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Key Points:

- A recent interpretation of climate model output is for a "warmer is more arid" scenario
- Geological records over several glacial cycles suggest that "warmer is less arid"
- We use a new framework to reinterpret climate model output and conclude that "warmer is less arid"

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On the assessment of aridity with changes in atmospheric CO_2

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Abstract A recent interpretation of climate model projections concluded that "warmer is more arid." In contrast, dust records and other evidence have led the geoscience community to conclude that "warmer is less arid" leading to an aridity paradox. The "warmer is more arid" interpretation is based on a projected increase in the vapour pressure deficit ($\sim$ 7–9% K^{-1}) that results in a projected increase in potential evaporation that greatly exceeds the projected increase in precipitation. However, the increase in potential evaporation does not result in an increase in (actual) evaporation which remains more or less constant in the model output. Projected changes in the long-term aridity can be assessed by directly interrogating the climate model output. To that end, we equate lack of precipitation with meteorological aridity and lack of runoff with hydrologic aridity. A third perspective, agro-ecological aridity, is not directly related to the water lost but rather to the carbon gain and is equated with the reduction in photosynthetic uptake of CO_2. We reexamine the same climate model output and conclude that "warmer is less arid" from all perspectives and in agreement with the geological records. Future research will need to add the critical regional and seasonal perspectives to the aridity assessments described here.

1. Introduction

How will climatic aridity change in future as a consequence of the ongoing accumulation of greenhouse gases in the atmosphere? One can readily envisage that some regions will become more arid while some become less arid but what might be the overall trend in aridity? Some earlier work concluded that aridity would generally increase because of increasing air temperature [*Dai et al.*, 2004]. That work was based on a model that assumed the change in potential evaporation to be solely determined by change in air temperature. With that assumption, an increase in air temperature will cause increased potential evaporation that inevitably leads to a calculated increase in aridity. While that temperature-based approach has been widely used, it has long been known to be based on incorrect physics [*Jensen*, 1973; *Palmer*, 1965; *Palmer and Havens*, 1958]. In particular, parallel research has highlighted the flaws of temperature-based approaches for estimating potential evaporation when used for climate change assessments [*Hobbins et al.*, 2008; *McKenney and Rosenberg*, 1993; *Milly and Dunne*, 2011; *Rosenberg et al.*, 1989; *Sheffield et al.*, 2012] and the issues arising from temperature-based assessments have been highlighted and discussed in the recent IPCC Special Report on Extremes [*Seneviratne et al.*, 2012]. It is now recognized that one should apply physically based approaches with potential evaporation calculated as a function of radiation, temperature, humidity, and wind speed.

While the method for estimating potential evaporation has now been brought closer to standard agricultural, ecological and hydrologic practices, some of the earlier conclusions about increasing aridity in a warming climate have not changed in several recent publications. For example, a recent analysis concluded that aridity has increased over the global drylands since 1948 [*Feng and Fu*, 2013] while several other studies have concluded that world-wide increases in aridity will occur over the next century primarily due to the CO_2-induced increase in temperature [*Cook et al.*, 2014; *Feng and Fu*, 2013; *Fu and Feng*, 2014]. Based on those results it has recently been argued that an increase in aridity with increase in temperature is general and is simply a consequence of thermodynamics [*Sherwood and Fu*, 2014]. Those interpretations can be summarized with the statement: "warmer is more arid."

The above-noted interpretation that "warmer is more arid" raises some very important scientific questions. First, two recent global analyses of trends in aridity since 1948 concluded that some regions have become more arid while others have become less arid but with little overall change in terms of global averages and little obvious trend in aridity within the global drylands [*Greve et al*., 2014; *Sheffield et al*., 2012]. Those contradictory results draw attention to the underlying methodologies involved. The conclusion of an increase in aridity in the global drylands since 1948 was based on observed climate variables and particular subsequent calculations [*Feng and Fu*, 2013]. That study used the FAO-56 methodology to calculate potential evaporation [*Allen et al*., 1998] which is known to be more or less linearly related to pan evaporation [*Lim et al*., 2013]. Hence, the calculated increase in potential evaporation since 1948 reported by [*Feng and Fu*, 2013] stands in contrast to the observed decline in pan evaporation in many regions over roughly the same time period [*McVicar et al*., 2012; *Roderick et al*., 2009]. The dichotomy between increasing potential evaporation deduced from calculations and the decrease from direct observations has yet to be resolved.

