

1 **Diagnosing the seasonal land-atmosphere correspondence over Northern Australia: Dependence**
2 **on soil moisture state and correspondence strength definition**

3

4

5

6

7

8 **Mark Decker¹ Andy Pitman¹ and Jason Evans¹**

9

10

11

12

13 [1] Climate Change Research Centre and ARC Centre of Excellence for Climate System Science,

14 University of New South Wales

15

16

17

18

19

20

21

22

23

24

25

26

27

28 Correspondence to: Mark Decker (m.decker@unsw.edu.au)

29

30 **Abstract**

31 The strength of the correspondence between the land and the atmosphere during the onset
32 (September) through to the peak (February) of the wet season over Northern Australia is statistically
33 diagnosed using ensembles of offline land surface model simulations that produce a range of different
34 background soil moisture states. We derive correspondence between the soil moisture and the
35 planetary boundary layer via a statistical measure of association. The simulated evaporative fraction
36 and the boundary layer are shown to be strongly associated during both SON and DJF despite the
37 differing background soil moisture states between the two seasons and among the ensemble members.
38 The sign and magnitude of the boundary layer-surface layer soil moisture association during the onset
39 of the wet season (SON) differs from the correspondence between the evaporative fraction and
40 boundary layer from the same season, and the correlation between the surface soil moisture and
41 boundary layer coupling during DJF. The patterns and magnitude of the surface flux-boundary layer
42 correspondence are not captured when the relationship is diagnosed using the surface layer soil
43 moisture alone. The conflicting results arise because the surface layer soil moisture lacks strong
44 association with the atmosphere during the monsoon onset because the evapotranspiration is
45 dominated by transpiration. Our results indicate that accurately diagnosing correspondence and
46 therefore coupling strength in seasonally dry regions, such as Northern Australia, requires root zone
47 soil moisture to be included.

48

49 1 Introduction

50

51 The land surface influences the atmosphere at multiple spatial and temporal scales (Pitman
52 2003; Pielke et al., 2011; Williams and Maxwell, 2011). Land-atmosphere coupling strength is the
53 degree to which land surface anomalies (e.g. soil moisture, vegetation characteristics, temperature,
54 snow cover) lead to changes in atmospheric states and fluxes (e.g. rainfall, cloud cover, moisture
55 convergence) as well as how anomalies in the atmosphere affect the land surface. The influence of
56 land surface anomalies on atmospheric anomalies (and vice versa) proceeds through a chain of non-
57 linear processes. The strength of these processes varies spatially and temporally and depend, in part,
58 on the background state of the system (Betts 2004; Koster and Suarez, 2003; Taylor and Ellis 2006).
59 The chain of mechanisms between soil moisture (SM) and precipitation (P) anomalies can be
60 summarized following Santanello et al. (2011) as

$$61 \quad \Delta SM \Rightarrow \Delta EF_{SM} \Rightarrow \Delta PBL \Rightarrow \Delta EF_{ATM} \Rightarrow \Delta CLD \Rightarrow \Delta P \quad (1)$$

62 where the changes in soil moisture (ΔSM) lead to changes in evaporative fraction (ΔEF_{SM}), which
63 alters the properties of the planetary boundary layer (ΔPBL) including the state (temperature,
64 humidity) and the entrainment rate. These three near surface coupling mechanisms (ΔSM , ΔEF_{SM} , and
65 ΔPBL) precede changes away from the land surface that further change evaporative fraction (ΔEF_{ATM}),
66 leading to changes in cloud development and growth (ΔCLD), and ultimately forcing changes in
67 precipitation (ΔP). The chain cycles with ΔP driving ΔSM to varying degrees depending on the region
68 and season (Zhang et al. 2008). Equation (1) is a conceptualization of complex and nonlinear
69 processes, such that the sign of the ΔCLD response to a ΔSM forcing can vary (Westra et al. 2012;
70 Gentine et al. 2013). Equation (1) is a simplification of the short (less than a day) timescale coupling
71 mechanisms and neglects large scale circulation and moisture feedbacks (Lee et al. 2012; Lintner and
72 Neelin 2009; Lintner et al. 2013). Additional feedbacks that operate on short timescales not shown in
73 (1), such as ΔEF_{SM} or ΔEF_{ATM} leading to ΔSM , may also be important (Seneviratne et al. 2010; Meng
74 et al. 2014a,b). Despite simplifications, Equation (1) highlights the primary control SM exerts on EF
75 as compared to secondary factors such as entrainment (Gentine et al. 2011). In a convective regime
76 ΔSM initiates a series of events that first alter the atmosphere (ΔPBL) prior to changing P. The series

77 of events from ΔSM - ΔPBL comprises the terrestrial portion of the coupling mechanisms and is the
78 focus of this study, with coupling examined here limited to these processes. The ΔSM through ΔPBL
79 sequence is a necessary, but not sufficient, set of processes that determine how P responds to changes
80 in SM.

81
82 The sensitivity of atmospheric processes to ΔSM has been quantified with observations (Koster
83 et al. 2003; Taylor and Ellis 2006) and multiple model experiments (Dirmeyer et al. 2006; Guo et al.
84 2006; Hirsch et al. 2013; Koster et al. 2000; Koster et al. 2006; Koster et al. 2011; Lee et al. 2012).
85 Ferguson et al. (2012) combined multiple sources of reanalysis data with LCL and SM observations to
86 examine the relationship between early morning surface layer SM (SM_1), and both the lifting
87 condensation level (LCL) and the EF in the afternoon during the convective season. The relationship
88 was quantified using the Kendall tau coefficient ($K\tau$), a non-parametric rank correlation coefficient
89 that measures the association between two time series. Ferguson et al. (2012) found strong coupling
90 ($K\tau$) between SM_1 -EF, EF-LCL, and SM_1 -LCL over many regions including monsoon regions such as
91 Northern Australia. These three coupling mechanisms span the first three components in Equation (1)
92 (ΔSM , ΔEF_{SM} and ΔPBL). While these represent only part of the processes involved in land-
93 atmosphere coupling, they comprise a fundamental pathway by which SM anomalies drive an
94 atmospheric response.

95
96 Several regional analyses have investigated the importance of land-atmosphere coupling in
97 Northern Australia (Evans et al. 2011). Koster et al. (2000) showed land-atmosphere coupling
98 increased the variance of P in both Northern and Eastern Australia. In agreement, Ferguson et al.
99 (2012) found high correlations in SM_1 -EF, EF-LCL, and SM_1 -LCL during the convective (monsoon)
100 season over the Northern savannas. These studies were limited in scope and did not explicitly explore
101 how the coupling behaves during periods with different background climate states.

102
103 To examine land-atmosphere coupling strength we explore the correspondence between model
104 derived soil moisture and water flux estimates with the observation based estimates of the boundary
105 layer state. Significant association between soil moisture or surface fluxes and the atmosphere

106 provides a necessary but not sufficient condition to demonstrate significant land-atmosphere coupling.
107 The lack of land-atmosphere feedbacks in offline simulations means we cannot assess cause and
108 effect, but by examining the statistical correspondence we can determine if the system states coevolve
109 in a manner consistent with strong coupling. We focus on Northern Australia to examine whether the
110 correspondence between soil moisture and the boundary layer can be diagnosed from SM_1 in regions
111 with a pronounced dry season, given the influence of groundwater on transpiration and deep SM
112 variability (Decker et al. 2013). Northern Australia has a pronounced May to September dry season
113 and a monsoon-driven wet season from November through February (Figure 1). The monsoonal
114 climate allows us to examine the SM_1 -LCL association as defined in Ferguson et al. (2012) in sharply
115 contrasting seasons (Figure 1) that exhibit contrasting background soil moisture states. By examining
116 the differences between correspondence during the onset (defined here as SON to coincide with the
117 initial increase in rainfall) of the wet season when soil moisture will be low, and then through to the
118 peak (defined as DJF to coincide with the precipitation maximum) of the wet season, we aim to
119 determine the reliability of diagnosing the terrestrial and near surface stages of land-atmosphere
120 correspondence using $K\tau$ derived from SM_1 and LCL during periods where total ET fluxes are
121 dominated by either soil evaporation or transpiration. The statistical association is defined here such
122 that the land surface processes in Equation (1) (ΔSM , ΔEF_{SM} and ΔPBL) are examined, while the
123 sequence of events in the atmosphere (ΔCLD and ΔP) are neglected. This terrestrial derived statistical
124 association captures how ΔSM relates to state changes in the mixed layer (ΔPBL). Strong association
125 as defined here is a necessary but not sufficient prerequisite for strong ΔSM - ΔP coupling. An
126 ensemble of offline simulations using two model configurations, one of which neglects groundwater
127 and therefore contains greatly reduced deep soil moisture, are driven using four forcing datasets. The
128 simulations provide estimates of SM_1 in addition to SM over the root zone (SM_{rz}), total ET and the ET
129 components. Afternoon (2 pm local time) LCL is derived using the near surface atmospheric variables
130 from the forcing datasets, and the sensitivity of the ensemble median $K\tau$ is examined for the onset and
131 peak of the monsoon season.

132

133 This manuscript is organized as follows. The model simulations, the SM_1 and ET observations
134 used for model evaluation, and the near surface atmospheric datasets are summarized in Section 2.

135 Section 3 outlines the statistical measure of association used for coupling strength, the derivation of
136 LCL from the atmospheric data, and the model experiments used to estimate the evaporative fraction
137 and soil moisture. The Results section consists of the SM₁-LCL and EF-LCL based association
138 strength, the impact of defining association strength with SM_{rrz} (the root zone SM) are presented in
139 Section 4. The results are explained in terms of the governing physical processes and previous
140 research in Section 5.

141

142 **2 Model Simulations and Data**

143 **2.1 Near Surface Atmospheric and Forcing Data**

144 The lifting condensation level (LCL see Section 3.2) over the entire study region is computed
145 from combinations of near surface atmospheric data using two reanalysis products. The LCL is also
146 calculated at the two flux sites using the tower observations. The model simulations (see Section 2.2)
147 are driven using a combination of atmospheric states and fluxes from reanalysis products, a gauge
148 based daily precipitation dataset, and a 3 hourly satellite-based precipitation product. We follow
149 Decker et al. (2014) and utilize four forcing datasets to drive model simulations.

150

151 The two gridded sources of temperature, humidity, wind speed, pressure, and radiative fluxes
152 are the Global Land Data Assimilation System (GLDAS, <http://disc.sci.gsfc.nasa.gov/hydrology/data->
153 [holdings](http://disc.sci.gsfc.nasa.gov/hydrology/data-holdings), Rodell et al., 2004) and the Modern-Era Retrospective Analysis for Research and
154 Applications (MERRA) product (Bosilovich et al., 2008). These two datasets are utilized due to the
155 high spatial resolution of GLDAS (0.25°) and high temporal resolution of MERRA (hourly). Two
156 forcing datasets are comprised of the uncorrected GLDAS and MERRA data interpolated to a common
157 0.25° x 0.25° grid. In addition two precipitation corrected datasets developed in Decker et al. (2014)
158 are used. The uncorrected atmospheric states and radiative fluxes from MERRA are combined with P
159 corrected via two algorithms. First, MERRA is corrected using the Australian Water Availability
160 Project (AWAP) daily gridded precipitation data (Jones et al. 2009) to remove the monthly biases
161 (labelled MERRA.B). Second, the MERRA precipitation is replaced with precipitation derived from
162 disaggregating the daily AWAP data with the 3 hourly Tropical Rainfall Measuring Mission (TRMM)
163 3B42 (Huffman et al. 2007) data (labelled MERRA.BT). These two corrected datasets have identical
164 monthly mean precipitation but different distributions of sub-monthly precipitation.

