2004, which was initialized with a spin-up period by running the first year repeatedly until the model reached equilibrium. From this simulation, we obtained the mean January rainfall over the 26 years, from which we derive the rainfall anomalies for each January over Australia. We have then selected the three driest and the three wettest Januaries, as determined by the rainfall over the Murray–Darling Basin (Fig. 1), which will constitute the dry case and the wet case, respectively. We have made a distinction between these two cases to determine how the changes in the land surface climate as a result of the stomatal response to the increasing CO₂ are affected by the amount of moisture available in the region of interest. The wet (dry) case is analogous to La Niña (El Niño) conditions over Australia.

We have conducted simulations only for January to represent the austral summer. Although this limits our study, it allows us to perform multiple realizations of this month from which we can obtain results that are statistically robust. A similar experimental design was used in Cruz et al. (2010). Fifty-one realizations were performed for each of the three Januarys selected for the dry case and for the wet case for each CO₂ concentration. Members of the ensemble vary by slight perturbations to the initial conditions. In each case the initial global soil temperature of the deepest soil layer, 4.6 m below the surface, was changed by −0.125° to 0.125°C with increments of 0.005°C. We changed the deep soil temperature by equal amounts mirrored about zero to avoid any bias introduced by this perturbation on the warming and cooling patterns in the results. The method used by the soil model to solve the soil temperature involves communication among layers at each time step. Thus, this perturbation within the deep soil layer affects the upper layers and the surface temperature enough to generate independent realizations for the ensemble.

Figure 2 shows the ensemble monthly-mean rainfall averaged over three different Januarys, where each January consists of 51 realizations, corresponding to the dry and the wet cases and the difference between them. High rainfall is simulated over the tropics and along the eastern coast of Australia during summer in both the dry and wet cases (Figs. 2a and 2b). In the dry case, it is particularly dry over South Australia (SA), Victoria (VIC), western New South Wales (NSW), and along the coast of Western Australia where rainfall is less than 1 mm day⁻¹ (Fig. 2a). The below-average rainfall over these areas is associated with El Niño episodes on the years chosen for this case, particularly 1983 and 2002. In the wet case, 2–5 mm day⁻¹ more rainfall occurs over northern Australia and along the coast of NSW and about 1 mm day⁻¹ more over central Australia and NSW (Fig. 2b). However, although the wet case is wetter on average, there is less rainfall over SWWA (Fig. 2c).

e. Statistical analysis

For each dry and wet case, the average of the monthly-mean transpiration, canopy temperature, and daily total rainfall is taken over all realizations. The difference of the ensemble monthly means from the 280-ppmv simulation (control run) is obtained for the simulations for the higher CO₂ levels. A two-sample F and t test had been conducted to determine the statistical significance of the mean differences. The type of standard t test used
was determined by the $F$ test, which checks if the two independent samples come from normal distributions with equal or unequal variance. Both tests are two-tailed, using a 95% significance level.

The percentage area of the statistically significant differences at the 95% level was obtained over the land points in Australia, which is our region of interest. This value is used to indicate the increase in the spatial extent of the significant changes with the increasing levels of CO$_2$. Findell et al. (2006) indicated that for a field to be considered statistically significant on the whole, taking into account spatial correlation within the field, more than 5% of the area of interest must pass the significance test at the 95% level locally. Citing Livezey and Chen (1983), the percentage of the field that should pass the local 95% significance test can be estimated corresponding to the number of spatial degrees of freedom of a field (Findell et al. 2006). We have not directly derived those estimates for this study. Instead, as a guide to interpretation of our results, we note that 6% of a field with 1000 degrees of freedom must pass the local 95% significance test to be considered statistically significant (Findell et al. 2006). A higher minimum value of the percentage area must pass the local 95% significance test if there are fewer degrees of freedom for a field (Findell et al. 2006).

