Modulation of Land-Use Change Impacts on Temperature Extremes via Land–Atmosphere Coupling over Australia

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ABSTRACT: The role of land–atmosphere coupling in modulating the impact of land-use change (LUC) on regional climate extremes remains uncertain. Using the Weather and Research Forecasting Model, this study combines the Global Land–Atmosphere Coupling Experiment with regional LUC to assess the combined impact of land–atmosphere coupling and LUC on simulated temperature extremes. The experiment is applied to an ensemble of planetary boundary layer (PBL) and cumulus parameterizations to determine the sensitivity of the results to model physics. Results show a consistent weakening in the soil moisture–maximum temperature coupling strength with LUC

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irrespective of the model physics. In contrast, temperature extremes show an asymmetric response to LUC dependent on the choice of PBL scheme, which is linked to differences in the parameterization of vertical transport. This influences convective precipitation, contributing a positive feedback on soil moisture and consequently on the partitioning of the surface turbulent fluxes. The results suggest that the impact of LUC on temperature extremes depends on the land–atmosphere coupling that in turn depends on the choice of PBL. Indeed, the sign of the temperature change in hot extremes resulting from LUC can be changed simply by altering the choice of PBL. The authors also note concerns over the metrics used to measure coupling strength that reflect changes in variance but may not respond to LUC-type perturbations.

**KEYWORDS:** Atmosphere–land interaction; Hydrometeorology; Land surface model

1. Introduction

The Australian continent is increasingly affected by extreme events, particularly those related to temperature (Alexander et al. 2006; Donat and Alexander 2012; Donat et al. 2013; Perkins and Alexander 2013; Seneviratne et al. 2014; Lewis and Karoly 2014). Increases in atmospheric carbon dioxide are a key driver of these increases in continental extremes (Cowan et al. 2014; Lewis and Karoly 2014) with trends superimposed on a background of high natural variability driven by large-scale modes of variability (Risbey et al. 2009).

Land-use change (LUC) also affects the mean climate (Pitman et al. 2009; Pielke et al. 2011; de Noblet-Ducoudré et al. 2012) and climate extremes (e.g., Pitman et al. 2012), particularly at regional scales (Deo et al. 2009; Kala et al. 2011; Nair et al. 2011; Avila et al. 2012). The persistence of droughts and heat waves has also been linked to land processes, mostly through the soil moisture limitation of evapotranspiration (Fischer et al. 2007; Lorenz et al. 2010; Jaeger and Seneviratne 2011; Zhang and Wu 2011; Mueller and Seneviratne 2012; Roundy et al. 2014). LUC modifies the biophysical characteristics of the land surface, the surface energy balance (Boisier et al. 2012), and how the land connects to the boundary layer. In Australia, LUC is most commonly associated with a change from native forest to grasslands and crops and is known through observations to affect the atmosphere. Early studies from the bunny fence experiment field campaign (Lyons et al. 1993) showed preferential cloud formation over native vegetation as compared to agriculture in Western Australia (Lyons et al. 1993; Lyons 2002; Ray et al. 2003). Ray et al. (2003) also showed that cumulus clouds occur preferentially over the native vegetation up to 10% of the time during the austral summer. Studies using single-column models have suggested that this is due to higher planetary boundary layer (PBL) heights over native vegetation exceeding the lifting condensation level (LCL) and increasing the possibility of cloud formation (Lyons et al. 1993; Lyons 2002). These mechanisms have been supported by aircraft observations (Lyons et al. 1993, 2001).

One way of quantifying how the land surface affects the atmosphere is the “coupling strength” approach introduced by Koster et al. (2004). The first Global Land–Atmosphere Coupling Experiment (GLACE-1; Koster et al. 2006) used a methodology for evaluating the strength of land–atmosphere coupling on seasonal time scales using general circulation models (GCMs). Regions of strong coupling, where precipitation and temperature are sensitive to soil
moisture variability, were located in the transition zones between humid and arid climates. GLACE-1 results focused on the Northern Hemisphere with simulations run over a single boreal summer [June–August (JJA)] and were not applicable for the Southern Hemisphere summer. This was the motivation for Hirsch et al. (2014b) to undertake the first assessment of coupling strength for the austral summer [December–February (DJF)] over Australia. By applying the GLACE-1 methodology in the Weather Research and Forecasting (WRF) Model (Skamarock et al. 2008), multiple simulations were conducted to understand the uncertainty associated with different model physics. Hirsch et al. (2014b) also considered the role of interannual variability of land–atmosphere coupling strength building on Northern Hemisphere work by Guo and Dirmeyer (2013). Hirsch et al. (2014b) showed Australia to be strongly coupled and therefore another land–atmosphere “hotspot” region, although the strength of the coupling was strongly influenced by large-scale modes of variability. Specifically, regions of strong coupling extend southward from the tropics during wetter (commonly La Niña) years and contract when less soil moisture is available (commonly El Niño) years. During dry years when soil moisture is limited, the land–atmosphere coupling strength was particularly sensitive to the choice of model physics in WRF.

