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Did aboriginal vegetation burning impact on the Australian summer monsoon?

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[1] Aboriginal vegetation burning practices and their role in the Australian environment remains a central theme of Australian environmental history. Previous studies have identified a decline in the Australian summer monsoon during the late Quaternary and attributed it to land surface-atmosphere feedbacks, related to Aboriginal burning practices. Here we undertake a comprehensive, ensemble model evaluation of the effects of a decrease in vegetation cover over the summer monsoon region of northern Australia. Our results show that the climate response, while relatively muted during the full monsoon, was significant for the pre-monsoon season (austral spring), with decreases in precipitation, higher surface and ground temperatures, and enhanced atmospheric stability. These early monsoon season changes can invoke far-reaching ecological impacts and set-up land surface-atmosphere feedbacks that further accentuate atmospheric stability. **Citation:** Notaro, M., K.-H. Wyrwoll, and G. Chen (2011), Did aboriginal vegetation burning impact on the Australian summer monsoon?, *Geophys. Res. Lett.*, *38*, L11704, doi:10.1029/2011GL047774.

1. Introduction

[2] Vegetation plays a fundamental role in the climate system, and its importance is emphasized by the Quaternary climate literature [*Wyputta and McAvaney*, 2001; *Crucifix and Hewitt*, 2005; *Claussen*, 2009], in which the role of vegetation feedbacks has been shown to be pivotal to the establishment and maintenance of regional climate states [*Gallimore et al.*, 2005; *Hales et al.*, 2006; *Liu et al.*, 2006a, 2006b; *Notaro et al.*, 2008; *Cheddadi and Bar-Hen*, 2009; *Dallmeyer et al.*, 2010; *Notaro and Gutzler*, 2011]. The impact of Aboriginal land-use practices has been a pervasive theme in Australian paleoecology [*Hallam*, 1975; *Bowman*, 1998; *Kershaw et al.*, 2002; *Lynch et al.*, 2007; *Mooney et al.*, 2011]. These practices have been linked to issues of megafauna extinction [*Johnson et al.*, 1999] and claims, based on an atmosphere general circulation model experiment [*Miller et al.*, 2005], that Aboriginal vegetation burning led to surface modifications on a scale sufficient to weaken the late Quaternary Australian summer monsoon. However, in building on this claim, other model results indicate a significantly more muted effect [*Pitman and Hesse*, 2007; *Marshall and Lynch*, 2008], but with their limited scope inviting a fuller evaluation. In particular, *Marshall and Lynch* [2008] applied a coarse global climate model coupled to a crude land surface model, while *Pitman and Hesse* [2007] only ran mid-monsoon simulations for January and did not consider land-atmosphere interactions during the flanks of the monsoon season. Here we undertake an ensemble-based examination of the influence of vegetation change on the northern Australian summer monsoon regime for the pre-, mid-, and post-monsoon period of November-March.

2. Model and Methods

[3] We use the National Center for Atmospheric Research Community Climate System Model Version 3.5 (NCAR CCSM3.5) [*Gent et al.*, 2010] with a finite volume dynamical core for the atmospheric component and 1.9° × 2.5° horizontal resolution. This fully coupled global atmosphere-ocean-land-ice model includes vegetation dynamics from the Community Land Model Version 3.5-Dynamic Global Vegetation Model, with annual vegetation processes based on the Lund-Potsdam-Jena model [*Levis et al.*, 2004].

[4] The major observed vegetation associations over the monsoon region of northern Australia comprise a dominance of eucalypt woodland/grassland and tussock and hummock grasslands; some acacia shrublands and acacia forest and woodland; and eucalypt open forest in higher rainfall regions. The model produces an idealized potential vegetation assemblage of grasses and tropical deciduous trees, similar to the observed, savanna-shrubland-grassland landscape.

[5] The Australian monsoon is characterized by a low-level monsoon trough over northern Australia, with cyclonic vorticity and equatorial westerlies that advect moisture into the region, and an upper-level anticyclone. Across much of the northern Australian prairie, roughly two-thirds of annual precipitation falls during the core monsoon season of January-March. The model simulates a reasonable representation of the spatial pattern of monsoon precipitation, but with precipitation amounts too high, and lacking the steep southward precipitation gradients. The simulated timing of monsoon precipitation is reasonable, but with an extended monsoon season into April. The model produces a good representation of lower tropospheric winds, which control monsoon 'inflow' into northwestern Australia, and a similar circulation pattern to the Fast Ocean Atmosphere Model, which has been demonstrated to give an acceptable representation of the northern Australian monsoon regime [*Marshall and Lynch*, 2006; *Wyrwoll et al.*, 2007].