The ensemble mean diurnal cycles for transpiration, canopy temperature, and rainfall are also obtained at each grid point. The changes in the mean diurnal cycles allow us to examine the effect of the increased CO$_2$ on the simulated quantities in more detail. The spatial average of these hourly values are derived over our regions of interest: the Murray–Darling Basin, defined here as 24$^\circ$–38$^\circ$S, 138$^\circ$–153$^\circ$E and southwest Western Australia, defined as 31$^\circ$–34$^\circ$S, 116$^\circ$–119$^\circ$E (Fig. 1). The MDB is important for Australia’s agricultural production, and it is also a major source of irrigation (available online at http://www.mdbc.gov.au). With the recent decline in rainfall as well as warmer and drier projections in the future, there have been significant efforts in understanding the climate changes in this region, mostly as part of the South Eastern Australian Climate Initiative (SEACI; Trewin 2006; Timbal and Murphy 2007; Maxino et al. 2008). The MDB is also a region used as a Continental Scale Experiment (CSE) in the Global Energy and Water Cycle Experiment (GEWEX) program (available online at http://www.gewex.org). There is considerable interest in SWWA, particularly investigating the sudden decline in winter rainfall in the mid-1970s (IOCl 2002; Pitman et al. 2004; England et al. 2006; Ummenhofer et al. 2008). These two areas are also representative of the changes simulated for the east and west regions of Australia.

![Graph](image-url)
3. Results

Although we have conducted global simulations, we will focus on the results over Australia. Changes in the daily mean transpiration, mean canopy temperature, and total rainfall have been averaged over 153 realizations, corresponding to the 51 realizations of each of the three Januarys in the dry and wet cases. Differences that are statistically significant at a confidence level of 95% are noted on the maps. The mean changes in the dry case will be first discussed, followed by a description of the results in the wet case. We then examine the changes in the mean diurnal cycle, particularly over the MDB and SWWA.

a. Mean changes in the dry case

For the dry case, there is almost no significant change in transpiration when CO₂ is initially increased from 280 to 375 ppmv, save for some areas along the east coast (Fig. 3a). Table 1 lists the percentage area of statistically significant differences at the 95% level for each level of CO₂ change. Changes occur only over 3.4% of the land area at 375 ppmv. However, this increases to 18.1% at 500 ppmv, which is larger than what we might expect by chance, as discussed in the previous section. Statistically significant decreases in transpiration of about 5 W m⁻² occur over the tropics and the eastern side of Australia (Fig. 3b). This reduction in transpiration is expected because the stomates reduce their opening when leaf-level CO₂ is increased. At 650 ppmv, changes over southwestern Australia become apparent (Fig. 3c). Thus, the area of reduced transpiration expands with further increases to CO₂. The maximum reductions in transpiration reach ~20 W m⁻² over forests along the east coast of NSW and Victoria (Figs. 3d and 3e). Because the patterns of change are coherent at various levels of CO₂, grow in magnitude and area as CO₂ is increased, and the results are statistically significant, this suggests that the changes in transpiration are not due to internal model variability.

No significant change in transpiration is simulated south of central Australia, which consists mostly of broadleaf shrubs with grass and bare soil (Fig. 1). The insufficient moisture available in this area (very low rainfall; Fig. 2a) limits the response of stomates to the elevated CO₂, which can be attributed to the representation of the stomatal conductance in the land surface model. Stomatal conductance is limited not just by increases in CO₂ but also by soil water availability [Eq. (1); Wang and Leuning 1998], and the soil moisture over this region is too low to support significant transpiration.

The effect of changes in transpiration can be seen in the changes in canopy temperature (Figs. 3f–3j). The areas of warming are coincident with the areas of reduced transpiration as expected, because of reduced evaporative cooling at the surface. Although there are some areas that have a lower canopy temperature when CO₂ is increased to 375 ppmv, these changes are not statistically significant (Fig. 3f). Some areas warm by up to 0.5°C in eastern Australia by 500 ppmv (Fig. 3g). As CO₂ is further increased, the regional warming of mostly 0.5–1°C extends toward the tropics and western Australia (Figs. 3h–3j). Temperatures increase by 2°C over the forested area in southeast Australia (Figs. 3i and 3j) where there was a significant reduction in transpiration of ~20 W m⁻² (Figs. 3d and 3e). The water stress in south-central Australia, as well as shallow-rooted vegetation, limits the effect of the elevated CO₂ on transpiration of this region. The coherent cooling observed here in Figs. 3g–3j is related to a decrease in net radiation over this area.