165

166 **2.2 Simulated estimates of Soil Moisture and Evaporative Fraction**

167 We use the community land model version 4 (CLM4, Oleson et al. 2010) to simulate the states
168 and fluxes of water and energy using configurations documented in Decker et al. (2013; 2014). The
169 land surface model simulations and reanalysis products allow for the terrestrial leg ($\Delta SM - \Delta PBL$ in
170 Equation (1)) to be explored explicitly. A detailed description of the groundwater configurations and
171 modifications are given in Decker et al. (2014).

172

173 The suite of simulations is utilized to address forcing data and model configuration
174 uncertainties in addition to exploring a large soil moisture state space. Two different configurations of
175 CLM4 are used. The first consists of the default CLM4 (referred to as CTRL). The second (referred
176 to as DRY) uses a modified CLM4 that replaces the two-way soil moisture coupling between the soil
177 column and the aquifer with a free drainage bottom boundary condition. The modifications
178 significantly reduce the soil moisture at depths below several centimeters and the ET flux during
179 periods of low rainfall while not imparting large differences on the changes in total column water
180 (Decker et al. 2014). The two model configurations thus enable the coupling between the atmosphere
181 and the land surface to be examined under two differing background soil moisture states.

182

183 The CLM4 evapotranspiration is computed as the sum of the soil evaporation, the canopy
184 evaporation and the transpiration. Transpiration is determined from the rate of photosynthesis and is,
185 in part, a function of SM. The dependence on SM is determined by the soil water potential in each
186 soil layer, the root distribution (prescribed by plant functional type, PFT), and the PFT dependence on
187 water stress. The spatial distribution and phenology of PFTs are specified and identical across all
188 simulations. The C3 grass PFT sets approximately 99% of the roots within 1m of the surface, while
189 approximately 90% of the roots are within this depth for the broadleaf evergreen forest PFT.

190

191 The experiment design follows the simulations outlined in Decker et al. (2014) that have been
192 shown to be in good agreement with observations over parts of Australia. One control (CTRL)
193 simulation and one dry simulation are equilibrated for the period 1948-1979 using the corrected
194 NCEP/NCAR data (Qian et al. 2006) after interpolating to the same $0.25^\circ \times 0.25^\circ$ grid as the other

195 forcing datasets. The CTRL and DRY simulations ending in 1979 provide initial conditions for the
196 four CTRL and four DRY simulations from 1979-2007. The model evaluation period spans the five
197 years coincident with the SM and ET data from 2003-2007. The associations are computed using the
198 period 1990-2008. Both the CTRL and the DRY simulations are forced with the four forcing datasets
199 (see Section 2.1): GLDAS, MERRA, MERRA.B, and MERRA.BT, generating a total of eight model
200 simulations. The SM (from all model layers), and turbulent energy fluxes are output at three hourly
201 intervals (coincident with the temporal resolution of the GLDAS forcing), while the remaining CLM4
202 output is saved as monthly means.

203

204 **2.3 Validation Data: Soil Moisture and Evapotranspiration**

205 The spatial-temporal behavior of the simulated surface soil moisture (SM_1) and
206 evapotranspiration (ET) are validated against gridded observationally based estimates. SM_1 is
207 evaluated against the daily Advanced Microwave Scanning Radiometer-Earth Observing System
208 (AMSR-E) L3 surface SM product. The data are derived from passive microwave measurements and
209 available for the period 2002 to 2011 (Njoku et al. 2003). AMSR-E based SM compares favorably
210 with in-situ measurements over Australia (Draper et al. 2009) and exhibits spatiotemporal variability
211 consistent with land model simulations (Liu et al. 2009). To simplify the comparison with the
212 simulated SM, the first model layer (~ 0.7 cm deep) SM is assumed comparable to SM from AMSR-E
213 despite the uncertain effective measurement depth (approximately 1 cm) that varies with SM.

214

215 The simulated evapotranspiration is evaluated against three ET products. Multiple ET datasets
216 based on different methodologies are included due to the uncertainty associated with deriving gridded
217 moisture flux data (Jimenez et al. 2011). The Global Land Evaporation Amsterdam Methodology
218 (GLEAM) (Miralles et al. 2010), the model-tree ensemble based dataset from MPI-Jena (J2010
219 hereafter) (Jung et al. 2010), and the Moderate Resolution Imaging Spectrometer (MODIS) MOD16
220 dataset (Mu et al. 2007; 2011) are used to estimate the observed mean seasonal ET fluxes. The
221 observed ET is estimated using the arithmetic mean of the three datasets after the GLEAM and
222 MOD16 data are aggregated to the coarse resolution ($0.5^\circ \times 0.5^\circ$) of the J2010 data. The simulations
223 are subsequently compared to the mean observed ET separately for the wet (December-February) and
224 the end of the dry (September-November) seasons.

225

226 In addition to the gridded SM and ET datasets the model is evaluated against observations
227 from two flux tower sites included in the OZ Flux network (ozflux.org.au). The Adelaide River site
228 (Beringer, 2013a) spans November 2007 through May 2009 and is located at 13.08°S 131.12°E. The
229 Howard Springs site (Beringer, 2013b) spans 2001 to present and is located at 12.48°S 131.15°E. Both
230 sites provide air temperature, water vapor, surface pressure, radiation, turbulent fluxes (including ET),
231 and soil moisture measurements at 30 minute intervals. The level 3 (L3) quality controlled data were
232 utilized in this study. Adelaide River provides SM data at 5cm depth while Howard Springs provides
233 SM at a depth from 10cm. The simulations are validated against the observed ET and SM at these two
234 locations.

235

236 **3 Methods**

237 **3.1 Kendall τ**

238 We evaluate the land-atmosphere coupling strength using Kendall tau ($K\tau$), a non-parametric,
239 rank correlation statistic (Press et al. 1992). Following Ferguson et al. (2012), $K\tau$ is used to indicate
240 the correspondence between two states important to land-atmosphere coupling. $K\tau$ does not assume
241 linearity between the variables being compared and tests for statistical significance. $K\tau$ ranges from -1
242 to 1 (positive values indicate the temporal variations are synchronized), with statistical significance
243 depending on the sample size (approximately 0.12 for the simulation based results in this study). $K\tau$ is
244 defined as

$$245 K_{\tau} = \frac{N_o - N_d}{0.5n(n-1)} \quad (2)$$

246 where N_o is the number of ordinate pairs, N_d is the number of disordinate pairs, and n is the number of
247 observations. Ordinate pairs are pairs of numbers for which the change between them have the same
248 sign, i.e. both are either positive or negative. The data are necessarily detrended separately over each
249 season prior to deriving $K\tau$ to prevent the strong seasonal cycle (Figures 2 and 3) from controlling the
250 statistical relationship. The spatially distributed $K\tau$ is calculated between the seasonally detrended
251 three hourly modeled SM_1 during the morning and the estimated three hourly LCL from the afternoon
252 at each grid cell for each month during both the wet and dry seasons. $K\tau$ is additionally derived with

253 detrended data at two flux tower sites using measurements of SM and LCL estimated from the tower
254 data. The morning SM_1 is utilized because SM will be highest in the morning prior to decreasing
255 during the day due to ET. The local time of SM and LCL varies because the simulations and forcing
256 data utilize GMT. The distributed $K\tau$ is found separately for each of the eight simulated (See Section
257 2.2) estimates of SM_1 and the four estimates of LCL (Section 3.2), generating a total of 32 estimates of
258 $K\tau$ for each month in both the wet and dry seasons. The median $K\tau$ is found separately for the wet and
259 dry seasons for the two different model configurations (Section 3.3) to give the final estimation of the
260 correspondence. The association is also diagnosed using $K\tau$ between the model simulated afternoon
261 evaporative fraction and the afternoon LCL. A second definition of association is found by calculating
262 $K\tau$ between the morning time root zone SM (SM_{rz}) and the afternoon LCL (SM_{rz} -LCL). SM_{rz} is
263 defined as the vertically averaged SM from the surface to a depth of 1m.

264 The physical meaning of a negative SM-LCL $K\tau$ association is as follows. A high value of SM
265 will cause a larger ET flux, moistening the lower atmosphere, causing a lower LCL. Thus we
266 hypothesis that in regions where the land-atmosphere are coupled the SM-LCL $K\tau$ should be negative.
267 If SM has no association with LCL than $K\tau$ is expected to be statistically insignificant. Similarly, if ET
268 is negatively associated with LCL ($K\tau < 0$), it means that high ET may be moistening the lower
269 atmosphere again leading to a lower LCL.

270

271 **3.2 Calculation of Lifting Condensation Level**

272 The state of the convective atmosphere is evaluated using the lifting condensation level (LCL),
273 defined as the height (in pressure) that a parcel reaches saturation when ascending adiabatically from
274 the surface. While a lower LCL is favorable to convection, it is not a sufficient constraint to guarantee
275 it. For convection to occur a parcel must reach the level of free convection (LFC), which may not
276 occur even if a parcel reaches the LCL. The height (in pressure) of the LCL is derived using only near
277 surface variables under the assumption that the boundary layer is well developed and therefore well
278 mixed. Estimating the LCL from near surface variables provides heights comparable to direct
279 observations (Ferguson and Wood 2009). Under these assumptions, the pressure at the LCL is given
280 by

$$LCL = P_{srf} - P_{srf} \left(\frac{T_{air}}{T_{dew}} \right)^{\frac{-c_p}{R}} \quad (3)$$

281
 282 where P_{srf} is the surface pressure (Pa), T_{air} is the near surface air temperature (K), T_{dew} is the near
 283 surface dew point temperature (K), R is the specific gas constant of dry air ($J K^{-1} kg^{-1}$), and c_p is the
 284 specific heat of dry air at constant pressure ($J K^{-1} kg^{-1}$). Four spatially explicit estimates of LCL are
 285 found by applying Equation (4) to several combinations of near surface forcing data, and two point
 286 wise estimated are derived from the flux tower data. While P_{srf} and T_{air} are directly provided by both
 287 reanalysis products and the tower measurements, T_{dew} is calculated using the available near surface
 288 atmospheric states. The four distributed estimates of LCL are calculated with Equation (4) by: 1)
 289 using GLDAS for pressure and both temperatures, 2) using MERRA for pressure and both
 290 temperatures, 3) using pressure from MERRA and temperatures from GLDAS, and 4) using pressure
 291 from GLDAS and temperatures from MERRA. The LCL is quality controlled by limiting LCL to be
 292 less than the surface pressure.

293

294

295 4. Results

296 4.1 Validation of Simulated Soil Moisture and Evapotranspiration

297 The two model configurations are separately validated against the observationally estimated
 298 soil moisture and evapotranspiration on monthly and seasonal timescales, respectively. Figure 2a
 299 shows the timeseries of the area averaged (10-15S to 120-150E) normalized ensemble mean first layer
 300 soil moisture from the CTRL and the DRY ensembles and the AMSR-E observed data. The simulation
 301 dynamics are evaluated using the normalized SM_1 due to the difficulties in direct comparison of
 302 simulated and observed soil moisture (Koster et al. 2009). The strong seasonal cycle of soil moisture
 303 owing to the monsoonal climate is evident in both the observationally based estimates and the
 304 simulations. CTRL and DRY are nearly identical aside from the dry season in 2005 where the soil
 305 moisture in CTRL decreases more than that from DRY. The observed moistening of the soil following
 306 the dry seasons in Figure 2a occurs within a month to that of the simulated moistening. The mean
 307 monthly soil moisture closely follows that of the observationally based estimates and exhibits dynamic
 308 behavior independent of the model configuration.

309

310 The bias of the ensemble mean time averaged surface layer soil moisture from the eight
311 simulations against the AMSR-E product is shown in Figure 2b. Over large regions of Northern
312 Australia, the simulated SM_1 is within $0.025 \text{ mm}^3 \text{ mm}^{-3}$ of AMSR-E. The difference in mean SM_1
313 between the two model configurations is similarly small (figure not shown). Figure 2 demonstrates
314 that the temporal evolution (Figure 2a) and mean state (Figure 2b) of the simulated SM_1 are similar to
315 the AMSR-E estimates.