[6] Unlike previous modeling studies, which primarily rely in single experiments, we investigate vegetation feedbacks across northern Australia using a comprehensive initial value ensemble approach. A multi-century, modern-day

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Table 1. Annual and seasonal Differences (Ensemble-Control) for 36 Variables, Averaged Across Northern Australia[a]

	ANN	JFM	AMJ	JAS	OND
Leaf area index m^2/m^2	**−0.56**	**−0.78**	**−0.46**	**−0.35**	**−0.66**
Stem area index m^2/m^2	**−0.21**	**−0.21**	**−0.21**	**−0.21**	**−0.21**
Net CO_2 flux umol/m^{2}s	**+0.35**	**+0.64**	**+0.26**	**+0.08**	**+0.44**
Surface stress kg/m/s^2	**−0.011**	**−0.009**	**−0.016**	**−0.015**	−0.003
Surface albedo	+0.0001	**−0.0054**	−0.0016	**+0.0046**	+0.0014
Net surface shortwave radiation flux W/m^2	**+1.01**	**+3.31**	−1.03	−1.00	**+2.78**
Net surface longwave radiation flux W/m^2	**+2.30**	**+1.98**	+0.52	**+2.38**	**+4.32**
Heat flux into soil layers W/m^2	**+0.49**	**+1.16**	−0.02	**+0.28**	**+0.54**
Total sensible heat flux W/m^2	−0.59	+0.51	−1.32	**−1.61**	+0.07
Ground sensible heat flux W/m^2	**+7.36**	**+4.35**	**+5.48**	**+9.75**	**+9.85**
Vegetation sensible heat flux W/m^2	**−7.94**	**−3.84**	**−6.79**	**−11.36**	**−9.78**
Total latent heat flux W/m^2	**−1.19**	−0.34	−0.26	**−2.05**	**−2.15**
Transpiration W/m^2	**−5.84**	**−7.34**	**−5.77**	**−3.97**	**−6.29**
Canopy evaporation W/m^2	**−3.98**	**−8.15**	**−1.74**	**−0.43**	**−5.60**
Ground evaporation W/m^2	**+8.63**	**+15.14**	**+7.30**	**+2.35**	**+9.75**
Bowen ratio	+0.024	+0.004	+0.003	+0.058	+0.031
Evapotranspiration cm/period	**−1.48**	−0.11	−0.06	**−0.64**	**−0.67**
Precipitation cm/period	**−4.78**	−1.92	+1.21	−0.03	**−4.04**
Precipitation minus evaporation cm/period	**−3.30**	−1.81	+1.28	**+0.61**	**−3.37**
Large-scale precipitation cm/period	**−1.61**	**−1.02**	+0.31	+0.04	**−0.93**
Convective precipitation cm/period	**−3.17**	−0.90	+0.91	−0.07	**−3.11**
Ground temperature °C	**+1.34**	**+1.36**	**+0.98**	**+1.09**	**+1.93**
Surface air temperature °C	**+0.42**	**+0.40**	**+0.35**	**+0.35**	**+0.57**
Vegetation temperature °C	**+0.26**	**+0.27**	**+0.25**	**+0.18**	**+0.35**
Maximum 2-m air temperature °C	**−0.09**	+0.01	**−0.20**	**−0.23**	+0.05
Minimum 2-m air temperature °C	**+0.60**	**+0.51**	**+0.60**	**+0.58**	**+0.71**
Diurnal temperature range °C	**−0.69**	**−0.50**	**−0.79**	**−0.81**	**−0.66**
Planetary boundary layer depth m	**+14.12**	**+8.66**	+6.24	**+13.63**	**+27.95**
850-hPa divergence s^{-1}	+0.31	**+0.47**	−0.40	−0.11	**+1.27**
700-hPa vertical velocity Pa/s	**+0.0013**	**+0.014**	−0.0010	+0.0008	**+0.0038**
Precipitable water kg/m^2	−0.26	−0.17	+0.31	−0.40	**−0.78**
Total cloud cover fraction	−0.004	−0.004	+0.004	−0.000	**−0.015**
Low cloud cover fraction	**−0.005**	**−0.011**	+0.002	+0.001	**−0.011**
Medium cloud cover fraction	−0.003	−0.007	+0.005	−0.000	**−0.011**
High cloud cover fraction	−0.004	−0.003	+0.004	−0.001	**−0.014**
Convection top pressure hPa	−2.2	**−8.1**	−6.2	+0.4	+5.2