Most research on land–atmosphere coupling emphasizes the role of soil moisture variability (e.g., Koster et al. 2004; Guo et al. 2006; Koster et al. 2006; Hirsch et al. 2014b). However, there are good reasons to expect that both LUC and land–atmosphere feedbacks linked with the PBL also influence extremes (Dirmeyer 2011). Hirsch et al. (2014a) therefore combined the GLACE-1 methodology with LUC and found that LUC had a profound effect on land–atmosphere coupling strength in WRF. Results showed a consistent weakening of soil moisture–maximum temperature coupling strength for regions that experienced a change from forest to crops, irrespective of the choice of WRF physics parameterizations. Lorenz and Pitman (2014) also connected LUC and coupling strength by examining the impact of increasing the scale of Amazonian deforestation on regional rainfall and mean temperature over a “strongly” and “weakly” land–atmosphere coupled region. They found that a small deforestation signal imposed over a strongly coupled region could have a larger impact on temperature than a large deforestation signal over a weakly coupled region. This suggests that the impact of LUC is closely tied to the model’s land–atmosphere coupling strength because regional coupling strength estimates are highly variable between models (Koster et al. 2004).

In this paper, we examine how LUC and land–atmosphere coupling modulates temperature extremes over Australia. We focus on the use of regional models to examine the impact of LUC. In virtually all regional modeling studies, LUC is examined using a single modeling system, or occasionally results are compared across models. We take a different approach by examining how the impact of LUC varies in an ensemble of simulations conducted using four configurations of the PBL and cumulus parameterization schemes. We seek to determine whether approaches used in modeling provide consistent estimates of the impact of LUC on temperature extremes or whether differences in land–atmosphere coupling directly influence how LUC affects the model simulations. We note here that we are not attempting to predict the real impact of LUC. Instead, we are examining the sensitivity of the simulated impact as a function of model structure, land cover, and
physics (PBL and cumulus convection) to determine whether single-model estimates of the impact of LUC can be reliable.

We focus on the Australian continent for several reasons. First, the Australian continent is a likely land–atmosphere coupling hotspot (Hirsch et al. 2014b; Lorenz et al. 2015), and we focus on temperature extremes as Hirsch et al. (2014b) found a spatially coherent signal in soil moisture–temperature coupling. Second, the Australian continent has experienced significant LUC since European settlement [Australian Surveying and Land Information Group (AUSLIG) 1990]. Finally, Australia has a wide range of environments ranging from tropical forests in the north through to temperate and Mediterranean climates in the south, east, and west, to hyperarid desert environments away from the coast (Gentilli 1972). Australia therefore provides an ideal test case for evaluating the relationship between LUC, temperature extremes, and land–atmosphere coupling.

2. Methodology

2.1. Model description

WRF is a community regional weather and climate model with a nonhydrostatic Eulerian dynamical core with terrain-following pressure-based vertical coordinates (Skamarock et al. 2008). WRF simulations are typically forced with reanalysis or GCM output (for scenario simulations) at 6-hourly intervals to define the lateral boundary conditions.

We use WRF coupled to the Community Atmosphere Biosphere Land Exchange (CABLE) land surface model (LSM) (Wang et al. 2011). CABLE includes a coupled model of stomatal conductance, photosynthesis, and partitioning of absorbed net radiation into latent and sensible heat fluxes. Soil and vegetation fluxes are calculated separately and then linearly combined into the total sensible and latent heat fluxes that form the lower boundary condition of the atmospheric model. A canopy turbulence model is used to calculate within canopy air temperatures and humidity. CABLE includes a multilayer soil model with six layers, with the deepest layer at 2.872 m. There are nine soil types used to prescribe hydraulic and thermal characteristics. The flow of water is parameterized using Darcy’s law, and the hydraulic conductivity is related to soil moisture via the Clapp and Hornberger (1978) relationship.

We use the NASA Land Information System (LIS; version 6.0) to couple CABLE (version 2.0) to WRF (version 3.2.1). LIS is a software framework for running high-resolution land data assimilation systems that integrate advanced LSMs with high-resolution satellite and in situ observational data to accurately characterize land surface states and fluxes (Kumar et al. 2006). We also used LIS to run CABLE offline simulations to provide equilibrated soil moisture and temperature initial conditions for the fully coupled WRF simulations following Hirsch et al. (2014a,b). LIS offline simulations require appropriate surface meteorological forcing to solve the governing equations of the soil–vegetation–snow system and predict the surface turbulent energy fluxes and soil states.

For coupled simulations, LIS–CABLE provides the surface turbulent energy fluxes to WRF, and WRF provides the surface temperature, humidity, wind speed, total precipitation, and short- and longwave radiation to LIS–CABLE. LIS–CABLE is treated like other LSMs within WRF, with the forcing fields mapped onto the
appropriate model tiles for LIS. The LIS software framework has been documented by Kumar et al. (2006) and Peters-Lidard et al. (2007).

2.2. Model configuration

The model domain is centered at 27.5°S, 133.0°E on a Lambert projection with a spatial resolution of 50 km and 30 atmospheric levels. A model time step of 180 s is used and up to four tiles per grid cell are implemented in CABLE to help resolve land surface heterogeneity. The model is initialized and updated at the lateral boundaries using ERA-Interim reanalysis (Dee et al. 2011) at 6-hourly intervals, with surface initial conditions obtained from prior LIS–CABLE offline simulations.