316

317 The seasonal mean ET is validated against the arithmetic mean of the three gridded ET
318 products for both DJF (Figures 3a, 3c, and 3e) and SON (Figures 3b, 3d, and 3f). The observed DJF
319 ET (Figure 3e) has a strong north-south gradient with a maxima centered around 13°S - 130°E . The
320 strong north-south gradient is also present in the ensemble mean ET (Figure 3a), however the
321 simulations overestimate DJF ET over much of the domain. The observationally based estimates
322 show an ET of less than 50 Wm^{-2} south of 18°S while the simulations remain above 60 Wm^{-2} in this
323 region. The mean SON ET is markedly lower compared to DJF ET in both the gridded data (Figure
324 3f) and the simulations (Figure 3b). Similar to DJF, both the model and the ET product show a strong
325 north-south gradient. The simulations underestimate the ET in the York Peninsula (East of 140°E and
326 North of 17°S) during SON and overestimate the ET in this region during DJF. The overestimation of
327 DJF ET compared to the gridded product is much more pronounced for the CTRL simulations (Figure
328 3a) than the DRY simulations (Figure 3c). The underestimation of the SON ET in the simulations is
329 largely a result of including the DRY model configuration. The CTRL simulations exhibit a 10-20
330 Wm^{-2} increase in SON ET over the DRY model runs (Figure 3b and Figure 3d). Overall, the model
331 exhibits spatial-temporal ET in close agreement with this gridded ET product.

332

333 Point measurements of SM and ET at two locations show reasonable agreement with the model
334 simulations. The Howard Springs SM observations 10cm depth (Figure 4a) typically increases from
335 0.05 to $0.2 \text{ mm}^3 \text{ mm}^{-3}$ from the dry to the wet season. The observations are drier during the wet season
336 and have a smaller (by a factor of two) seasonal cycle than both the DRY and CTRL simulations.
337 DRY is much drier ($\sim 0.08 \text{ mm}^3 \text{ mm}^{-3}$) than CTRL ($\sim 0.18 \text{ mm}^3 \text{ mm}^{-3}$) during the dry season and in
338 better agreement with the measurements ($\sim 0.05 \text{ mm}^3 \text{ mm}^{-3}$). This contrasts with the agreement at the

339 Adelaide River site (Figure 4b) where the measurements and CTRL peak around $0.30 \text{ mm}^3\text{mm}^{-3}$
340 during the 2008 wet season. DRY ($0.02\text{-}0.07 \text{ mm}^3\text{mm}^{-3}$) is again much drier than CTRL (0.15
341 $\text{mm}^3\text{mm}^{-3}$) during the 2008 dry season but CTRL is in better agreement with the data ($0.15 \text{ mm}^3\text{mm}^{-3}$
342 3). The AMSR-E estimate, CTRL, and DRY are similar in Figures 4a and 4b (the Y-axis scale is the
343 same in both figures), while the SM observations at the two sites differ drastically. The disagreement
344 in the mean as well as the amplitude of the seasonal variability is likely due to both the difference in
345 scale between the measurements and simulations and poor representation of soil properties in the
346 model. When the SM comparison is normalized using the first two moments as in Figure 2a (not
347 shown) there is greater agreement between the measurements, AMSR-E, and the simulations.

348

349 The ET data at Howard Springs (Figure 4c) demonstrates that the CTRL simulation always
350 produces too little ET during the dry season. While the gridded ET estimate in Figure 4c falls within
351 10 Wm^{-2} of the CTRL simulation during the dry season, the tower data are nearly 20 Wm^{-2} greater than
352 both during the 2007 and 2008 dry seasons. The wet season peak in ET is well simulated by both
353 CTRL and DRY at Howard Springs. The model performance is different at Adelaide river as both
354 CTRL and DRY have a wet season peak ET of around 120 Wm^{-2} while the measurements peak closer
355 to 150 Wm^{-2} . Figure 4d further demonstrates that DRY has too little dry season ET.

356

357 The results from Figures 2, 3 and 4 demonstrate that CLM4 simulates the monthly and
358 seasonal first layer soil moisture and evapotranspiration reasonably. While the details of the model
359 performance vary depending on which site, season, and ensemble member is used for validation,
360 overall the spatial and temporal patterns of ET and SM are generally captured by the modeling system.
361 The accuracy of the estimated land surface states and fluxes therefore enables the use of the simulated
362 variables in the diagnoses of the land-atmosphere association strength during SON and DJF.

363

364 **4.2 Background SM state**

365 The sharp contrast in background SM state can be illustrated by taking a spatial-temporal
366 average of SM as a function of depth for CTRL and DRY for DJF (Figure 5a) and SON (Figure 5b).
367 The soil moisture away from the surface is markedly different between CTRL and DRY. During DJF,
368 CTRL shows a slight increase in soil moisture with depth, reaching a peak of $\sim 0.35 \text{ mm}^3\text{mm}^{-3}$ at

369 depths near 3 m. In contrast, DRY has a peak soil moisture of only $\sim 0.24 \text{ mm}^3 \text{ mm}^{-3}$ at the surface and
370 decreases with depth to near zero at 3 m. Similar patterns of SM with depth are seen over SON,
371 however SM_1 is considerably lower for both CTRL and DRY compared to DJF.

372
373 Despite the similar mean and temporal behavior of SM_1 shown in Figure 2, SM away from the
374 surface differs substantially between the two model configurations (Figure 3). The mean DJF ET is
375 similar between CTRL and DRY, with differences between the two only $10\text{-}20 \text{ Wm}^{-2}$, corresponding to
376 roughly 10-20% of the mean value. The fractional contribution of transpiration to the total ET during
377 DJF is roughly 10-30% for both DRY and CTRL (figure not shown) indicating that the evaporation is
378 the dominant ET mechanism. The enhanced mean SM in CTRL causes the CTRL ET to be greater
379 than the DRY ET during DJF, yet both compare reasonable well to the observationally based estimates
380 (Figure 3). However the lack of SM at depths below several centimeters for DRY during SON causes
381 the reduced ET as compared to CTRL during this period. The mean ET during SON is sensitive to the
382 mean SM away from the surface, indicating that transpiration significantly contributes to the total ET
383 during this period. The reduced root zone SM in DRY leads to an increase in water stress and reduced
384 transpiration. This reduction during SON is large relative to the mean ET during the period (Figure 3).

385

386 **4.3 Correspondence: EF-LCL and SM_1 -LCL**

387 The statistical association between the evaporative fraction and the LCL is shown in Figure 6,
388 with the results from the two flux towers shown in enclosed squares around 13°N and 131°E . The
389 insignificant associations are greyed out while the statistically significant results are shown in color.
390 During DJF, CTRL (Figure 6a) and DRY (Figure 6c) exhibit strong surface flux-atmosphere
391 correspondence, with the strongest association over the Cape York Peninsula (East of 140°E and North
392 of 17°S) and the Southwestern part of the domain. Similarly, the EF-LCL coupling is significant
393 during SON (Figures 6b and 6d) over much of the domain, although the magnitude is reduced relative
394 to DJF. Both ensembles show strong associations independent of the season, however the differences
395 between CTRL and DRY vary with season. The DJF EF-LCL correspondence near 15°S 132°E is
396 statistically significant in DRY (Figure 6c) but not in CTRL (Figure 6a), contrasting the similar SON
397 EF-LCL association in this region exhibited by both DRY (Figure 6d) and CTRL (Figure 6b). The
398 flux towers (boxed squares in Figures 6a-6c) both show statistically significant association between

399 EF and the LCL during both seasons. The EF-LCL correspondence from the tower observations agree
400 more closely with DRY in DJF as CTRL show statistically insignificant association in the region (13°S
401 131°E). The reduced deep layer soil moisture resulting from the removal of the groundwater module
402 enhances the DJF correspondence in agreement with the tower data.

403
404 Figure 7 shows the median Kendall tau ($K\tau$) between SM_1 and the LCL (see Section 3.3) for
405 CTRL and DRY separately during DJF (Figures 7a and 7c) and SON (Figures 7b and 7d). Several
406 important features are present in Figure 7. The SM_1 -LCL association during DJF and SON is largely
407 similar between the two model configurations. CTRL (Figure 7a) and DRY (Figure 7c) exhibit similar
408 spatial patterns and magnitudes of $K\tau$. Some regions (17°S 126°E) exhibit increases in the magnitude
409 of $K\tau$ in CTRL relative to DRY in DJF (Figures 7a and 7c) although the differences are statistically
410 insignificant over most of the domain. Regardless of these slight variations in $K\tau$, CTRL and DRY
411 exhibit a strong association between SM_1 and the boundary layer during the peak of the wet season
412 over coincident parts of the domain. Both model configurations also show areas (15°S 131°E) with
413 insignificant correspondence adjacent to the strongly associated regions. In contrast, CTRL and DRY
414 both contain regions of significant positive $K\tau$ demonstrating a negative correspondence during SON,
415 in disagreement with the results from the Adelaide River tower site. The tower sites both show
416 statistically significant negative SM-LCL association during DJF adjacent to a region (15°S 131°E) of
417 insignificant correspondence both simulations. The similarity in SM_1 -LCL correspondence between
418 CTRL and DRY during both DJF and SON implies a similar temporal variability of SM_1 as related to
419 the LCL. From Figure 3, the mean ET fluxes are considerably different during SON. The similar
420 temporal behavior relative to the LCL for both DRY and CTRL indicates that the SM_1 variability is
421 physically independent of the season mean ET fluxes.

422
423 Contrasting Figures 6 and 7 reveals that the surface fluxes (Figures 6b and 6d) are associated
424 with the LCL despite the simulated surface layer soil moisture (Figures 7b and 7d) lacking similar
425 correspondence. The regions of positive $K\tau$ in Figure 7 contradict the strongly negative $K\tau$ in Figure
426 6 during SON. The flux towers show negative association for both EF-LCL and SM-LCL during DJF
427 and SON in Figures 6 and 7. The EF-LCL coupling during DJF is much stronger than the SM_1

428 coupling, and DRY exhibits regions of stronger EF-LCL correspondence than CTRL, however the
429 differences are not statistically significant over much of the domain. A key difference between the
430 flux tower and model simulation estimated $K\tau$ is the depth of the SM. The measurement depth at the
431 tower sites are 5cm and 10cm for Adelaide River and Howard Springs respectively, while the model
432 surface layer soil moisture is taken from a depth of 0.7 cm. The change in sign of SM_1 -LCL $K\tau$ from
433 SON (Figures 7b and 7d) to DJF (Figures 7a and 7c) demonstrates that applying Equation (4) to SM_1
434 and the LCL doesn't always capture the coupling between the land and the atmosphere during periods
435 where deep SM and transpiration dominate the ET flux.

436

437 In short, our results demonstrate that the simulated surface layer soil moisture cannot adequately
438 capture the SM-LCL association during both DJF and SON. The significant contributions of
439 transpiration to the total ET fluxes (especially during SON) are responsive to perturbations in SM_{rz} and
440 not SM_1 .