[a] Bolded values indicate statistically significant differences. ANN, annual; JFM, January-March; AMJ, April-June; JAS, July-September; OND, October-December.

simulation was run, and the last 80 years were considered as the control simulation for this study. The dynamic vegetation code in CCSM3.5 allows the model to generate inter-annually-varying distributions of plant functional types (PFTs) in the control simulation. To develop an 80-member ensemble set, for each year in the control simulation, the total vegetation cover fraction was reduced by 0.2 across northern Australia and a one-year ensemble simulation produced. The total vegetation cover fraction is reduced while maintaining the original proportions of PFTs. Since CCSM3.5 only updates its distribution of PFTs once per year, the vegetation cover fraction remains constant all year; in the ensemble experiments, the reduction in vegetation cover fraction is imposed at the beginning of the year. By comparing the climate of the ensemble and control, the impact of reduced vegetation cover on regional climate is determined.

[7] Traditional burning practices across northern Australia were undertaken in a systematic and purposeful manner, in part, to promote grassland regrowth. Burning occurred throughout the dry season, but generally initiated in the early-mid dry season (May-August) under cooler, less fire-prone weather conditions [*Russell-Smith*, 2002]. To mimic these changes, the total model vegetation cover fraction was reduced across the monsoon region. These experiments only roughly emulate burn practices across northern Australia, given that the quantitative extent of prior vegetation burn and its impact on individual PFTs are not known.

[8] The discussion focuses on October-March, with results summarized in Table 1 and Figure 1. October-December represents the pre-to-early monsoon season. December is included, recognizing that the onset date for the northern Australian summer monsoon is middle to late December [*Kajikawa et al.*, 2010]. All differences that are described in the text are statistically significant at the $p<0.1$ level, based on student t-tests, unless identified otherwise.

3. Results

[9] In response to reduced vegetation cover, atmospheric subsidence increases, with changes in 700-hPa vertical velocity of +0.0038 Pa/s in OND and +0.0014 Pa/s in JFM (Figure 2 and Table 1). This corresponds with an increase in 850 hPa divergence of $+1.27*10^{-7}$ s^{-1} for OND and $+0.47*10^{-7}$ s^{-1} for JFM (Table 1). The changes in vertical velocity and low-level divergence are most pronounced during November and December (Figure 1), indicating a more stable and drier atmosphere. This is reflected in a reduction in precipitable water by −0.78 kg/m^2 in OND (Table 1), again with the greatest decrease in Nov-Dec (Figure 1). Total cloud cover fraction decreases by −0.015 in OND (Table 1). These moisture and cloud cover changes, attributed to diminished evapotranspiration and moisture

convergence, carry clear implications for the radiation/energy balance.

[10] With the imposed reduction in vegetation cover, the area-averaged leaf area index decreases by −0.66 m^2/m^2 (−0.79 m^2/m^2) in OND (JFM) (Table 1). The response in surface albedo is quite small, as the background soil is not highly reflective and the change in upper soil water fraction is not dramatic [*Notaro et al.*, 2008; *Notaro and Gutzler*, 2011]. Incoming solar radiation has increased due to diminished cloud cover and atmospheric moisture, accounting for a net shortwave radiation increase of +2.78 W/m^2 for OND and +3.31 W/m^2 for JFM (Table 1). The increased net (downward) shortwave flux leads to an increase in soil heat flux of +0.54 W/m^2 for OND and +1.16 W/m^2 for JFM and higher net upward longwave radiation flux of +4.32 W/m^2 for OND and +1.98 W/m^2 for JFM – related to increases in surface temperatures, with the lower downward longwave radiation flux due to a decrease in cloud cover and atmospheric moisture (Table 1). Surface sensible heat flux increases in OND by +0.07 W/m^2 and in JFM by +0.51 W/m^2, although not statistically significant (Table 1). Latent heat flux is reduced in OND by −2.15 W/m^2 and in JFM by −0.34 W/m^2, although only the former is significant (Table 1).