A second contradictory result is that independent satellite observations show a general greening over many parts of the global drylands since the 1980s [*Dardel et al*., 2014; *Donohue et al*., 2009, 2013; *Herrmann et al*., 2005; *Piao et al*., 2005; *Young and Harris*, 2005]. This greening trend has been, at least in part, attributed to the biological impact of elevated atmospheric [CO_2] in dryland regions [*Donohue et al*., 2013]. That is important because the UNEP methodology [*Barrow*, 1992] used by *Feng and Fu* [2013] does not explicitly consider the direct biological impact of changes in atmospheric [CO_2] on the water use efficiency of vegetation [*Farquhar*, 1997].

An even bigger puzzle—a genuine paradox—awaits resolution. Analyses of 800,000 year old ice cores has revealed large increases in dust during cold glacial periods (atmospheric [CO_2] $\sim$180 ppm) with dust virtually absent during the warmer inter-glacial periods (atmospheric [CO_2] $\sim$280 ppm) [*Lambert et al*., 2008]. Those dust records, along with a range of other evidence, have led the geoscience community to the interpretation that the warmer (and higher atmospheric [CO_2]) inter-glacial periods are, in general, less arid than the colder (and lower atmospheric [CO_2]) glacial periods [*Mahowald et al*., 1999; *Muhs*, 2013]. We call this dichotomy between observations of the past interpreted as "warmer is less arid" versus climate model projections for the future interpreted as "warmer is more arid," the *global aridity paradox*.

At face value, there are at least two ways to resolve the global aridity paradox. The first is that there is no paradox and we accept both the observations of the past as well as the above-noted interpretation of the climate model projections. The conclusion is that the earth is currently at a global minimum with respect to aridity and that any change, either to warmer, or to cooler, conditions, would increase global aridity. An alternative approach is to follow the observation-based assessments of the geoscience community with their "warmer is less arid" interpretation. This implies a problem with either (i) the climate model output, or (ii) the interpretation of climate model output.

The aim here is to investigate this global aridity paradox. We begin with a brief overview of the aridity concept. We then summarize the basis for the recent interpretation that "warmer is more arid" and subsequently investigate that interpretation using the same climate model output. We find that the previous interpretation of climate model output does not support the proposition that warmer is more arid if projected changes in land precipitation and runoff are used as indicators of aridity. We propose a new flux-based framework for measuring long-term aridity that recognizes meteorological, hydrologic and agro-ecological perspectives and explicitly includes the biological impact of changing CO_2. We then use that framework and the same climate model projections to investigate the global aridity paradox.

2. Aridity—Some Background

In this section we first briefly summarize historical studies of aridity and then describe modern practice. We focus on the widely used aridity index approach.

2.1. Development of the Aridity Index Approach

Efforts to study climatic aridity are grounded in early research that sought to understand broad scale climatic controls on vegetation (and biome) distribution. For example, Köppen and Geiger proposed well-known aridity indices based on correlating spatial variations in the long-term averages of precipitation (P) and air temperature (T) with the spatial distribution of major vegetation types. [See *Tuhkanen* [1980] and *Maliva and Missimer* [2012] for an overview of historical developments in the field.) Subsequent work by *Thornthwaite* [1948] introduced the notion of potential evaporation (E_p) that was used instead of T to quantify the effectiveness of P in sustaining vegetation. Similar and almost parallel developments occurred in

Hydrology with *Budyko* [1958] using the ratio of P to net radiation (with the latter expressed as an equivalent depth of liquid water) as a measure of aridity. That latter index (or close variants) has been widely used and particularly effectively in the Hydrology and Land Surface community over the last 50 years [*Blöschl et al*., 2013]. More recently, the UNEP adopted the ratio of P to E_p as a measure of aridity in their landmark 1992 atlas on global desertification [*Barrow*, 1992]. In summary, the aridity index approach has been widely used in one form or another for at least the last 60 years.

2.2. Basis of the P/E_p Index

The basic idea underlying the aridity index approach is that P measures the supply and E_p the demand with the P/E_p ratio measuring how well the supply can meet the demand. To understand how the "warmer is more arid" interpretation was made we begin with the Penman-Monteith equation [*Shuttleworth*, 2012],

$$LE = \frac{\Delta(R_n - G) + \frac{\rho c_p D}{r_a}}{\Delta + \gamma(1 + \frac{r_s}{r_a})}, \tag{1}$$

where L ($\sim 2.4 \times 10^6$ J kg^{-1}) is the latent heat of vaporization, E (kg m^{-2} s^{-1}) the evaporation rate, Δ (Pa K^{-1}) the change in saturated vapour pressure with respect to temperature, R_n (W m^{-2}) the net irradiance, G (W m^{-2}) the ground heat flux, ρ (~ 1.2 kg m^{-3}) the air density, c_p (~ 1006 J kg^{-1} K^{-1}) the specific heat of air at constant pressure, D (Pa) the vapour pressure deficit of the air, r_a (s m^{-1}) and r_s (s m^{-1}) the aerodynamic and surface resistance, respectively, and γ (~ 67 Pa K^{-1}) the so-called psychrometric constant.