441

442 **4.6 Proposed Association strength definition: SM_{rz} -LCL**

443 The definition of land-atmosphere coupling using land surface moisture states and fluxes must
444 encompass the relevant physical mechanisms. Previously, Ferguson et al. (2012) was limited to using
445 SM_1 in deriving $K\tau$ because the AMSR-E (or other microwave) SM measurements typically measure
446 to depths less than a few centimeters beneath the soil surface. Computing $K\tau$ between SM_1 and the
447 LCL incorporates the surface layer soil moisture that is important for surface evaporation from the
448 soil. Therefore the DJF coupling (or other periods where the ET is largely comprised of soil
449 evaporation) should be adequately defined using SM_1 . $K\tau$ computed from SM_1 neglects SM_{rz}
450 variations that drive transpiration during the initial increase in precipitation following the dry season
451 and therefore may not fully encompass the extent of land-atmosphere associations. Acknowledging
452 the importance of transpiration during the Northern Australian wet season, we further evaluate the
453 land-atmosphere association by computing $K\tau$ between the vertically averaged SM_{rz} and the LCL. As
454 opposed to remotely sensed SM from AMSR-E (or other satellite products), the use of simulated SM
455 facilitates the estimation of SM_{rz} . Applying Equation (4) using SM_{rz} imposes a different set of
456 problems, as the rooting depth is model dependent and generally only approximately known. There is

457 substantial evidence that eucalypts have rooting depths exceeding 20 meters (Schenk and Jackson
458 2002), however neither CLM4 nor the direct observations in this study extend that deep. Due to these
459 limitations, SM_{rz} is computed as the weighted mean of the SM observations at 10, 40, and 100cm for
460 the Howard Springs site. We assume that the SM_{rz} consists of the soil layers between the surface and a
461 depth of 1m, as greater than 90% of the prescribed roots in CLM4 are within 1m of the surface
462 (Oleson et al. 2010). This assumed rooting depth is consistent with the model formulation but not
463 realistic given the rooting depths of eucalypts.

464

465 Figure 8 shows the ensemble median coupling strength diagnosed between SM_{rz} and the LCL.
466 Comparing Figures 7 and 8 it is clear that including the portion of SM that partially controls
467 transpiration increases the magnitude of the DJF SM-LCL associations and eliminates the region near
468 14°S 131°E with statistically insignificant correspondence (Figures 7a and 7c) despite soil evaporation
469 contributing significantly to the simulated ET. Large differences between the SON SM_{rz} -LCL and
470 SM_1 -LCL $K\tau$ are seen south of 15°S and east of 130°E. Despite large regions of statistically
471 significant SON SM_{rz} -LCL $K\tau$ for CTRL (Figure 8b) and DRY (Figure 8d) regions of insignificant
472 association are prevalent near 13°S 131°E. The flux tower derived SON SM_{rz} -LCL correspondence is
473 insignificant in agreement with the DRY and CTRL results near 13°S 131°E. The similarity between
474 the DRY and CTRL SM_{rz} -LCL $K\tau$ highlights the negligible groundwater impact (Figures 8b and 8d).
475 Comparing Figures 8b and 8d with Figure 3b and 3d reveals that despite the impact of groundwater on
476 the mean ET flux over SON, the mean state of the deep SM imparts little influence on the temporal
477 dynamics of SM_{rz} in relation to the LCL. Neglecting the SM beneath the surface layer in the
478 calculation of $K\tau$ results in a weak diagnosed SM-LCL association during SON because transpiration
479 is partly governed by the water availability within the root zone. By defining the association using
480 SM_{rz} it is clear that the land is strongly linked to the LCL during both DJF and SON. The DJF SM-
481 LCL association in CTRL near flux tower sites is stronger when defined in this manner, although both
482 sets of simulations still show SM_{rz} to be statistically associated to the LCL.

483

484 The SM_1 -LCL, and SM_{rz} -LCL $K\tau$ shown in Figures 7, and 8 are the median from ensembles
485 with 32 estimates. The ensembles explicitly use multiple constructions of LCL to sample the possible

486 range atmospheric states given the near surface MERRA and GLDAS estimates, and may lead to
487 uncertain estimates of $K\tau$. The inter-ensemble uncertainty of the $K\tau$ metric is examined to
488 demonstrate the robustness of the results. The standard deviation of the coupling strength between
489 SM_{rz} and the LCL and between SM_1 and the LCL among the ensemble members is generally less than
490 0.15 (Figures 8a-8d). The variation among the ensemble members is smaller than the median coupling
491 strength shown in Figures 7 and 8. The low standard deviation relative to the median demonstrates
492 that the association shown in Figures 7 and 8 is robust in that greater than 83% of the $K\tau$ estimates
493 (assuming they are normally distributed) have a correspondence of the same sign reported in the
494 figures. The coupling using SM_1 (Figure 8c) show larger ensemble uncertainty near the coast centered
495 around 135E compared to the SM_{rz} coupling in DJF (Figure 8a) and over the Cape York Peninsula in
496 SON (Figures 8b and 8a). Aside from the region near 15S and 130E during SON, the larger ensemble
497 uncertainty is found when using SM_1 to define the coupling strength.

498

499 **5 Discussion**

500 The seasonal ET from CTRL, DRY, and the gridded ET products from DJF through SON
501 provide insight into the mechanisms that limit the SON DRY ET. The ET from CTRL and DRY are
502 similar (within +/- 10%) during the large DJF precipitation forcing. The dry season commences
503 between MAM and JJA (Figure 1) resulting in increased vapor pressure deficit (VPD) between the
504 vegetation and the atmosphere and increased photosynthetically active radiation (PAR). The changes
505 in VPD and PAR promote increased transpiration from DJF through MAM, although the actual
506 transpiration is also governed by SM_{rz} . Comparing Figure 3 and Figure 5 indicates that the DRY ET is
507 relatively SM limited and unable to maintain ET similar in magnitude to CTRL and the
508 observationally based estimates during SON. Within the model, the soil column-groundwater
509 interactions parameterized in CTRL inhibit the large, ET limiting SM_{rz} reduction present in DRY. In
510 reality the inability of DRY to maintain ET during SON may result from the shallow rooting depths
511 assumed in CLM4. The depths are substantially shallower than the rooting depths of eucalypts.
512 Realistic rooting depth profiles reaching nearly 20 meters in Australia (Schenk and Jackson 2002) and
513 corresponding soil layer depths may negate the impact of the parameterized soil column-groundwater
514 impacts current in CLM4.

515

516 The EF-LCL association (Figure 6) is similar for both model configurations despite the mean
517 ET (Figure 3) and SM (Figure 5) differing considerably between CTRL and DRY. The similarity
518 holds for both DJF and SON despite the differing background soil moisture states between the two
519 periods. The results indicate that while the mean ET is a strong function of mean soil moisture, the
520 SM-LCL coupling as diagnosed here is insensitive to the background state. The coincidence of the
521 temporal variations in SM, EF, and LCL are demonstrated by the large values of $K\tau$. These
522 seemingly counterintuitive results may be an artifact of using a rank correlation coefficient to
523 determine the strength of the correspondence. $K\tau$ only measures the temporal coincidence of the two
524 time-series while neglecting the magnitude of these variations. Although $K\tau$ is largely independent to
525 the background soil moisture state, alternative definitions of association may not remain as invariant.
526

527 While association in Figure 6 is largely unaffected by the mean SM state, the mean ET flux is
528 largely derived from deeper SM through transpiration during the onset of the wet season prior to DJF.
529 Correspondence under these conditions is poorly defined using SM_1 . The strong EF-LCL coupling
530 during SON and DJF highlights the inadequacy of coupling diagnosed with SM_1 . The physically
531 improbable SM_1 -LCL coupling varies from positive (Figure 5b) to negative (Figure 5a) as the wet
532 season is established (Figure 1). Despite the domain mean precipitation increasing from roughly zero
533 to several mm/day during SON, $K\tau$ from SM_1 -LCL exhibits both positive (i.e. 15°S 126°E) and
534 negative (i.e. 15°S 134°E) coupling over this period. The transition from negligible (or positive) to
535 strong land-atmosphere association during the wet season is an artifact resulting from the use of SM_1 .
536 More consistent correspondence in general agreement with the EF-LCL dynamics throughout the wet
537 season exists between SM_{rz} -LCL because transpiration is incorporated into the coupling diagnostic.
538 During SON, the dry surface layer SM is responsible for little ET, with variations in ET not associated
539 with SM_1 . The SM_{rz} -LCL $K\tau$ eliminates the insignificant association around 17°S 128°E exhibited in
540 the SM_1 -LCL association. Despite regions of significant SM_{rz} -LCL association in DRY and CTRL, the
541 Howard Springs data show insignificant SM_{rz} -LCL correspondence during SON. The lack of observed
542 association is possibly related to the inability to sample SM at depths that correspond to the physically
543 relevant rooting depths. The necessity of using SM_{RZ} agrees with Lee et al. (2012) where transpiration
544 was found to limit precipitation variability over tropical regions. The importance of transpiration

545 among the various ET components is not limited to Northern Australia or monsoon regions (Coenders-
546 Gerrits et al. 2014; Schlesinger and Jasechko 2014) highlighting the need to characterize land-
547 atmosphere dynamics using SM well beneath the surface.

548
549 Statistically determining the association using only near surface variables from land surface
550 model simulations of SM and atmospheric data as done in this study (i.e. Ferguson et al. 2013, Betts
551 2004) is limited due to only examining a part of the full land-atmosphere coupling processes. While
552 the LCL is an important determinant in the formation of precipitation, moisture convergence, upper
553 level inversions, convective available potential energy, wind shear, and many other factors play
554 important roles in the formation of convection. The correspondence diagnosed in this study with
555 Equation (4) is by definition limited in scope to only part of the coupling continuum described in
556 Equation (1). Therefore association defined using these methods provides a necessary but not
557 sufficient condition for strong land-atmosphere interactions between soil moisture and precipitation.

558
559 Our results likely extend to monsoonal regions beyond Northern Australia. GLACE (Koster
560 et al. 2006) revealed multiple areas of strong land-atmosphere coupling coincide with major monsoon
561 systems. The strong coupling during the wet season (September-February) diagnosed using SM_{rz} and
562 $K\tau$ in our results qualitatively agrees with the strong coupling in monsoon regions from GLACE. The
563 dry season antecedent to the large precipitation fluxes induces low evaporation while allowing deeply
564 rooted plants to transpire despite the prolonged dry period. These conditions over Northern Australia
565 (Figures 3 and 4) lead to transpiration dominating the ET flux during the onset of the wet season.
566 Coupling diagnosed using $K\tau$ under these conditions must be defined using SM_{rz} rather than SM_1 to
567 ensure the relevant pathways of the moisture fluxes are not neglected.

568

569 **6 Conclusions**

570 The land-atmosphere coupling strength is analyzed utilizing ensembles of land surface
571 simulations and near surface atmospheric data. Using four forcing datasets, ensembles of CLM
572 simulations over Northern Australia are performed, using configurations that either include or neglect
573 soil column-groundwater interactions. The seasonal dynamics of the simulated SM_1 is insensitive to
574 the mean soil moisture state and all simulations compare favorably with the AMSR-E soil moisture

575 product. Further, the simulated ET from December to February is similar between the CTRL and DRY
576 runs, with both configurations largely consistent with the DJF ET estimated from three gridded ET
577 products.

578

579 The strength of the land-atmosphere association is diagnosed between both SM_1 and EF from
580 the simulations and the LCL as calculated from the near surface atmospheric state. During the peak
581 wet season strong SM_1 -LCL and EF-LCL associations are shown. The wet season onset (SON) shows
582 strong EF-LCL association that contrasts the weak SM_1 -LCL association demonstrating the SON
583 coupling is not properly characterized with SM_1 . The land-atmosphere interactions during periods
584 with non-negligible transpiration necessitates the use of root zone soil moisture instead of the surface
585 soil moisture to properly capture the physical processes. Properly defining the association with SM_{rz}
586 differs considerably from the SM_1 diagnosed association and shows strong associations throughout the
587 wet season. The SM_{rz} -LCL association is consistent with that from EF-LCL. It should be noted that
588 these associations are a necessary but not sufficient condition to diagnose full land-atmosphere
589 coupling.