The reduced transpiration and warming could affect rainfall because there is less water vapor flux from the land surface. Figures 3k–3l show few significant changes in rainfall until 650 ppmv (Fig. 3m), at which level there is a decrease of 0.1–0.5 mm day⁻¹ over 8.6% of the land. By 1000 ppmv, 51.7% of the continent is significantly drier, coincident with surface warming (Fig. 3o). The scale of this continental reduction in rainfall is not enormous, but it is a drying on an already dry continent.

Finally, note the coherence between changes in transpiration, temperature, and rainfall. There is a clear emergence of a coherent and consistent pattern of change in each quantity through the various CO₂ levels.

b. Mean changes in the wet case

A stronger response to variations in CO₂ is observed on the vegetated surface in the wet case (Fig. 4), relative to the dry case (Fig. 3) because of increased moisture availability (Fig. 2c). With the initial increase from 280 to 375 ppmv, significant changes in the transpiration occur over 16.4% of the land (Fig. 4a), compared to 3.4% in the dry case (Table 1). At levels of CO₂ below 500 ppmv, transpiration is increased over northeast Australia (Figs. 4a and 4b), which can be attributed to increased atmospheric evaporative demand associated with decreased rainfall (Figs. 4k and 4l). However, further increases in CO₂, from 650 ppmv onward, limit the stomatal conductance and result in reducing transpiration by 1–5 W m⁻² in these areas (Figs. 4c–4e). Both the spatial extent and the magnitude of the transpiration reduction increase in response to the elevated levels of CO₂.

Above 650 ppmv, there is a strong similarity between the regional changes in transpiration between the wet (Figs. 4d and 4e) and dry cases (Figs. 3d and 3e). A
FIG. 3. Monthly average changes in (a)–(e) transpiration (W m$^{-2}$), (f)–(j) canopy temperature (°C), and (k)–(o) rainfall rate (mm day$^{-1}$) for changes in leaf-level CO$_2$: 375–280, 500–280, 650–280, 840–280, and 1000–280 ppmv. Averages are taken over three Januarys with 51 model realizations each for the dry case. Changes that are statistically significant at a 95% confidence level are marked with “1.”
significantly warmer surface, particularly over eastern Australia, results from the reduced evaporative cooling due to the decrease in stomatal conductance in a richer CO₂ environment (Figs. 4i and 4j). This is enhanced by a decrease in the soil evaporation (not shown). The area of warming grows as CO₂ is increased, although there appears to be minimal increase in the area where the temperature change is significant (Table 1). Changes in rainfall of up to ±0.5 mm day⁻¹ are also simulated at these higher levels of CO₂ (Figs. 4n and 4o).

Comparison of Figs. 3 and 4 show substantial differences between the detail of the wet and dry cases. However, the most confronting difference is the contrasting response of the surface climate to the increase in CO₂ over WA. The reduction in transpiration over this region indicates that the stomates have the same response to the elevated CO₂ levels as other regions. However, this area is cooler (Figs. 4g–4i) and wetter (Figs. 4l–4n) at levels of 500–840 ppmv, compared to the dry case (Figs. 3g–3i and 3l–3n). The temperature and rainfall contrast is even stronger between the western and eastern regions in the wet case (Figs. 4g–4l and 4o–4n).

This decrease in canopy temperature over WA (Figs. 4f–4j) can be attributed to a higher total latent heat flux in this region. Although there was a decrease in transpiration, the increase in soil evaporation, associated with more rainfall (Figs. 4k–4o) over this region, leads to an overall increase in the total latent heat flux (not shown). This leads to the question of what caused the increase in rainfall, accompanying the increases in leaf-level CO₂ over WA. The fundamentally different response over WA to reduced transpiration under elevated CO₂, in contrast to other parts of Australia, will be discussed in section 4.

c. Impact on the diurnal cycles over the MDB and SWWA

The effect of the stomatal response to increased CO₂ on the diurnal cycle of transpiration, canopy temperature, and rainfall is examined over the MDB and SWWA for the dry and wet cases. This is to further characterize the changes found in the mean quantities discussed in the previous sections. These two regions have been selected because they have been identified as key areas in understanding the changes in climate over Australia.