We implement four configurations of the PBL and cumulus parameterization schemes to evaluate the uncertainty of our results to WRF model physics. This includes two PBL schemes, Yonsei University (YSU; Hong et al. 2006) and Mellor–Yamada–Janjic (MYJ; Janjic 1994), and two cumulus schemes, Kain–Fritsch (KF; Kain 2004) and Betts–Miller–Janjic (BMJ; Betts 1986; Betts and Miller 1986). The PBL schemes differ in closure order and the representation of vertical mixing. The YSU PBL scheme uses countergradient terms to represent fluxes due to nonlocal gradients and explicit treatment of entrainment at the top of the PBL defined by the buoyancy profile. The MYJ scheme represents a 2.5 turbulent closure model where the upper limit depends on the turbulent kinetic energy, buoyancy, and shear of the driving flow, with different limits applied depending on the atmospheric stability. The cumulus schemes differ in their triggering assumptions for precipitation, which can also be influenced by the choice of the PBL scheme, given this affects how heat and moisture are transferred from the surface. The other WRF physics parameterizations implemented in this study include the WRF Single-Moment 5-Class Microphysics scheme, the Rapid Radiative Transfer Model longwave scheme (Mlawer et al. 1997), and the Dudhia shortwave scheme (Dudhia 1989).

2.3. Implementation of land-cover change

The Australian Surveying and Land Information Group (AUSLIG 1990) provide land-cover information for pre-European (1788, hereafter NATIVE) and post-European (1988, hereafter PRESENT) settlement of Australia. Both land-cover distributions are mapped onto the CABLE plant functional types (PFTs). There are fewer CABLE PFTs than those provided by AUSLIG (1990) and most forest-type PFTs are mapped onto CABLE’s evergreen broadleaf PFT, and most grasses are mapped onto CABLE’s C3 grass or crop PFT following Hirsch et al. (2014a). The leaf area index (LAI) is prescribed using PFT-specific values to maintain consistency with the land-cover representation. We ensured that the surface albedo and roughness length were consistent with the land-cover representation. The main LUC transition occurs over southwest Western Australia (SWWA) and southeast Australia (SEA), coinciding with large-scale clearing of native forests that were replaced with crops and grasses (Figures 1a,b). This change in vegetation potentially provides a strong and spatially coherent forcing that should enable us to
examine the impact of land-cover change on climate extremes and how this is modulated by land–atmosphere interactions. We note that the land-cover descriptions were identical for each WRF physics configuration for PRESENT land cover and likewise for NATIVE land cover, with the change in surface albedo (Figure 1c), LAI (Figure 1d), and roughness length (not shown) identical for each WRF physics configuration. Note that there is a region of Queensland (20°S, 145°E) where the LAI and albedo increase; this is where sparse vegetation is replaced by shrubs.

2.4. Experimental design

We follow the GLACE-type experiments of Hirsch et al. (2014a) and run simulations for the austral summer (DJF) for each land-cover distribution: PRESENT and NATIVE. Each experiment consists of a fully coupled and uncoupled 16-member ensemble. Two-way land–atmosphere interactions are enabled for the fully coupled ensemble, with subsurface soil moisture written to file at every model step.
time step from one of the ensemble members. These values are then used to pre-
scribe subsurface soil moisture in the uncoupled ensemble member simulations to
effectively decouple the land from atmospheric feedback.

The ensemble initialization and boundary conditions use 16 different austral
summers starting on 1 December from the ERA-Interim lateral boundary condi-
tions to capture the influence of interannual variability across the region. We focus
on austral summer because this is the season when high net radiation combined
with an active boundary layer has the highest potential to affect temperature ex-
tremes, consistent with observations of consistently higher PBL heights during this
season over native vegetation in Western Australia (Nair et al. 2011). We examine
one dry (El Niño: December 1982 to February 1983) and one wet (La Niña: De-
cember 1999 to February 2000) summer in the prescribed soil moisture ensembles
to capture the range of soil moisture states on the coupling. The coupled ensemble
simulations differ only by the forcing data, whereas for the uncoupled ensemble
simulations, the prescribed soil moisture states are selected from one of the coupled
ensemble member simulations.

Coupling strength $\Delta \Omega$ is evaluated following Koster et al. (2006). For the 90-day
ensemble member simulations, the first 6 days are discarded as spinup (for the
atmosphere), with the remaining 84 days used to compute fourteen 6-day averages.
For each ensemble, coupled and uncoupled, $\Omega$ is calculated as
\[
\Omega = \frac{16\sigma_{X}^2 - \sigma_{X}^2}{15\sigma_{X}^2},
\]
where $\sigma_{X}^2$ corresponds to the all-ensemble variance of variable $X$, and $\sigma_{X}^2$ corresponds to the variance of the ensemble-mean time series. The $\Omega$ values ap-
proaching 1 indicate that the ensemble members are identical. The term $\Delta \Omega$ is then
inferred by taking $\Omega[\text{uncoupled}]$ minus $\Omega[\text{coupled}]$, where values of $\Delta \Omega > 0$
indicate that the members of the uncoupled ensemble simulations converge more
quickly than the coupled ensemble.