590

591 Our results also show that the diagnosed land-atmosphere coupling is insensitive to the mean
592 vertical profile of soil moisture. It is however, sensitive to the depth of the soil column considered.
593 The implication of our findings therefore indicates a need to include the root zone in order capture
594 periods when the ET is dominated by transpiration. We recommend that future studies of land-
595 atmosphere coupling should include groundwater and focus on root zone soil moisture rather than
596 surface layer soil moisture.

597

598

599 **Acknowledgments**

600 We acknowledge support from the Australian Research Council Super Science scheme
601 (FS100100054). AJP was also supported by the Australian Research Council Centre of Excellence for
602 Climate System Science (CE110001028). JPE was supported by an Australian Research Council
603 Future Fellowship (FT110100576). The GLDAS data used in this study were acquired as part of the
604 NASA's Earth-Sun System Division and archived and distributed by the Goddard Earth Sciences

605 (GES) Data and Information Services Center (DISC) Distributed Active Archive Center (DAAC). We
606 also thank Diego Miralles for providing the GLEAM ET product.

607

608

609 **References**

610 Betts, A. K. 2009: Land-Surface-Atmosphere Coupling in Observations and Models. *J. Adv. Model.*
611 *Earth Syst.*, **1 (3)**, doi:10.3894/JAMES.2009.1.4.

612
613 Betts, A. K., J. H. Ball, A. C. M. Beljaars, M. J. Miller, and P. A. Viterbo. 1996: The Land Surface-
614 atmosphere Interaction: A Review Based on Observational and Global Modeling Perspectives, *J.*
615 *Geophys. Res-Atmos.*, **101 (D3)**, 7209–7225. doi:10.1029/95JD02135.

616
617 Betts, A. K., B. Helliker, and J. Berry, 2004: Coupling Between CO₂, Water Vapor, Temperature, and
618 Radon and Their Fluxes in an Idealized Equilibrium Boundary Layer over Land, *J. Geophys. Res.-*
619 *Atmos.*, **109 (D18)**, doi:10.1029/2003JD004420.

620
621 Beringer, J., 2013a: Adelaide River OzFlux tower site OzFlux: Australian and New Zealand Flux
622 Research and Monitoring hdl: 102.100.100/14228

623
624 Beringer, J., 2013b: Howard Springs OzFlux tower site. OzFlux: Australian and New Zealand Flux
625 Research and Monitoring. hdl: 102.100.100/14234

626
627 Bosilovich, M. G., J. Chen, F. R. Robertson, and R. F. Adler, 2008: Evaluation of Global Precipitation
628 in Reanalyses, *J. Appl. Meteor. Climatol.*, **47 (9)**, 2279–2299.

629
630 Coenders-Gerrits, A. M. J., van der Ent, R. J., Bogaard, T. A., Wang-Erlandsson, L.,
631 Hra-chowitz, M., and Savenije, H. H. G., 2014: Uncertainties in transpiration estimates,
632 *Nature*, **506**, doi:10.1038/nature12925.

633
634 Decker, M., and X. Zeng. 2009: Impact of Modified Richards Equation on Global Soil Moisture
635 Simulation in the Community Land Model (CLM3.5), *J. Adv. Model. Earth Syst.*, **1 (3)**,
636 doi:10.3894/JAMES.2009.1.5.

637
638 Decker M., A. Pitman, J. P. Evans, 2013: Groundwater constraints on simulated transpiration

639 variability over Southeastern Australian forests, *J. Hydrometeor.*, **14**, 543–559.
640

641 Decker M., A. Pitman, J. P. Evans, 2014: Applying scaled vegetation greenness metrics to constrain
642 simulated transpiration anomalies: A study over Australia, *J. Hydrometeor.*, **15**, 1607–1623.
643

644 Dirmeyer, Paul A., Xiang Gao, Mei Zhao, Zhichang Guo, Taikan Oki, Naota Hanasaki, 2006: GSWP-
645 2: Multimodel Analysis and Implications for Our Perception of the Land Surface, *Bull. Amer. Meteor.*
646 *Soc.*, **87**, 1381–1397, doi: <http://dx.doi.org/10.1175/BAMS-87-10-1381>
647

648 Draper, C. S., J. P. Walker, P. J. Steinle, R. A.M. de Jeu, and T. R.H. Holmes, 2009: An Evaluation of
649 AMSR-E Derived Soil Moisture over Australia, *Remote Sens. Environ.*, **113 (4)**, 703–710.
650

651 Eamus, D., and R. Froend. 2006: Groundwater-dependent Ecosystems: The Where, What and Why of
652 GDEs, *Aust. J. Bot.*, **54 (2)**, 91–96.
653

654 Evans, J. P., A. J. Pitman, and F. T. Cruz, 2011: Coupled atmospheric and land surface dynamics over
655 southeast Australia: A review, analysis and identification of future research priorities. *International*
656 *Journal of Climatology*, **31**, 1758–1772, doi:10.1002/joc.2206.
657

658 Fan, Y., G. Miguez-Macho, C. P. Weaver, R. Walko, and A. Robock, 2007: Incorporating Water Table
659 Dynamics in Climate Modeling: 1. Water Table Observations and Equilibrium Water Table
660 Simulations, *J. Geophys. Res.-Atmos.*, **112 (D10)**, doi:10.1029/2006JD008111.
661

662 Ferguson, C. R., and E. F. Wood, 2009: Observing land–atmosphere interaction globally with satellite
663 remote sensing, Proc. Earth Observation and Water Cycle Science Symp., Frascati, Italy, European
664 Space Agency, P17.
665

666 Ferguson, C. R., E. F. Wood, and R. K. Vinukollu, 2012: A Global Intercomparison of Modeled and
667 Observed Land-Atmosphere Coupling, *J. Hydrometeor.*, **13 (3)**, 749–784, doi:10.1175/JHM-D-11-
668 0119.1.

669
670 Gedney, N., and P. M. Cox, 2003: The Sensitivity of Global Climate Model Simulations to the
671 Representation of Soil Moisture Heterogeneity, *J. Hydrometeor.*, **4** (6), 1265–1275.
672
673 Gentine, P., D. Entekhabi, and J. Polcher, 2011: The Diurnal Behavior of Evaporative Fraction in the
674 Soil–Vegetation–Atmospheric Boundary Layer Continuum, *J. Hydrometeor.*, **12**, 1530-1546.
675
676 Gentine, P., A. A. M. Holtslag, F. D'Andrea, and M. Ek, 2013: Surface and Atmospheric Controls on
677 the Onset of Moist Convection over Land. *J. Hydrometeor.*, **14**, 1443–1462.
678
679 Guo, Z., and Coauthors, 2006: GLACE: The Global Land–Atmosphere Coupling Experiment. Part II:
680 Analysis. *J. Hydrometeor.*, **7**, 611–625, doi: <http://dx.doi.org/10.1175/JHM511.1>.
681
682 Hirsch, A., J. Kala, A. J. Pitman, C. Carouge, J. P. Evans, V. Haverd, and D. Mocko, 2013: Impact of
683 Land Surface Initialisation Approach on Sub-seasonal Forecast Skill: A Regional Analysis in the
684 Southern Hemisphere, *J. Hydrometeor.*, under review.
685
686 Huffman, G. J., D. T. Bolvin, E. J. Nelkin, D. B. Wolff, R. F. Adler, G. Gu, Y. Hong, K. P. Bowman,
687 and E. F. Stocker, 2007: The TRMM Multisatellite Precipitation Analysis (TMPA): Quasi-Global,
688 Multiyear, Combined-Sensor Precipitation Estimates at Fine Scales, *J. Hydrometeor.*, **8** (1), 38–55.
689 doi:10.1175/JHM560.1.
690
691 Jimenez, C., and Coauthors, 2011: Global Intercomparison of 12 Land Surface Heat Flux Estimates, *J.*
692 *Geophys. Res.*, **116**, D02102, doi:[10.1029/2010JD014545](http://dx.doi.org/10.1029/2010JD014545).
693
694 Jones, D. A., W. Wang, and R. Fawcett, 2009: High-quality spatial climate data-sets for Australia, *J.*
695 *Aust. Meteor. Oceanogr.*, **58**, 233-248.
696
697 Jung, M., and Coauthors, 2010: Recent Decline in the Global Land Evapotranspiration Trend Due to
698 Limited Moisture Supply, *Nature*, **467**, 951–954, doi:10.1038/nature09396.

699

700 Koster, R. D., M. J. Suarez, and M. Heiser. 2000: Variance and Predictability of Precipitation at
701 Seasonal-to-Interannual Timescales, *J. Hydrometeor.*, **1** (1), 26–46.

702

703 Koster, R. D., and M. J. Suarez, 2003: Impact of Land Surface Initialization on Seasonal Precipitation
704 and Temperature Prediction, *J. Hydrometeor.*, **4** (2), 408–423.

705

706 Koster, R. D., and Coauthors, 2006: GLACE: The Global Land–Atmosphere Coupling Experiment.
707 Part I: Overview. *J. Hydrometeor.*, **7**, 590–610.

708

709 Koster, R. D., Z. Guo, R. Yang, P. A. Dirmeyer, K. Mitchell, and M. J. Puma, 2009: On the Nature of
710 Soil Moisture in Land Surface Models, *J. Clim.*, **22**, 4322–4335.

711

712 Koster, R. D., and Coauthors, 2011: The Second Phase of the Global Land-Atmosphere Coupling
713 Experiment: Soil Moisture Contributions to Subseasonal Forecast Skill, *J. Hydrometeor.*, **12** (5), 805–
714 822. doi:10.1175/2011JHM1365.1.

715

716 Lee, J.-E., and Coauthors, 2012: Reduction of tropical land region precipitation variability via
717 transpiration, *Geophys. Res. Lett.*, **39**, L19704, doi:10.1029/2012GL053417.

718

719 Lintner, B. R. and J. D. Neelin, 2009: Soil Moisture Impacts on Convective Margins. *J. Hydrometeor.*,
720 **10**, 1026–1039.

721

722 Lintner, B. R., P. Gentine, K. L. Findell, F. D’Andrea, A. H. Sobel, and G. D. Salvucci, 2013: An
723 Idealized Prototype for Large-Scale Land–Atmosphere Coupling. *J. Climate*, **26**, 2379–2389.

724

725 Lo, M.-H., and J. S. Famiglietti, 2010: Effect of water table dynamics on land surface hydrologic
726 memory, *J. Geophys. Res.*, **115**, D22118, doi:[10.1029/2010JD014191](https://doi.org/10.1029/2010JD014191).