[11] In response to drier conditions and weakened turbulent fluxes, ground and 2-m surface air temperatures increase in OND by +1.93°C and +0.57°C, respectively, and in JFM by +1.36°C and +0.40°C, respectively (Table 1), with the largest warming in November (Figure 1). Vegetation temperature also increases by +0.35°C in OND and +0.27°C in JFM (Table 1). There is a notable decrease in the diurnal temperature range of −0.66°C in OND and −0.50 °C in JFM (Table 1). With the reduction in LAI, heat flux into the soil increases, elevating ground and surface air temperatures at night and decreasing the diurnal temperature range [*Zhou et al.*, 2007]. Precipitation declines by −4.04 cm in OND and −1.92 cm in JFM (latter not significant) (Table 1). The decreases in precipitable water and precipitation are greatest in November and December (Figure 1). The reduction in vegetation cover causes decreases in net evapotranspiration of −0.67 cm for OND and −0.11 cm for JFM (latter not significant) (Table 1), related to diminished transpiration and canopy evaporation. The net precipitation minus evaporation declines by −3.37 cm in OND and −1.81 cm in JFM (latter not significant) (Table 1). The overall conclusion that emerges is of a drier, warmer pre-monsoon season, but with relatively minor changes once the active monsoon has set-in.

4. Discussion

[12] Our model results show that the impacts of vegetation changes on the peak monsoon season in northern Australia are relatively minor. The precipitation reduction during the mid-monsoon months of JFM in the experiments amounts to

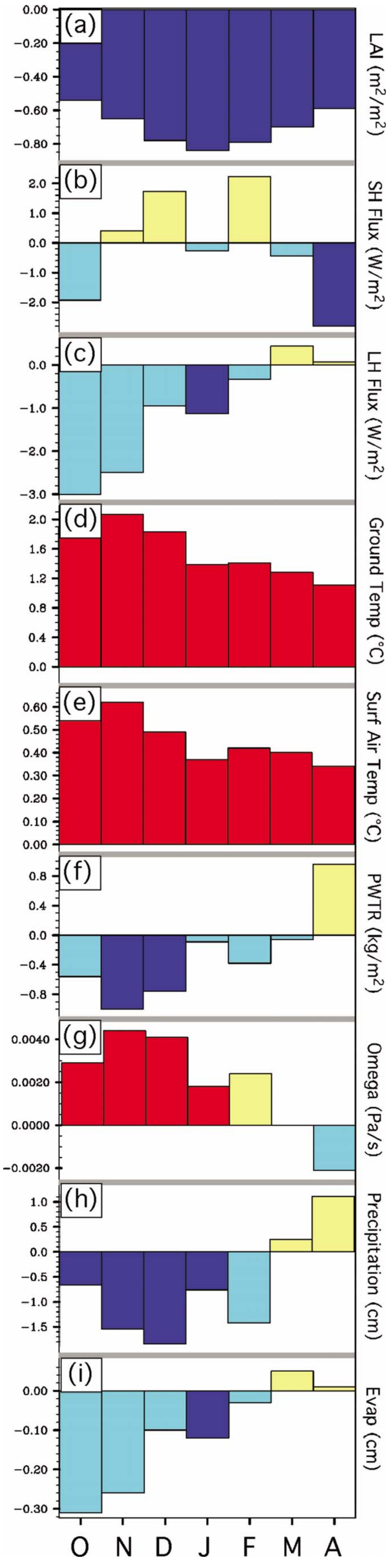

Figure 1. Differences (ensemble-control) in monthly (October-April) (a) leaf area index (m^2/m^2), (b) sensible heat flux (W/m^2), (c) latent heat flux (W/m^2), (d) ground temperature (°C), (e) surface air temperature (°C), (f) precipitable water (kg/m^2), (g) vertical velocity (Pa/s), (h) precipitation (cm), and (i) evapotranspiration (cm), averaged across northern Australia. Red and dark blue bars indicate statistically significant increases and decreases, respectively.

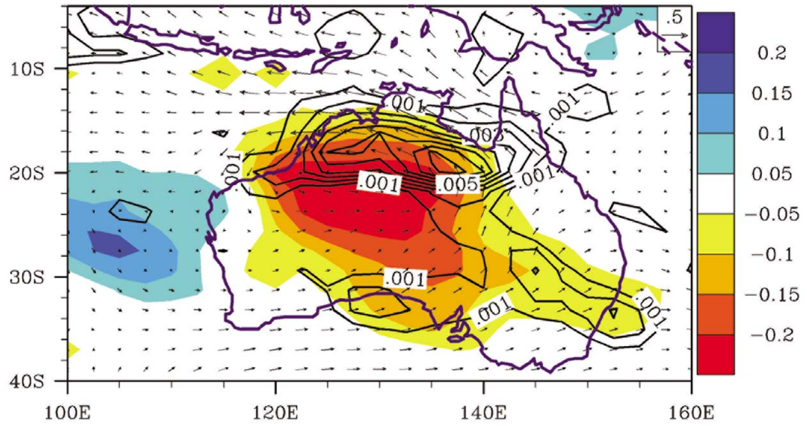

Figure 2. Differences (ensemble-control) during November-February in 850 hPa specific humidity (shading: g/kg), 850 hPa wind (vectors: m/s), and 700 hPa vertical velocity (contour: Pa/s).