We examine the impacts of LUC and land–atmosphere coupling using the ex-
tremes indices sourced from daily temperature values as recommended by the
Expert Team on Climate Change Detection and Indices (ETCCDI; Zhang et al.
2011). Since we are simulating seasonal time scales over the austral summer only,
we limit our analysis to the highest (hottest) maximum temperature $T_{\text{x}}$ and the
lowest (coldest) minimum temperature $T_{\text{Nn}}$. These indices are calculated from
daily maximum $T_{\text{MAX}}$ and minimum $T_{\text{MIN}}$ temperatures. Other ETCCDI indices
based on the frequency of temperatures exceeding percentage thresholds were also
examined but found to share common attributes with $T_{\text{Nn}}, T_{\text{MIN}}, T_{\text{MAX}},$ and $T_{\text{x}}$
and therefore are not reported here.

3. Results

3.1. Impact of LUC on land–atmosphere coupling

Hirsch et al. (2014a) evaluated the change in the GLACE-1 soil moisture–
maximum temperature coupling strength $\Delta \Omega_{T_{\text{MAX}}}$ associated with LUC. Figure 2
reproduces this result, illustrating $\Delta \Omega_{T_{\text{MAX}}}$ for PRESENT and NATIVE land cover
and the change (PRESENT minus NATIVE) for each of the WRF physics
Figure 2. Soil moisture–maximum temperature coupling strength $\Delta \Omega$ for WRF–LIS–CABLE for (left) PRESENT, (middle) NATIVE, and (right) PRESENT minus NATIVE land-cover description. (a)–(c) YSU–KF physics with dry soil moisture, (d)–(f) YSU–KF physics with wet soil moisture, (g)–(i) YSU–BMJ physics with dry soil moisture, (j)–(l) YSU–BMJ physics with wet soil moisture, (m)–(o) MYJ–KF physics with dry soil moisture, (p)–(r) MYJ–KF physics with wet soil moisture, (s)–(u) MYJ–BMJ physics with dry soil moisture, and (v)–(x) MYJ–BMJ physics with wet soil moisture.
configuration and soil moisture cases examined. Consistent with Hirsch et al. (2014b), the estimates of $\Delta T_{\text{MAX}}$ for PRESENT and NATIVE show that the coupling is stronger under wetter soil conditions for all WRF physics configurations. The change $\Delta T_{\text{MAX}}$ between PRESENT and NATIVE (Figure 2, right column) generally shows a weakening of approximately $-0.25$ across regions where forest is replaced with crops. There are some spatially coherent regions where the coupling strengthens by $+0.20$, particularly in the wet soil moisture cases (Figures 2f,l,r,x). These regions are associated with sparse woodland regions becoming more densely vegetated.

### 3.2. Impact of LUC on temperature extremes

We next focus our attention on temperature extremes as they show a coherent response to LUC (Figure 3) over the regions where a change was imposed. Figure 3 shows the actual change (PRESENT minus NATIVE) in the coldest minimum temperature $T_{\text{MIN}}$, mean minimum temperature $T_{\text{MIN}}$, mean temperature $T_2$, mean maximum temperature $T_{\text{MAX}}$, and hottest maximum temperature $T_{\text{MAX}}$ for DJF for each WRF physics configuration. The impact of LUC on each temperature extreme is commonly larger than the impact on the mean temperature, a result consistent
across all WRF physics configurations. This is most obvious for the more extreme temperatures (TXx and TNn), where changes exceeding 2°C occur across SEA and SWWA in contrast to less than 1°C for mean temperature. The extreme temperature response also shows dependence on the choice of PBL scheme, with the MYJ scheme (Figure 3, bottom two rows) showing larger changes in the temperature extremes than the YSU scheme (Figure 3, top two rows). Furthermore, for  \( T_{\text{MAX}} \) and TXx, the LUC impact varies in sign between the PBL schemes, particularly over SWWA, with a decrease for the YSU scheme (\(-1.5°C\); Figures 3d,i) and an increase in the MYJ scheme (0.5°–1.0°C; Figures 3n,s). The larger temperature response with the MYJ scheme, associated with a decrease in TNn and  \( T_{\text{MIN}} \) and the increase in TXx and  \( T_{\text{MAX}} \), causes a widening of the diurnal temperature range (DTR; not shown), particularly over SEA. In contrast, for the YSU scheme, the decrease in  \( T_{\text{MIN}} \) (\(-1°C\)) and  \( T_{\text{MAX}} \) (\(-2°C\)) over SWWA causes a reduction of the DTR.

The critical result here is the difference in the sign of the temperature response to LUC between the PBL schemes, particularly over SEA. We explore this further in Figures 4 to 7 by examining how the LUC signal propagates through the components of the surface energy balance and within the boundary layer. Across all the WRF physics schemes, the same LUC signal was imposed as demonstrated for

\[
\begin{align*}
\Delta R_{\text{net}} & \quad \Delta Q_h & \quad \Delta Q_{\text{le}} & \quad \Delta (Q_h + Q_{\text{le}}) & \quad \Delta \text{Soil Moisture} \\
\text{YSU-KF} & \quad \text{(a)} & \quad \text{(b)} & \quad \text{(c)} & \quad \text{(d)} & \quad \text{(e)} \\
\text{YSU-BMJ} & \quad \text{(f)} & \quad \text{(g)} & \quad \text{(h)} & \quad \text{(i)} & \quad \text{(j)} \\
\text{MYJ-KF} & \quad \text{(k)} & \quad \text{(l)} & \quad \text{(m)} & \quad \text{(n)} & \quad \text{(o)} \\
\text{MYJ-BMJ} & \quad \text{(p)} & \quad \text{(q)} & \quad \text{(r)} & \quad \text{(s)} & \quad \text{(t)} \\
\end{align*}
\]