727

728 Meng X.H., Evans J.P., McCabe M.F. (2014a) The Impact of Observed Vegetation Changes on Land–

729 Atmosphere Feedbacks During Drought. *J Hydrometeor.*, **15**, 759–776. doi: 10.1175/JHM-D-13-
730 0130.1
731
732 Meng, X. H., Evans, J. P. and McCabe, M. F. (2014b): The influence of inter-annually varying albedo
733 on regional climate and drought, *Clim Dyn*, 42(3-4), 787–803, doi:10.1007/s00382-013-1790-0.
734
735 Miralles, D. G., R. A. M. De Jeu, J. H. Gash, T. R. H. Holmes, and A. J. Dolman. 2011: Magnitude and
736 Variability of Land Evaporation and Its Components at the Global Scale, *Hydrol. Earth Syst. Sc.*, **15**
737 **(3)**, 967–981, doi:10.5194/hess-15-967-2011.
738
739 Miralles, D. G., T. R. H. Holmes, R. A. M. De Jeu, J. H. Gash, A. G. C. A. Meesters, and A. J.
740 Dolman, 2010: Global Land-surface Evaporation Estimated from Satellite-based Observations,
741 *Hydrol. Earth Syst. Sc.*, **7 (5)**, 8479–8519. doi:10.5194/hessd-7-8479-2010.
742
743 Mu, Q., F. A. Heinsch, M. Zhao, and S. W. Running, 2007: Development of a global
744 evapotranspiration algorithm based on MODIS and global meteorology data, *Remote Sens. Environ.*,
745 **111**, 519-536, doi:10.1016/j.rse.2007.04.015.
746
747 Mu, Q., L. A. Jones, J. S. Kimball, K. C. McDonald and S. W. Running, 2009: Satellite assesment of
748 land surface evapotranspiration for the pan-Arctic domain, *Water Resour. Res.*, **45**,
749 doi:10.1029/2008WR007189.
750
751 Niu, G.-Y., Z.-L. Yang, R. E. Dickinson, L. E. Gulden, and H. Su, 2007: Development of a Simple
752 Groundwater Model for Use in Climate Models and Evaluation with Gravity Recovery and Climate
753 Experiment Data, *J. Geophys. Res.-Atmos.*, **112 (D7)**, doi:10.1029/2006JD007522.
754
755 Njoku, E.G., T.J. Jackson, V. Lakshmi, T.K. Chan, and S.V. Nghiem, 2003: Soil Moisture Retrieval
756 from AMSR-E, *Geoscience and Remote Sensing, IEEE Transactions*, **41 (2)**, 215–229,
757 doi:10.1109/TGRS.2002.808243.
758

759 Oleson, K., and Coauthors, 2010: Technical Description of Version 4.0 of the Community Land Model
760 (CLM), [Http://dx.doi.org/10.5065/D6FB50WZ](http://dx.doi.org/10.5065/D6FB50WZ),
761 <http://nldr.library.ucar.edu/repository/collections/TECH-NOTE-000-000-000-848>.
762

763 Pielke, R. A., and Coauthors, 2011: Land Use/land Cover Changes and Climate: Modeling Analysis
764 and Observational Evidence, *WIREs. Clim.*, **2 (6)**, 828–850. doi:10.1002/wcc.144.
765

766 Pitman, A. J. 2003: The Evolution of, and Revolution in, Land Surface Schemes Designed for Climate
767 Models, *Int. J. Climatol.*, **23 (5)**, 479–510, doi:10.1002/joc.893.
768

769 Press, W. H., B. P. Flannery, S. A. Teukolsky, and W. T. Vetterling, 1992: Nonparametric or rank
770 correlation, Numerical Recipes in C: The Art of Scientific Computing, L. Cowles and A. Harvey, Eds.,
771 2nd ed. Cambridge University Press, 639–644.
772

773 Qian, T., A. Dai, K. E. Trenberth, and K. W. Oleson. 2006: Simulation of Global Land Surface
774 Conditions from 1948 to 2004. Part I: Forcing Data and Evaluations, *J. Hydrometeor.*, **7 (5)**, 953–975,
775 doi:10.1175/JHM540.1.
776

777 Rodell, M., and Coauthors, 2004: The Global Land Data Assimilation System, *Bull. Amer. Meteor.*
778 *Soc.*, **85 (3)**, 381–394, doi:10.1175/BAMS-85-3-381.
779

780 Santanello, J. A., C. D. Peters-Lidard, and S. V. Kumar, 2011: Diagnosing the Sensitivity of Local
781 Land-Atmosphere Coupling via the Soil Moisture-Boundary Layer Interaction, *J. Hydrometeor.*, **12**
782 **(5)**, 766–786, doi:10.1175/JHM-D-10-05014.1.
783

784 Schenk, H. J., and R. B. Jackson, 2002: The Global Biogeography of Roots, *Ecol. Monographs*, **73**,
785 311-328.
786

787 Schlesinger, W. H. and Jasechko, S., 2014: Transpiration in the global water cycle, *Agr. Forest*
788 *Meteorol.*, **189–190**, 115–117, doi:10.1016/j.agrformet.2014.01.011.

789
790 Seneviratne, S. I., T. Corti, E. L. Davin, M. Hirschi, E. B. Jaeger, I. Lehner, B. Orlowsky, and A. J.
791 Teuling, 2010: Investigating Soil Moisture-climate Interactions in a Changing Climate: A Review,
792 *Earth-Science Reviews*, **99**, 125–161, doi:10.1016/j.earscirev.2010.02.004.
793
794 Small, E. E., and S. A. Kurc, 2003: Tight Coupling Between Soil Moisture and the Surface Radiation
795 Budget in Semiarid Environments: Implications for Land-atmosphere Interactions, *Water Resour. Res.*,
796 **39** (10), doi:10.1029/2002WR001297.
797
798 Taylor, C.r M., and R. J. Ellis, 2006: Satellite Detection of Soil Moisture Impacts on Convection at the
799 Mesoscale, *Geophys. Res. Lett.*, **33** (3), doi:10.1029/2005GL025252.
800
801 Vergnes, J.-P., B. Decharme, R. Alkama, E. Martin, F. Habets, and H. Douville, 2012: A Simple
802 Groundwater Scheme for Hydrological and Climate Applications: Description and Offline Evaluation
803 over France, *J. Hydrometeor.*, **13** (4), 1149–1171, doi:10.1175/JHM-D-11-0149.1.
804
805 Westra, D., G. J. Steeneveld, and A. A. M. Holstag, 2012: Some Observational Evidence for Dry Soils
806 Supporting Enhanced Relative Humidity at the Convective Boundary Layer Top, *J. Hydrometeor.*, **13**,
807 1347-1358.
808
809 Williams, J. L., and R. M. Maxwell, 2011: Propagating Subsurface Uncertainty to the Atmosphere
810 Using Fully Coupled Stochastic Simulations, *J. Hydrometeor.*, **12** (4), 690–701,
811 doi:10.1175/2011JHM1363.1.
812
813 Yu, Z., Da. Pollard, and L. Cheng, 2006: On Continental-scale Hydrologic Simulations with a Coupled
814 Hydrologic Model, *J. Hydrology*, **331**, 110–124, doi:10.1016/j.jhydrol.2006.05.021.
815
816 Zeng, X., and M. Decker, 2009: Improving the Numerical Solution of Soil Moisture–Based Richards
817 Equation for Land Models with a Deep or Shallow Water Table. *J. Hydrometeor.*, **10**, 308-319.
818

819 Zhang, J., W.-C. Wang, and J. Wei, 2008: Assessing Land-atmosphere Coupling Using Soil Moisture
820 from the Global Land Data Assimilation System and Observational Precipitation, *J. Geophys. Res.-*
821 *Atmos.*, **113**, doi:10.1029/2008JD009807.

822

823 **List of Figures**

824 **Figure 1.** Observations of the domain (18°S-11°S and 120°E-150°E) averaged mean annual cycle of
825 precipitation (P in mm day⁻¹).

826

827 **Figure 2.** (a) The mean normalized (using the first two moments) first layer soil moisture (SM₁) from
828 the CTRL and DRY simulations and the AMSR-E observations. (b) The difference between the mean
829 SM₁ (from all simulations over all months from 2004-2009) and the AMSR-E observations.

830

831 **Figure 3.** The mean ET (Wm⁻²) from the wet season (DJF shown in the left hand column) and the
832 transition between the dry and wet seasons (SON shown in the right hand column). The ensemble
833 mean ET from (a) CTRL over DJF, (b) CTRL over SON, (c) DRY for DJF, (d) DRY from SON, (e)
834 OBS (the mean of three gridded ET products) over DJF, and (f) OBS for SON.

835

836 **Figure 4.** The monthly soil moisture (SM in mm³mm⁻³) from the ensemble mean from CTRL and
837 DRY, AMSR-E, and flux tower measurements (OBS_{tower}) from flux tower sites at (a) Howard Springs
838 at 10cm depth and (b) Adelaide River at 5cm depth. The monthly evapotranspiration (ET Wm⁻²) from
839 CTRL, DRY, the mean of three ET products (OBS_{gridded}) and the measurements at the (c) Howard
840 Springs and (d) Adelaide River flux tower sites.

841

842 **Figure 5.** Spatiotemporal mean soil moisture (mm³mm⁻³) SM as a function of depth (m) for (a) DJF
843 and (b) SON.

844

845 **Figure 6.** The ensemble median Kendall-tau correlation metric (Kτ) between the afternoon time
846 (local) evaporative fraction (EF) and the afternoon computed lifting condensation level (LCL) at from
847 (a) CTRL over DJF, (b) CTRL from SON, (c) DRY over DJF, and (d) DRY from SON. The black-
848 outlined squares in (a)-(d) denote the values from the flux tower sites. Only statistically significant
849 (95% confidence level) results are shown in (a)-(d).

850

851 **Figure 7.** The ensemble median Kendall-tau correlation metric (Kτ) between the morning first layer
852 soil moisture (SM₁) and the afternoon computed lifting condensation level (LCL) from (a) CTRL over

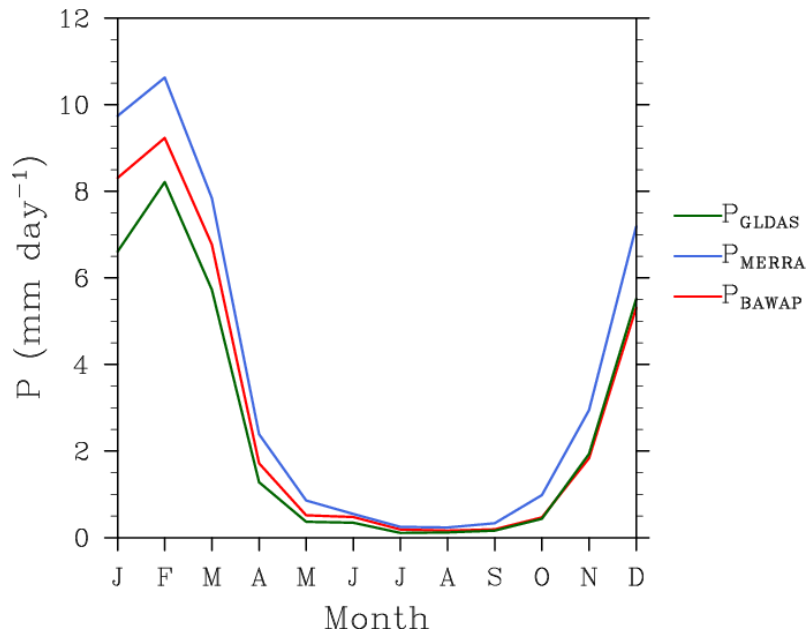
853 DJF, (b) CTRL from SON, (c) DRY over DJF, and (d) DRY from SON. The black-outlined squares in
854 (a)-(d) denote the values from the flux tower sites. Only statistically significant (95% confidence
855 level) results are shown in (a)-(d).

856

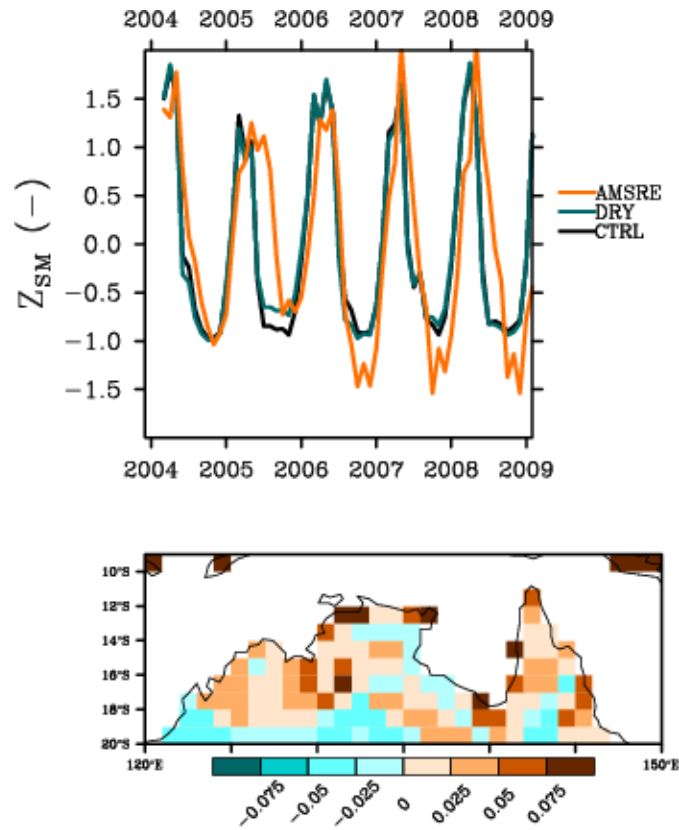
857 **Figure 8.** The ensemble median Kendall-tau correlation metric ($K\tau$) between the morning root zone
858 soil moisture (SM_{rz}) and the afternoon computed lifting condensation level (LCL) from (a) CTRL over
859 DJF, (b) CTRL from SON, (c) DRY over DJF, and (d) DRY from SON. The black-outlined squares in
860 (a)-(d) denote the values from the Howard Springs flux tower site. Only statistically significant (95%
861 confidence level) results are shown in (a)-(d).