The MDB, situated in southeast Australia, consists of grass and shrubland in the west, crops in the south, and forest along the eastern coast (Fig. 1). Figure 5a shows a small effect on transpiration over the MDB in the dry case when CO₂ is increased from 280 to 375 ppmv. However, further increases in CO₂ clearly show reductions in transpiration, with maximum changes before midday. This results in a warmer canopy at all times during the day, with temperature increases reaching 0.7°C at 1000 ppmv. The diurnal pattern in rainfall indicates an effect on convective rainfall where peak reductions occur in the early afternoon.

The reduction in the transpiration rates are higher over the MDB in the wet case compared to the dry case. The increase in moisture availability enhances the suppression of transpiration where daytime reductions are about double the magnitude seen in the dry case (Fig. 5b). This leads to a large increase in the canopy temperatures, with maximum changes of 1.6°C during the day at 1000 ppmv. Nighttime temperatures are also warmer when CO₂ is increased. Rainfall is decreased at all times, with peak reductions from noon until afternoon, again suggesting that the effect on rainfall is via the convective rainfall.

The dry case over SWWA is wetter than the wet case based on the rainfall difference maps (see Fig. 2c). As such the transpiration reduction over the predominantly agricultural region (Fig. 1) during the day in the dry case (Fig. 6a) is similar to the changes observed over the MDB in the wet case. The loss in evaporative cooling leads to a warmer canopy during the day where the magnitude of the warming increases with CO₂, which is consistent with the decreases in transpiration. On the other hand, nighttime temperatures appear to be unaffected by the CO₂ change. Rainfall over this area also declines at higher levels of CO₂ with peak reductions late in the afternoon, again indicating a convective response.

Changes in the transpiration over SWWA only occur during the day in the wet case (Fig. 6b). Though we see a reduction in the daily mean transpiration in this area, the diurnal cycle indicates a strong reduction in the morning but an increase as time progresses toward the afternoon. Unlike the warming observed in the dry case, daytime canopy temperatures decrease as CO₂ increases. This cooling can be attributed to the increase in rainfall at higher CO₂ levels during the day (Fig. 6b). Note that the afternoon canopy temperature is cooler by about 0.5°C at
FIG. 4. As in Fig. 3, but for the wet case.
650 ppmv even though there is more rainfall at 1000 ppmv as a result of a higher transpiration at 650 ppmv, which enhances the cooling over the canopy.

The daytime increase in rainfall with the associated cooling over SWWA was evident in the mean changes over WA in the wet case as discussed earlier. Hence, the following section will present a proposed mechanism that explains the anomalous response of this region, in contrast to other parts of Australia, under elevated levels of CO₂.

4. Cooling anomaly over Western Australia

In this section, we examine the unexpected cooling simulated over WA in the wet case (section 3) and examine the mechanisms involved in this response. It was found that although a reduced transpiration led to a warmer surface almost everywhere in Australia, the temperature over WA is decreased because of an increase in rainfall, accompanying the increases in leaf-level CO₂. Here we investigate the cause of the increase
in rainfall over WA in the wet case and why it was not simulated in the dry case. Although these changes in rainfall are simulated at various levels of CO₂, only two cases will be examined: 650 ppmv, at which level CO₂ is increased substantially from 280 ppmv, and 1000 ppmv, which is the maximum level of CO₂ in this study.

Figure 7 shows the mean surface air temperatures and the lowest-level winds for the dry and wet cases at 650 ppmv and their differences from 280 ppmv. The land surface is warmer by about 5°–15°C than the surrounding ocean, except for the area north of 15°S in the wet case (Fig. 7b). The eastern region is warmer by more than 5°C than the west in the dry case. The surface is generally cooler in the wet case. There is a strong west–northwesterly monsoonal flow over the tropics, characteristic of the summer monsoon season. The southeasternly trade winds over eastern Australia and the westerly flow south of Australia can also be seen in both cases.
Note the stronger winds over the ocean, east of Australia in the dry case, and a slightly stronger onshore flow south of Australia in the wet case.