about −0.2 mm/day, which is insignificant given the magnitude of the precipitation received during the peak of the monsoon. In response to reduced evapotranspiration and cloud cover, the ground and surface air temperatures increase. While the changes in precipitable water, cloud cover, vertical velocity, and 850 hPa divergence are modest, they are indicators of a statistically significant drier, more stable atmosphere.

[13] The early/pre-monsoon months show a significant climate response to vegetation changes, highlighted by the precipitation changes. For the early monsoon season, precipitation decreases by −40 mm in response to diminished evapotranspiration and a triggered reduction in moisture convergence across northern Australia. The increases in subsidence, low-level divergence, ground temperature, and air temperature and decreases in precipitable water, evapotranspiration, turbulent fluxes, and cloud cover are most marked in the early monsoon season and indicate a drier, more stable atmosphere. These findings are consistent with our additional long-term deforestation experiments over northern Australia with CCSM3.5, not discussed here.

[14] Given the shallow, relatively subdued nature of the northern Australian summer monsoon [*Hung and Yanai*, 2004], it could be expected that it is sensitive to land surface changes, especially during seasonal phases in which the large-scale monsoonal controls are less important. The changes that have been identified for the early/pre-monsoon season can be seen as a pre-conditioning process prior to the onset of the 'full' monsoon. Once the full monsoon is established, it becomes less likely that biophysical feedbacks can have a significant impact. During the active monsoon phases (bursts), large-scale circulation controls (e.g., Madden-Julian oscillation, cold surges linked to outflow from the East Asian winter monsoon, links with the Southern Hemisphere mid-latitudes), as well as stronger moisture sources [*Hung and Yanai*, 2004; *Wheeler and McBride*, 2005], are likely to overwhelm regional-scale changes, leading to a more muted response to biophysical changes.

[15] We find that a decrease in vegetation cover can delay the Australian monsoon and reduce early monsoon rainfall, and in this respect, our findings lend some support to the claim that Aboriginal vegetation burning practices over late Quaternary time-scales impacted the northern Australian summer monsoon regime [*Johnson et al.*, 1999]. However, our results clearly demonstrate that the effect on the peak monsoon was limited [*Pitman and Hesse*, 2007; *Marshall and Lynch*, 2008]. But we have equally demonstrated that biophysical feedbacks associated with reduced vegetation cover can have a clear impact on the early/pre-monsoon season, leading to a significant reduction in precipitation and essentially extending the dry season.

[16] The reduction in precipitation, in combination with the increased temperatures, can have significant impact on the biota, as broad biogeographical patterns in biodiversity show strong relationships with temperature and rainfall gradients, especially 'warmest period mean temperature' and 'precipitation seasonality' [*McKenzie et al.*, 2000]. It also follows that the strength of feedbacks can become of greater significance during years of reduced monsoon activity or during prolonged 'breaks' during the monsoon season – essentially invoking a 'Charney' type response. A prolonged dry season and higher temperatures during the pre-monsoon months could further promote the occurrence of late dry season 'hot burns,' which would carry significant ecological implications [*Russell-Smith*, 2002] and could have induced additional land-atmosphere feedbacks.

[17] The overall conclusion that emerges from this ensemble-based model study is that changes in vegetation cover can impact the regional hydrological cycle of northern Australia, particularly during the early/pre-monsoon months. During the peak monsoon, the more local-scale biophysical feedbacks provoked by the vegetation changes are less significant and the broader hemispheric-interhemispheric/land-ocean forcing mechanisms become dominant. This scenario is not all that different from that which occurs with the progression of the monsoon season, where rainfall linked to a local 'thunderstorm scale' characterizes the early/pre-monsoon season or breaks during the monsoon season. With the onset of low level westerlies, marking the onset of the monsoon over northern Australia, the monsoon trough is established over the region and precipitation is dominated by synoptic-scale clusters and bands embedded within the trough [*McBride*, 1998]. Our model results lead us to conclude that Aboriginal vegetation burning practices, while significantly affecting pre-monsoon events, did not have a

major impact on the late Quaternary summer monsoon of northern Australia. Our conclusions further prompt a fuller evaluation of the significance of Aboriginal burning practices for the Australian environment.

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