862

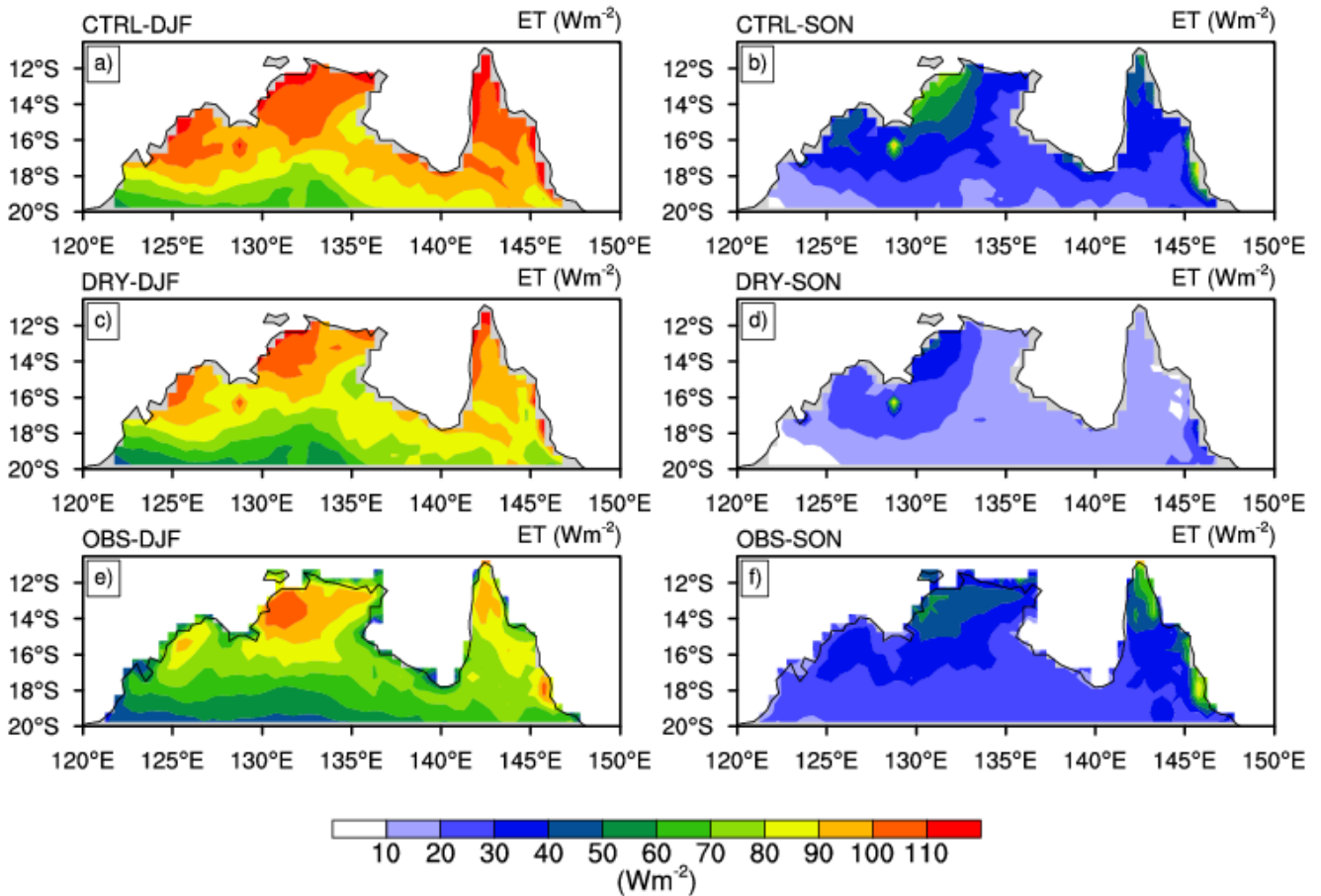
863 **Figure 9.** The standard deviation of the Kendall-tau correlation metric ($K\tau$) among the ensemble
864 members between the afternoon computed lifting condensation level (LCL) and either the morning
865 root zone soil moisture (SM_{rz}) over (a) DJF, (b) SON, or the morning first layer soil moisture (SM_1)
866 over (c) DJF, and (d) SON.



868 **Figure 1.** Observations of the domain (18°S-11°S and 120°E-150°E) averaged mean annual cycle of (a)
 869 precipitation (P in mm day⁻¹).
 870



872 **Figure 2.** (a) The mean normalized (using the first two moments) first layer soil moisture (SM_1) from
 873 the CTRL and DRY simulations and the AMSR-E observations. (b) The difference between the mean
 874 SM_1 (from all simulations over all months from 2004-2009) and the AMSR-E observations.
 875

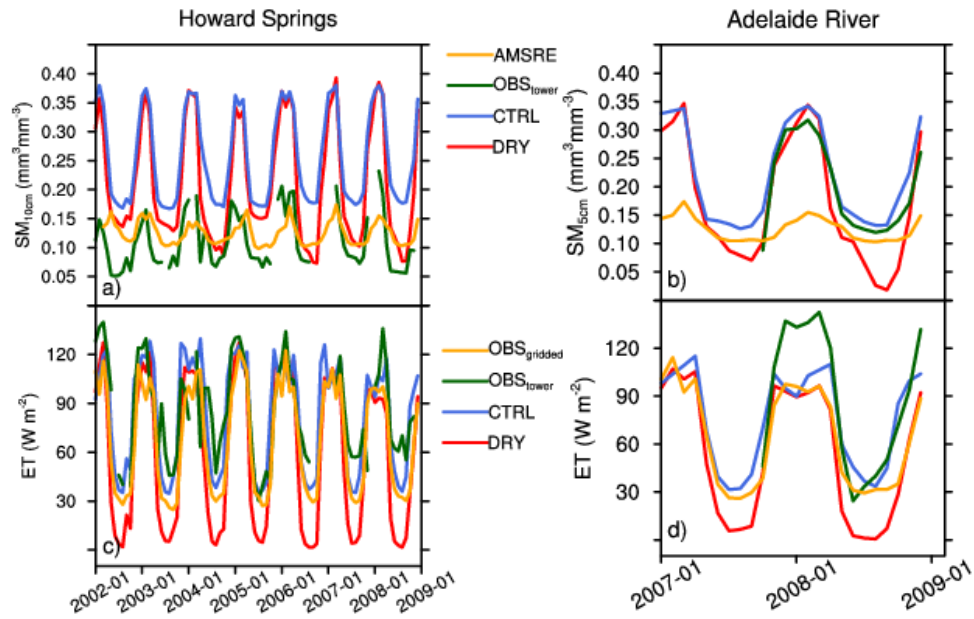


878

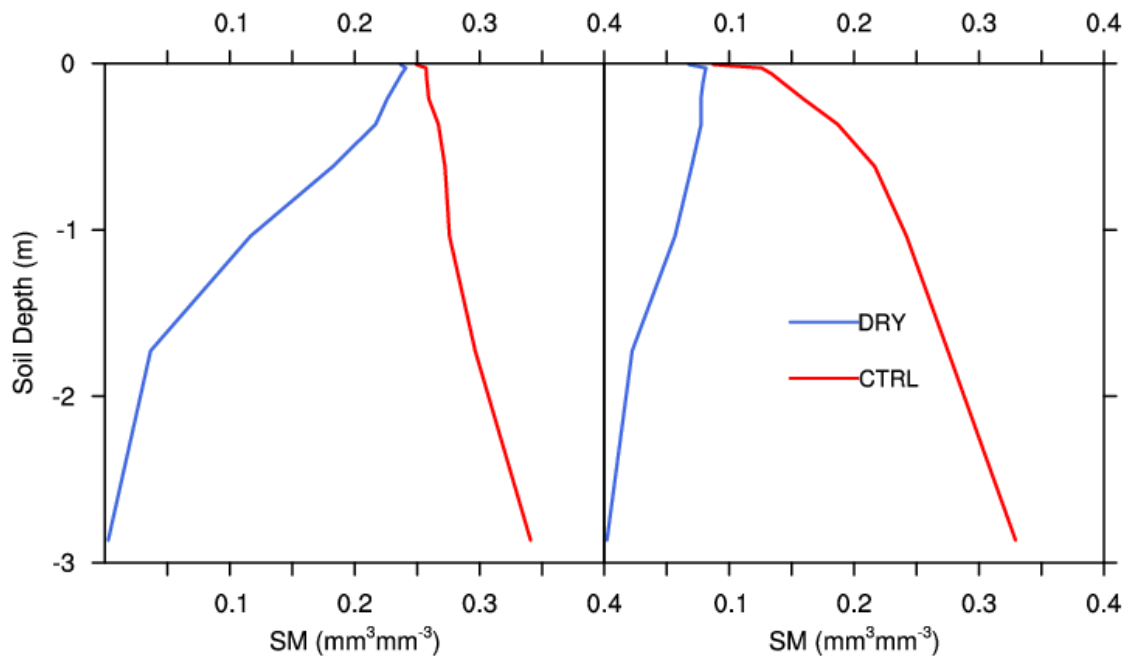
879

880 **Figure 3.** The mean ET (Wm⁻²) from the wet season (DJF shown in the left hand column) and the
 881 transition between the dry and wet seasons (SON shown in the right hand column). The ensemble
 882 mean ET from (a) CTRL over DJF, (b) CTRL over SON, (c) DRY for DJF, (d) DRY from SON, (e)
 883 OBS (the mean of three gridded ET products) over DJF, and (f) OBS for SON.

884



885 **Figure 4.** The monthly soil moisture (SM in $\text{mm}^3\text{mm}^{-3}$) from the ensemble mean from CTRL and
 886 DRY, AMSR-E, and flux tower measurements ($\text{OBS}_{\text{tower}}$) from flux tower sites at (a) Howard Springs
 887 at 10cm depth and (b) Adelaide River at 5cm depth. The monthly evapotranspiration (ET W m^{-2}) from
 888 CTRL, DRY, the mean of three ET products ($\text{OBS}_{\text{gridded}}$) and the measurements at the (c) Howard
 889 Springs and (d) Adelaide River flux tower sites.