The CO₂ increase from 280 to 650 ppmv in the dry case results in a warming over most of the continent, except over south-central Australia (Fig. 7c). On the other hand, there is a clear contrast in the temperature change in the wet case, in which the eastern region is up to 1°C warmer but cooler in the west (Fig. 7d). The moist conditions in the wet case, particularly over south-central Australia (see Fig. 2c) allow the vegetated surface to respond to the increased CO₂, resulting in a reduced transpiration and an increase in both the area and magnitude of warming over central and east Australia (Fig. 7d). The enhanced temperature gradient between the land and sea surface leads to a southward shift of the northwesterly winds in northwest Australia, which brings more tropical moisture into WA. There are also

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**Fig. 7.** Maps of the surface air temperature (°C) and lowest model-level winds (m s⁻¹) at 650 ppmv for the (a) dry and (b) wet cases, and the difference in the surface temperature and winds between 650 and 280 ppmv for the (c) dry and (d) wet cases.
onshore winds from the south, which bring relatively dry air from the Southern Ocean but it is still a net gain of moisture compared to the dry case. These winds converge over WA with the winds from the north, and they overall produce significant accumulation of atmospheric moisture over this area.

Figure 8 shows the changes in the vertically integrated (1000–700 hPa) low-level moisture convergence and vertical velocities for the dry and wet cases, as CO₂ is increased from 280 to 650 ppmv. In the dry case over WA, there is no notable change in the moisture convergence (Fig. 8a) with minimal changes in the vertical velocity (Fig. 8c). On the other hand, both the increased moisture convergence in the wet case (Fig. 8b) and the enhanced rising motion (Fig. 8d) over WA enhance the convection over this area. Once the convection is initiated, it can further enhance the moisture convergence and lead to further rainfall. The increased rainfall also enhances evaporation and consequently lowers surface temperature.

As CO₂ is increased to 1000 ppmv, there is a similar pattern in the changes observed in the dry case, although the magnitude of change is generally larger (Fig. 9c). In the wet case, Fig. 9d indicates a reduction in the cool area over WA. This may be attributed to the weakening of the northwesterly onshore winds in the north that brought moisture-laden winds into Western Australia, compared to when CO₂ was increased to 650 ppmv and the strengthening of the southeasterly dry winds at the south, both of which reduced moisture in the area. Hence, the warming effect of the reduced transpiration begins to dominate over the cooling brought about by increased rainfall (Fig. 9d). At 1000 ppmv, there is a decrease in the low-level moisture convergence over the west coast of WA with stronger changes in vertical velocities inland, relative to 650 ppmv in the dry case (Figs. 10a and 10c). On the other hand, the area of enhanced moisture convergence and rising motion over WA is decreased in the wet case (Figs. 10b and 10c), which lessens the area of increased rainfall and cooling as seen in Fig. 9d.

5. Discussion

Multiple realizations have been conducted for different Januarys where only the CO₂ at the leaf level is changed in the land surface model of a climate model, to characterize the effect of the physiological feedback to various levels of CO₂ over Australia. Our results have shown the clear effect of increasing leaf-level CO₂ on the transpiration via the stomatal response, particularly when sufficient moisture is available. The reductions in transpiration generally result in regional-scale warming with decreases in rainfall, in agreement with results from doubled CO₂ (Sellers et al. 1996; Bounoua et al. 1999; Levis et al. 2000; Betts et al. 2007) or doubled stomatal resistance model experiments (Pollard and Thompson 1995; Henderson-Sellers et al. 1995; Martin et al. 1999).

In addition, we have shown that further increases in CO₂ up to 1000 ppmv lead to statistically significant increases in the magnitude and areal extent of reduced transpiration and rainfall and increased temperature, which are also seen at the global scale (Cruz et al. 2010). Examination of the changes in the diurnal cycle shows that the change in temperature is mainly driven by the daytime changes in transpiration. The changes in the daily means are mainly due to the peak changes that occur from midday through the afternoon.

This paper also examined how the availability of moisture affects the effect of increases in the leaf-level CO₂ on the land surface. Our results indicate the changes are minimal at the present-day value of 375 ppmv in the dry case. However, the wet case showed that changes can occur over a significantly large area, even with an increase of less than 100 ppmv, comparable to the changes seen at levels greater than 500 ppmv in the dry case. In the wet case, a reduction in the water stress allows the stomates to respond more to the CO₂ fertilization, which led to more intense changes in transpiration, temperature, and rainfall. More interestingly, this enhanced the thermal gradient between the land and the ocean, thereby affecting wind flow. As seen over WA, this led to the advection of moist air from the tropics, leading to increases in convective rainfall in the area. Hence, a cooler surface that dominated the warming effect of the reduced transpiration is observed.