Figure 4. The change (PRESENT minus NATIVE) in (first column) DJF net radiation  \( R_{\text{net}} \), (second column) sensible heat flux  \( Q_h \), (third column) latent heat flux  \( Q_{\text{le}} \), (fourth column) the total turbulent flux  \( Q_h + Q_{\text{le}} \), and (fifth column) total column soil moisture. (a)–(e) YSU–KF physics, (f)–(j) YSU–BMJ physics, (k)–(o) MYJ–KF physics, and (p)–(t) MYJ–BMJ physics. Net radiation and all fluxes are in units of W m\(^{-2}\) and soil moisture is in units of 1000 m\(^3\) m\(^{-3}\).
surface albedo in Figure 1c. This corresponds to a 10%–20% increase across regions where forests were replaced with crops (mostly SWWA and SEA) and a 10% decrease over regions that became more densely vegetated. Focusing on SWWA and SEA where there was deforestation, the increase in surface albedo increases the amount of shortwave radiation reflected by the surface. This then contributes to reducing the net available energy at the surface (Figures 4a,f,k,p). The decrease in net radiation also reduces the magnitude of the surface turbulent energy fluxes. The decrease in sensible heat flux $Q_h$ (Figures 4b,g,l,q) is similar across all the WRF physics configurations (10–20 W m$^{-2}$). This contributes to decreasing mean surface air temperature for those WRF physics configurations using the MYJ PBL scheme (Figures 3m,r). For the YSU PBL scheme, the decrease in $Q_h$ does not always contribute to a decrease in mean surface temperature over SEA (Figures 3c,h). This is examined further in our deconstruction of the surface energy balance.
The change in latent heat flux $Q_{le}$ (Figures 4c,m,r) illustrates a split between the PBL schemes. When the YSU PBL scheme is used, LUC causes a mixture of increases and decreases in $Q_{le}$ over SEA (Figures 4c,h). If the MYJ PBL scheme is used, LUC leads to a clear and large increase in $Q_{le}$ (commonly by 5–10 W m$^{-2}$; Figures 4m,r) over both SWWA and SEA. In short, the impact of the same LUC perturbation is translated differently into an atmospheric temperature response depending on the PBL scheme because the change in net radiation is more strongly converted into an increase in $Q_{le}$ when using MYJ relative to YSU. Despite the contrast in the $Q_{le}$ response to LUC between the PBL schemes, the total turbulent flux ($Q_{h} + Q_{le}$) (Figures 4d,i,n,s) shows a decrease for both PBL schemes. This is consistent with the decrease in net radiation and decrease in the aerodynamic roughness length (not shown). However, the magnitude is smaller for the MYJ (decrease is about 5–10 W m$^{-2}$) than the YSU PBL scheme (decrease is about 10–20 W m$^{-2}$). The contrast in the $Q_{le}$ result between the PBL schemes is related to the

![Figure 6. The percentage change (PRESENT minus NATIVE) in DJF (first column) LCL, (second column) LFC, (third column) CAPE, and (fourth column) convective precipitation RAINC. (a)–(d) YSU–KF physics, (e)–(h) YSU–BMJ physics, (i)–(l) MYJ–KF physics, and (m)–(p) MYJ–BMJ physics. For ease of comparison, all variables are presented in units of percentage (100x(PRESENT – NATIVE)/NATIVE).]
change in soil moisture (Figures 4e,j,o,t). The MYJ scheme shows a larger increase in soil moisture (Figures 4o,t) than the YSU scheme (Figures 4e,j). Thus, more soil moisture is available for vegetation to transpire in the MYJ scheme and hence contributes to higher $Q_{le}$. Since the same LUC perturbation is applied across all WRF physics configurations, the difference in the soil moisture response to LUC between the PBL schemes suggests differences in the precipitation feedback on soil moisture and consequently in $Q_{le}$. This difference is likely associated with how vertical mixing contributes to triggering convective precipitation, which is parameterized differently between the physics configurations considered in this paper. We evaluate this in Figures 5 and 6 with further analysis of the surface energy balance and the boundary layer.

The surface turbulent energy fluxes form the bottom boundary conditions for the PBL scheme, which is responsible for the vertical subgrid-scale fluxes within the whole atmospheric column. The different treatment of vertical mixing between the YSU and MYJ schemes therefore contributes to differences in the atmospheric temperature and moisture. This is evident in the shortwave and longwave radiation components of the surface energy balance that lead to further differences between $Q_{le}$ and $Q_h$. Changes in the outgoing longwave radiation $LW_{up}$ (Figure 5, first column) result from changes in skin temperature and cloud cover. Given the changes in surface temperature shown in Figure 3, a corresponding change in $LW_{up}$ is anticipated, although in this case it is much smaller in magnitude ($<5 \text{ W m}^{-2}$; Figures 5a,e,i,m) than the change in $Q_{le}$ and $Q_h$. For the YSU scheme, the increase in $LW_{up}$ (Figures 5a,e) is consistent with the higher temperatures (Figures 3c,h). For the MYJ scheme, the change in $LW_{up}$ is negligible (Figures 5i,m) and not always consistent with the temperature change (Figures 3m,r). There is greater influence from the changes in cloud cover for configurations that use the MYJ scheme (Figures 5j,n). Changes in cloud cover (Figures 5b,f) are generally consistent with the changes in $LW_{up}$ for the YSU PBL scheme.