891

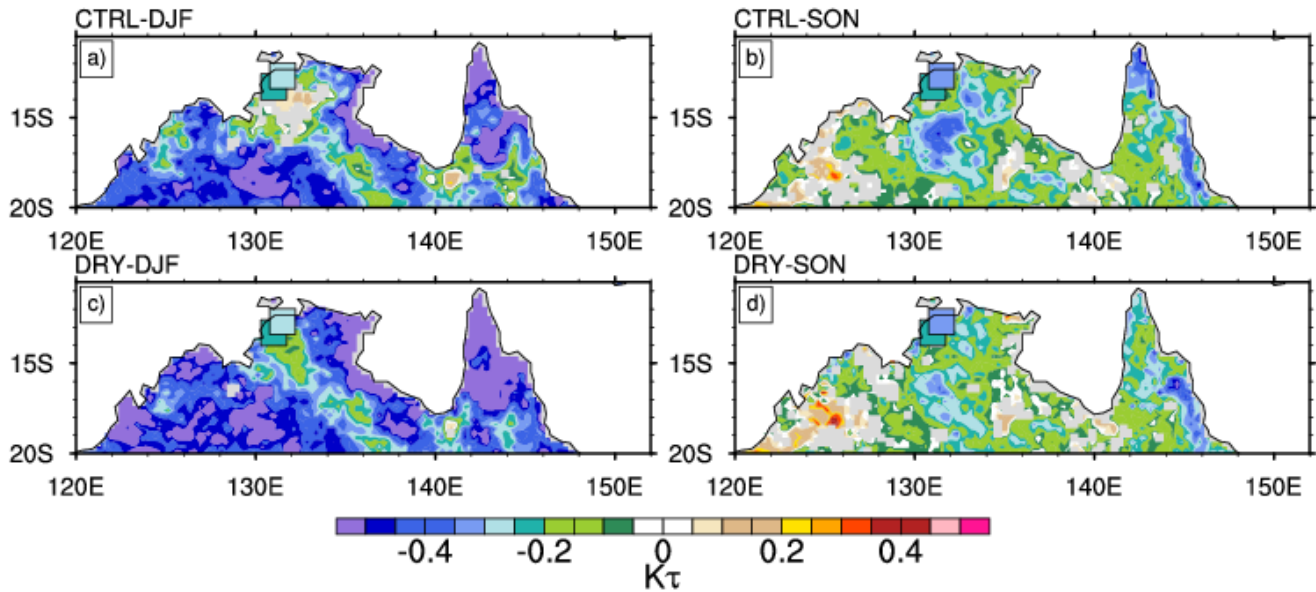
892

893 **Figure 5.** Spatiotemporal mean soil moisture ($\text{mm}^3\text{mm}^{-3}$) SM as a function of depth (m) for (a) DJF

894 and (b) SON.

895

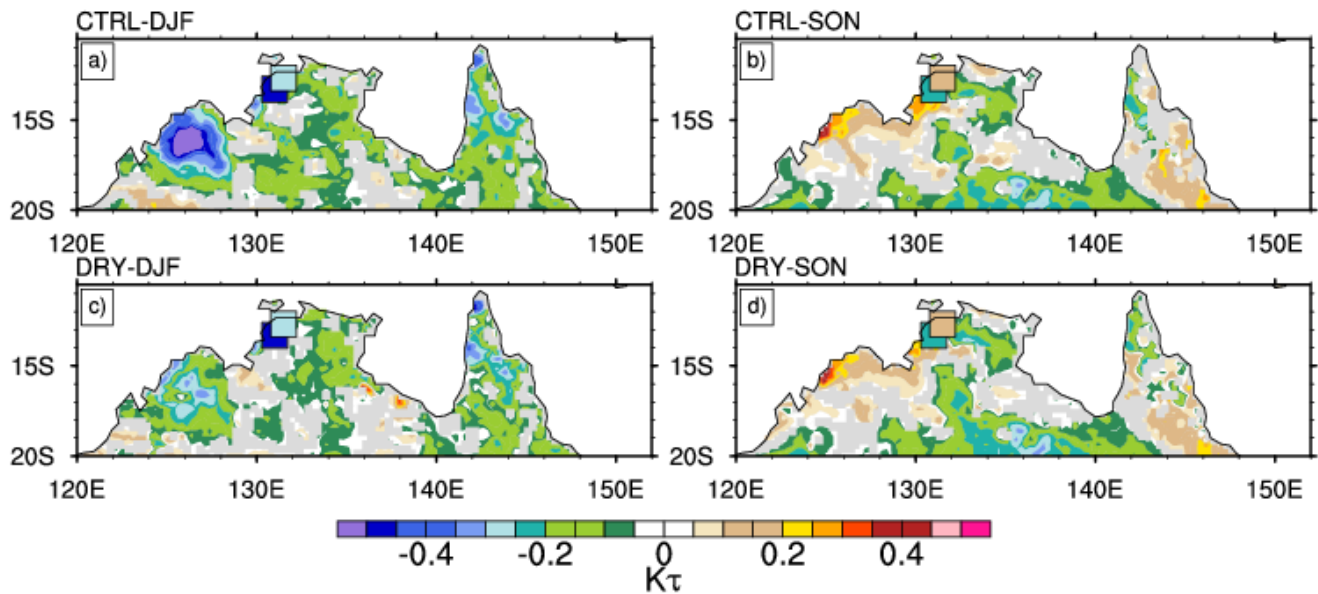
896



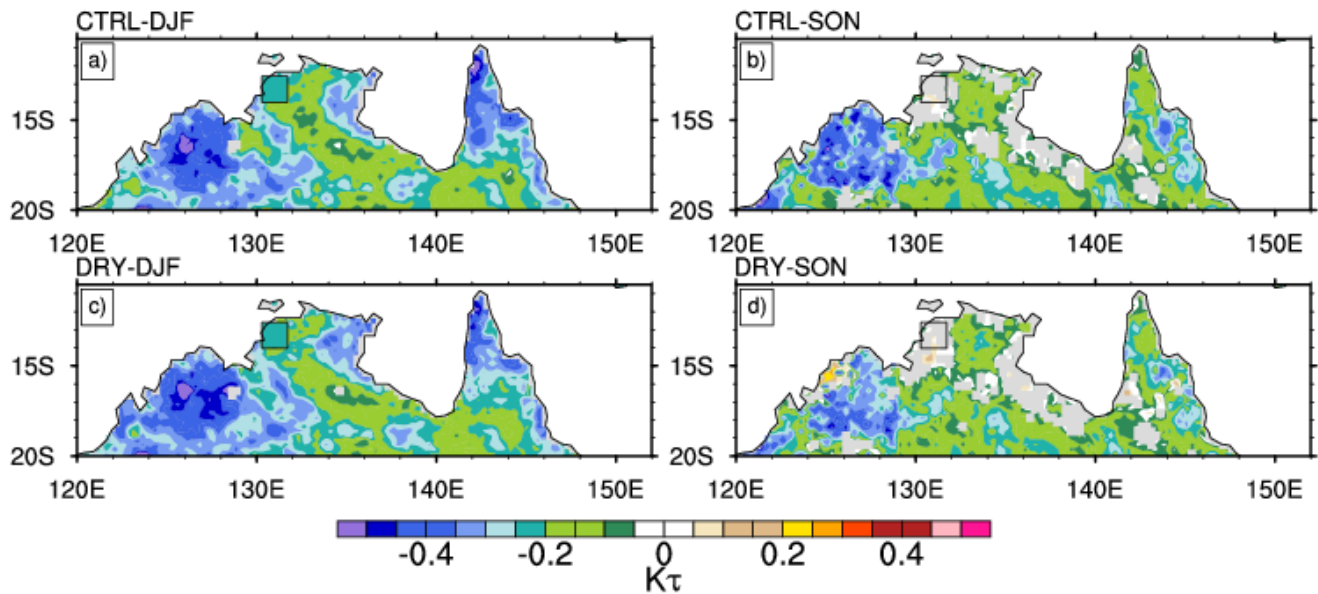
899 **Figure 6.** The ensemble median Kendall-tau correlation metric ($K\tau$) between the afternoon time
 900 (local) evaporative fraction (EF) and the afternoon computed lifting condensation level (LCL) at from
 901 (a) CTRL over DJF, (b) CTRL from SON, (c) DRY over DJF, and (d) DRY from SON. The black-
 902 outlined squares in (a)-(d) denote the values from the flux tower sites. Only statistically significant
 903 (95% confidence level) results are shown in (a)-(d).

904

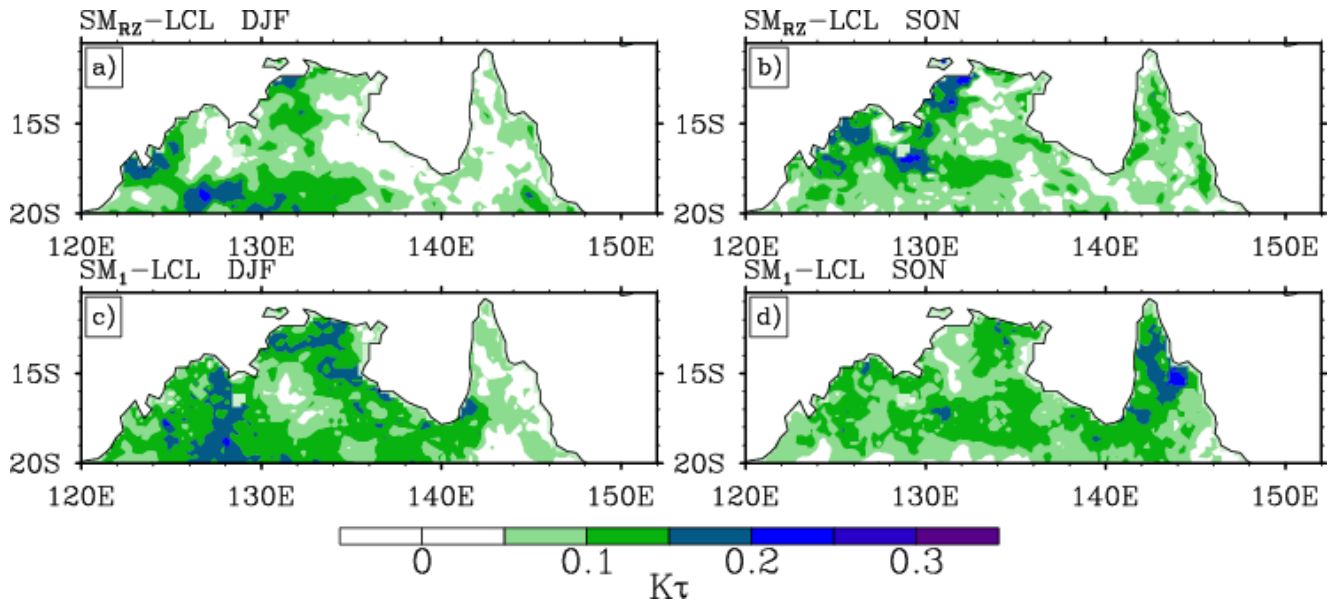
905



907 **Figure 7.** The ensemble median Kendall-tau correlation metric ($K\tau$) between the morning first layer
 908 soil moisture (SM_1) and the afternoon computed lifting condensation level (LCL) from (a) CTRL over
 909 DJF, (b) CTRL from SON, (c) DRY over DJF, and (d) DRY from SON. The black-outlined squares in
 910 (a)-(d) denote the values from the flux tower sites. Only statistically significant (95% confidence
 911 level) results are shown in (a)-(d).
 912



914 **Figure 8.** The ensemble median Kendall-tau correlation metric ($K\tau$) between the morning root zone
 915 soil moisture (SM_{rz}) and the afternoon computed lifting condensation level (LCL) from (a) CTRL over
 916 DJF, (b) CTRL from SON, (c) DRY over DJF, and (d) DRY from SON. The black-outlined squares in
 917 (a)-(d) denote the values from the Howard Springs flux tower site. Only statistically significant (95%
 918 confidence level) results are shown in (a)-(d).
 919



921 **Figure 9.** The standard deviation of the Kendall-tau correlation metric ($K\tau$) among the ensemble
 922 members between the afternoon computed lifting condensation level (LCL) and either the morning
 923 root zone soil moisture (SM_{RZ}) over (a) DJF, (b) SON, or the morning first layer soil moisture (SM_1)
 924 over (c) DJF, and (d) SON.