The effect of moisture availability on the stomatal response is more likely significant in water-limited regions, such as central Australia, in contrast to wetter regions, such as the tropics. Our results agree with Niyogi and Xue (2006) that the soil moisture status needs to be considered when studying the physiological effects of increased CO₂.

The effect of the physiological feedback on the annual cycle was not examined in this study. Our results may be limited to the changes in the January climate; however, we are confident that these results generated from multiple realizations are statistically robust and not a result of internal model variability. Note that previous studies have also indicated a stronger effect on the climate in the summer months, when vegetation is most active (Henderson-Sellers et al. 1995; Bounoua et al. 1999; Martin et al. 1999). Therefore, our results may be interpreted as a potential maximum effect of the stomatal response to increased CO₂. There are also longer-term feedbacks relating to water use efficiency that we have not resolved in this paper. If stomatal conductance decreases in
response to higher CO₂ and transpiration decreases per unit of carbon fixed, then less water is used by the plants. This means that more water remains in the soil that can be transpired or evaporated for subsequent seasons. The complex interaction between the seasonal changes in soil moisture and the physiological response to increased CO₂ over Australia can be explored in the future using simulations over multiple seasons.

In this study, we have also only changed the CO₂ at the leaf level in our simulations. Sellers et al. (1996) and Bounoua et al. (1999) suggest that including the down-regulation of photosynthesis, as well as changes in the radiative forcing due to higher CO₂, has the potential to enhance the physiological feedback. We have also not considered structural changes in vegetation, which can also result from the increased productivity in an enriched CO₂ environment with sufficient moisture and nutrient availability. The structural responses may partially offset the decrease in transpiration as a result of increased CO₂ (Betts et al. 1997; Levis et al. 2000; Kergoat et al. 2002;
Calvet et al. 2008) as well as affect the effect of the physiological feedback on global runoff (Piao et al. 2007; Betts et al. 2007). Our results may be modified when these factors have been included in the experiments.

Finally, we give two notes of caution. First, the parameterization of stomatal responses to elevated CO$_2$ used in CABLE is based on Leuning (1995) and Wang and Leuning (1998) and is believed to be reasonable. However, elevated CO$_2$ experimental studies using real vegetation are difficult, are not run at very high CO$_2$ levels, and may reflect how vegetation will respond in the real world. Clearly, to improve our confidence in the effect of increased leaf-level CO$_2$ on Australia, systematic field experiments are essential over sustained periods, including sampling El Niño and La Niña conditions, and using Australian native ecosystems. Second, the response of the atmosphere to a land surface perturbation is dependent on the coupling strength between the land and the atmosphere (Koster et al. 2006), and the precise nature of how the land is perturbed (Pitman et al. 2009).
Ideally, a multimodel experiment needs to be conducted to determine those changes that are robust and simulated by multiple models and those changes that appear to be model specific.

6. Conclusions

In this paper we described the regional changes in the Australian January climate, which result from stomatal response to increases in leaf-level CO$_2$ under differing conditions of moisture availability, using a coupled atmosphere–land model. Statistically significant reductions in transpiration generally resulted in significant regional-scale warming with decreases in rainfall, especially when there was enough soil moisture. Various levels of increase in CO$_2$ from 280 to 1000 ppmv led to statistically significant increases in the magnitude and areal extent of the changes in the surface climate. Our results also showed that the availability of moisture substantially affects the effect of increases in the leaf-level CO$_2$, at which level the physiological feedback can indirectly lead to more rainfall and decrease the warming effect of reduced transpiration, such as the anomalous cooling simulated over Western Australia.
The influence of moisture availability suggests that the potential of the physiological feedback to affect the future climate may be affected by uncertainties in rainfall projections, particularly for water-stressed regions. The effect of the physiological feedbacks may still be moderated, if not offset, when other factors are considered, for example, radiative and structural effects of elevated CO₂. The changes in the surface climate presented in our results suggest that the exclusion of the physiological feedbacks to increased CO₂ may inadequately quantify the effect of global warming and human-induced land cover changes on the Australian summer climate. It is therefore important to incorporate these feedbacks into future climate assessments and projections for Australia.

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