Figure 7. Schematic of the process by which the PBL schemes differ in the surface energy balance response to the same LUC perturbation. For (left) YSU and (right) MYJ.
Vertical transport of heat and moisture are also parameterized differently between the PBL schemes. The differences in the sign of the changes in $Q_{le}$, $Q_h$, and cloud cover are associated with different treatment of vertical transport and triggering of convection. Changes in cloud cover affect incoming shortwave and longwave radiation (last two columns of Figure 5), and the sign of the change due to LUC in both longwave LW$_{down}$ (Figures 5c,g,k,o) and shortwave SW$_{down}$ (Figures 5d,h,l,p) incoming radiation is dependent on the PBL scheme. The differences in LW$_{down}$ and SW$_{down}$ contribute a positive feedback on the net radiation that enhances the deviation in $Q_{le}$ between the schemes (Figures 4c,h,m,r). Our results are consistent with a hypothesis that as the model integration moves forward in time, this split between the PBL schemes of the surface energy balance becomes more pronounced, and therefore the asymmetry in the temperature response illustrated in Figure 3 becomes more clearly defined. To resolve this hypothesis, a suite of additional simulations would be required, saving many atmospheric and land surface variables at every time step, something that is prohibitively expensive at this time.

The PBL scheme asymmetry in the surface energy balance (Figures 4, 5) is examined further by comparing the change in key atmospheric properties of the boundary layer: LCL, the level of free convection (LFC), and the convective available potential energy (CAPE). The LCL is the height above which an air parcel has to be raised for condensation to occur (Figures 6a,e,i,m). A reduction in LCL implies less energy is required for lifting air parcels high enough for condensation to occur. The LFC is the height at which a saturated parcel of air becomes warmer than its surrounding environment and rises freely (Figures 6b,f,j,n), and hence a lower LFC would imply less energy is required for convection to occur. CAPE is the amount of positive buoyant energy of an air parcel (Figures 6c,g,k,o), and higher CAPE values are more likely to result in convective rainfall. Figure 6 shows that the YSU scheme generally results in lower LCL (Figures 6a,e) and LFC (Figures 6b,f) but also lower CAPE (Figures 6c,g) values over large parts of SEA. This results in an overall decrease in convective rainfall in that region (Figures 6d,h). There are regions of increase in convective rainfall, which tend to coincide well with grid points where there is a reduction in the LCL and LFC and an increase in CAPE. The MYJ scheme results in an overall increase in LCL (Figures 6i,m) and LFC (Figures 6j,n), which would imply lower chances of convective rainfall; however, the increase in CAPE (Figures 6k,o) is sufficiently large to result in overall increases in convective rainfall over most of the domain (Figures 6l,p). Overall, the changes in CAPE can explain the changes in convective rainfall very well, with an increase in CAPE leading to an increase in convective rainfall and vice versa for all experiments. The larger increase in convective rainfall for the MYJ as compared to the YSU experiments explains why the change in soil moisture is larger and more positive in the MYJ scheme (Figures 4o,t) than the YSU scheme (Figures 4e,j) and consequently why the $Q_{le}$ response to LUC varies in sign, particularly over SEA.

The asymmetry between the PBL schemes over SEA can be summarized by the surface energy balance response to the same LUC perturbation (Figure 7). For the YSU scheme, there is a decrease in both $Q_h$ and $Q_{le}$, a counterintuitive result, whereas for the MYJ scheme increases in $Q_{le}$ are accompanied by decreases in $Q_h$, as expected. An examination of the changes in the Bowen ratio (i.e., $Q_h/Q_{le}$)
showed an overall decrease across all simulations (not shown), which could not explain this discrepancy. Instead, the YSU result can be explained by the increase in surface albedo, decrease in roughness length, and the changes in precipitation. The increase in albedo would be the primary mechanism leading to lower $Q_h$, and the decrease in surface roughness, together with the reduction in convective rainfall can explain the reduced $Q_{le}$. The decrease in $Q_{le}$ is sufficient to cause an increase in mean temperature $T_2$. The $T_2$ increase contributes to an increase in LW$_{up}$ and the $Q_{le}$ decrease contributes to less cloud, leading to an increase in SW$_{down}$ and decrease LW$_{down}$. For the MYJ scheme, the same increase in surface albedo and decrease in roughness length contributes to a decrease in $Q_h$, with a corresponding decrease in $T_2$ and therefore also in LW$_{up}$. The increase in precipitation, and therefore soil moisture for the MYJ scheme over most of the domain would also lead to higher $Q_{le}$. Hence, a decomposition of the surface energy balance can explain the asymmetry between the YSU and MYJ schemes in response to LUC and subsequently the impact on mean temperature (Figures 3c,h,m,r). This analysis of mean temperature breaks down when we consider the response in the temperature extremes, particularly for $T_{\text{MAX}}$ (Figures 3d,i,n,s) and TX$_x$ (Figures 3e,j,o,t), because we use seasonal means of the fluxes while the temperature extremes are derived using daily values, which would be driven by daily fluxes.

3.3. Contrasting the coupling response and the extremes response

In the preceding analysis, we found a consistent shift in coupling strength with LUC irrespective of the WRF physics configuration (Figure 2). However, the impact of LUC on temperature extremes (Figure 3) varies between the PBL schemes, in part linked to how each scheme parameterizes vertical transport and how that influences the surface energy balance (Figures 4–7). This inconsistency between the coupling response and the temperature extremes response can be explained using the probability density functions (PDFs) of $T_{\text{MAX}}$ (Figures 8a–c) and $T_{\text{MIN}}$ (Figures 8d–f). The $T_{\text{MAX}}$ PDFs for PRESENT (Figure 8a) and NATIVE (Figure 8b) land cover show that although each WRF physics configuration has different $T_{\text{MAX}}$ means, the variance and spread of the distributions are similar. Given that the GLACE-I $\Delta \Omega_{\text{TMAX}}$ is a function of the $T_{\text{MAX}}$ variance, the consistent variance across the WRF physics configurations explains the consistent coupling response to LUC (Figure 2). For $T_{\text{MIN}}$, the temperature distributions are generally indistinguishable between the different WRF physics configurations for both land-cover distributions (Figures 8d,e). For both $T_{\text{MAX}}$ and $T_{\text{MIN}}$, the PDFs of the temperature change with LUC (PRESENT minus NATIVE; Figures 8c,f) are slightly shifted between the two PBL schemes, particularly for $T_{\text{MIN}}$ where the MYJ scheme has a longer negative tail than the YSU scheme combinations (Figure 8f). These changes are consistent with the temperature extreme response to LUC illustrated in Figure 3, that is, a stronger cooling in the cold extremes with the MYJ scheme.

A similar examination of the PDFs of the coupling strength for $T_{\text{MAX}}$ and $T_{\text{MIN}}$ shows that the change in the coupling strength between PRESENT and NATIVE was generally consistent across the different WRF physics configurations (not shown). In summary, the difference between the coupling and extremes response
to LUC is due to LUC shifting the temperature mean but not the temperature variance.

3.4. Relationship between LUC, temperature extremes, and coupling

To investigate the relationship between temperature extremes and land–atmosphere coupling, we next examine the joint PDF between each temperature index and the corresponding soil moisture–temperature coupling strength for PRESENT and NATIVE land cover separately. To derive the PDFs we aggregated results across all WRF physics configurations (Figure 9). The PDF of the temperature index is also included in addition to identifying the evapotranspiration regime (soil moisture limited or energy limited) for each 2°C temperature bin. For each temperature bin, the corresponding evaporative fraction \(\left[\text{EF} = \frac{Q_{le}}{Q_{le} + Q_h}\right]\) and soil moisture values were used to determine the rate of change in EF. Here, a positive EF gradient indicates a soil moisture limited regime (red dots) and a shallow or no gradient indicates an energy-limited regime (blue dots). For TXx (Figures 9a–d) and TMAX (Figures 9e–h), the hotter temperatures, where the temperature PDF peaks, coincide with regions that are more strongly coupled. These regions also tend to be those where evapotranspiration is soil moisture limited. Cooler TMAX
and TXx temperatures tend to be associated with weaker $T_{\text{MAX}}$ coupling over regions where evapotranspiration is energy limited. This relationship between hot temperature extremes ($T_{\text{MAX}}$ and TXx) and soil moisture–$T_{\text{MAX}}$ coupling is consistent between the PRESENT and NATIVE land-cover descriptions. It is also
stronger for TXx than $T_{MAX}$ and holds for both the dry and wet soil moisture cases, indicating an influence from interannual variability. The relationship between temperature extremes and soil moisture–$T_{MAX}$ coupling strength did not change among the different WRF physics configurations (not shown). For $T_{MIN}$ (Figures 9i–l) and TNn (Figures 9m–p), soil moisture–$T_{MIN}$ coupling is generally weaker with the corresponding flatter temperature distributions. Evapotranspiration is mostly soil moisture limited, although the less extreme minimum temperatures (particularly TNn) are generally located in energy-limited evapotranspiration regimes (Figures 9m–p).

In summary, therefore, higher temperature extremes tend to coincide with stronger coupling where evapotranspiration is soil moisture limited. This relationship becomes stronger for higher temperature extremes, particularly when more soil moisture is available to support the higher evapotranspiration.

4. Discussion

In this paper, we seek to determine how LUC and land–atmosphere coupling modulate simulated temperature extremes over Australia. We used the WRF model, and four physics configurations, to explore whether there is any dependence on how LUC affects temperature extremes associated with model physics.

Our results show that LUC leads to a change in the modeled surface energy balance influencing the nature of land–atmosphere interactions and temperature extremes for the Australian summer. This is not a surprise, of course; it is well known that LUC affects the surface energy balance. What is new is that the sign of the temperature change due to LUC can be changed depending on the choice of PBL scheme. We cannot definitively conclude that this asymmetry stems from the partitioning of the turbulent fluxes. Differences in precipitation timing, associated with the triggering of convective precipitation, may also contribute to differences in soil moisture and therefore in the partitioning of the fluxes. We also note that the choice of PBL scheme influences the LUC impact on temperature extremes more than the choice of cumulus scheme. This is associated with the treatment of vertical mixing in the respective schemes. Despite the implementation of the same LUC perturbation in each WRF physics configuration, the different treatment of vertical mixing contributed to a different feedback on $Q_{le}$.

Our results point to some model physics dependence for the impact of LUC on temperature extremes, but not for soil moisture–temperature coupling strength. This inconsistency was related to how the LUC perturbation contributes to changing the mean, and therefore also extreme temperatures, but not the temperature variance in the fully coupled simulations that are used to calculate the coupling strength. This suggests that there is an incompatibility in using land–atmosphere coupling strength calculated from variances to explore the impact of LUC on temperature extremes.

The consistent decrease in GLACE soil moisture–maximum temperature coupling strength with LUC reported here is associated with a decrease in the temperature variance of the uncoupled simulations reported by Hirsch et al. (2014a). Changes in the surface energy balance in response to LUC showed a consistent decrease in $Q_{h}$ across all WRF physics configurations (Figure 4, second column) that have a direct effect on surface temperature. This change is associated with
changes in net available energy at the land surface that is partitioned into $Q_h$ and $Q_{le}$ as a function of soil moisture. We examined the partitioning of net radiation for each WRF physics configuration (Figure 10, left column) and the partitioning of $Q_{le}$ between soil evaporation and transpiration (Figure 10, right column). Generally, EF increases for all WRF physics configurations over regions that transition from forest to crops (Figure 10, left column). This is consistent with the decreases shown in $Q_h$ (Figure 4, second column). However, the magnitude of the EF change varies between each of the WRF physics configurations, particularly over SWWA where there is a clear dependence on the choice of PBL scheme. This is associated with the differences in how $Q_{le}$ changes over this region where there is negligible change using the YSU model, but a large increase using the MYJ model (Figure 4, third column). We examine this change in the partitioning of $Q_{le}$ in CABLE using soil evaporation and transpiration (Figure 10, right column). With the exception of SWWA, the partitioning of $Q_{le}$ is generally consistent across the WRF physics configurations, and the increase in soil evaporation over regions where forests are replaced by crops is consistent with a decrease in transpiration. Conversely, over regions where the vegetation becomes denser, soil evaporation decreases and transpiration increases. These changes in $Q_{le}$ partitioning are generally consistent with the changes in the coupling strength.

5. Conclusions

We have combined the GLACE-1 methodology with LUC to explore the relationship between LUC, land–atmosphere coupling, and temperature extremes for the Australian summer. Regions experiencing deforestation are associated with a weakening of soil moisture–maximum temperature coupling strength that is robust to the choice of WRF physics. In contrast, the LUC impact on temperature extremes is dependent on the choice of PBL scheme, although deforestation is usually associated with a decrease in minimum and increases in maximum temperature extremes. The deviation between the two PBL schemes employed in this study was associated with their distinct parameterizations of vertical mixing, leading to differences in precipitation that contributed a positive feedback on soil moisture. The inconsistency between the coupling response and temperature extremes response to LUC was associated with how LUC only affects the mean temperature distribution and not temperature variance.

The weakening of the coupling strength was associated with changes in the partitioning of net radiation into $Q_h$ and $Q_{le}$ and further with the partitioning of $Q_{le}$ between soil evaporation and transpiration. In particular, the change in coupling strength appears to be inversely related to the change in soil evaporation. The response of temperature extremes to LUC therefore depends on simulating the appropriate evapotranspiration regime, which in turn depends on the coupling. The impact of LUC on Australian regional temperature extremes is therefore associated with changes in the partitioning of the surface energy balance, which affect the coupling between the land and the atmosphere. This is model physics dependent, and therefore large ensembles sampling a range of parameterizations of the boundary layer could help quantify the uncertainties associated with the simulated impacts of LUC on regional climate extremes. Our results certainly suggest that any single-model estimate of the impacts of LUC on temperature extremes will
Figure 10. PRESENT minus NATIVE change in the (left column) evaporative fraction \( \Delta EF = \frac{Q_{le}}{Q_{le} + Q_{h}} \) and (right column) soil evaporation fraction \( \Delta Q_{soil}/Q_{le} \) for (a), (b) YSU–KF physics, (c),(d) YSU–BMJ physics, (e),(f) MYJ–KF physics, and (g), (h) MYJ–BMJ physics.
depend critically on the choice of PBL parameterization, and therefore any single model has the potential to be highly misleading. Our results also point to a challenge in using the $\Delta \Omega$ metrics for coupling strength to evaluate the impact of LUC on temperature extremes since the response of the $\Delta \Omega$ metric to LUC will not change with model physics, at least, in WRF. Finally, we emphasize that different combinations of convection and PBL scheme clearly affect how LUC impacts WRF. It is inevitable that some of these modules reflect the real world better than others. We have not attempted to examine the skill of each combination in this paper because the high-resolution data we would require do not exist in Australia and are rare elsewhere. Our results suggest that to understand the impact of LUC on extreme temperatures will require investment in intensive field measurements that connect the land with the PBL and atmospheric convection.

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