The current method used to estimate E_p is well known. Briefly, one measures all the radiative fluxes (to estimate R_n), air temperature (to estimate Δ and also to calculate the saturated vapour pressure that is one part of the calculation of D), wind speed (to estimate r_a) and the vapour pressure of the air that is used (along with the saturated vapour pressure) to calculate D. (See *McMahon et al*. [2013] for numerous worked examples.) With those data, the calculation then proceeds by assuming there is no supply limit and r_s is set to zero. With a modern computer it is easy to do the calculations but there are some conceptual difficulties with the standard procedure because E_p is a hypothetical flux. For example, assume we measure the radiation, temperature, wind speed and D in a dry and hot desert environment many months after the last rainfall. The surface resistance to water vapour flux will be substantial (e.g., $r_s >> 0$) under such dry conditions. We follow the standard procedure and input the meteorological observations into the equation and then set r_s equal to zero to calculate E_p. It is easy to see that if r_s were really zero (say there had been recent rainfall or perhaps irrigation) then the radiation, temperature, wind speed and D would likely be different from the measurements that were made when the surface was drier. The difference between actual ($r_s > 0$) and hypothetical ($r_s = 0$) conditions has led many to question the E_p concept [*Brutsaert and Stricker*, 1979; *Granger*, 1989; *McNaughton*, 1976; *Shuttleworth*, 2006, 2012; *Shuttleworth and Calder*, 1979; *Shuttleworth et al*., 2009; *Wallace*, 1995]. To further complicate matters, there have been a large number of approximations (e.g., ignore wind speed, ignore D, estimate radiation using T, etc.) in use over the last 60 years in various measures that are themselves often called potential evaporation in the literature [*Donohue et al*., 2010; *McMahon et al*., 2013]. Even specialists in the field can find it bewildering. As it turns out, the above-noted objections, while important from a scientific viewpoint, are not relevant to our reassessment of the "warmer is more arid" interpretation and are not discussed further.

3. Assessing Changes in P/E_p Over Time

In this section we first describe the recent research from which the "warmer is more arid" interpretation arose. We then describe why the "warmer is more arid" conclusion is not consistent with the climate model output and we explain the basis of that inconsistency.

3.1. Climate Model Projections of P/E_{ref}

The research described in the introduction used E_{ref} (instead of E_p). E_{ref} is calculated using the Penman-Monteith equation (equation (1)) but with prescribed surface properties appropriate to the assumed reference crop, i.e., a hypothetical well-watered agricultural crop with a canopy that completely covers the soil (fixed surface properties are: albedo = 0.23, $r_s = 70$ s m^{-1}, height = 0.12 m) [*Allen et al*., 1998]. For readers not familiar with surface resistances, we note that an r_s of 70 s m^{-1} is considered well watered. For a pure water surface (e.g., open water body like a lake), r_s is zero, while in many arid environments r_s routinely exceeds 5000 s m^{-1}.

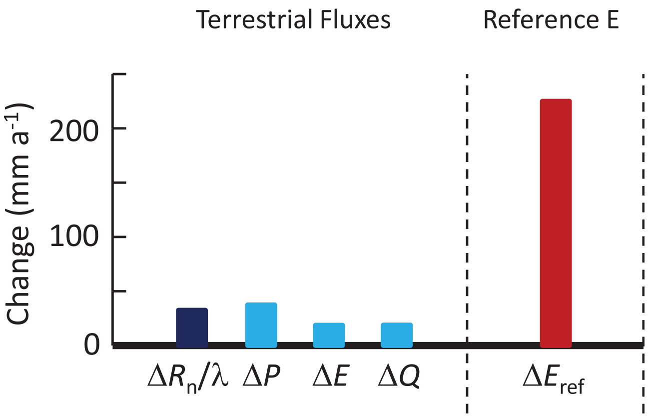


Figure 1. Projected changes in mean annual water cycle over the global land surface from the end of the 20th century (1970–1999) to the end of the 21st century (2070–2099). (left) (denoted Terrestrial fluxes) Projected changes in net irradiance expressed on a water-equivalent basis ($\Delta R_n/\lambda$, +39 mm a^{-1}), precipitation (ΔP, +41 mm a^{-1}), evaporation (ΔE, +20 mm a^{-1}) and runoff (ΔQ, +21 mm a^{-1}) [*Roderick et al.*, 2014]. (right) (denoted Reference *E*) The projected change in E_{ref} (ΔE_{ref}, +230 mm a^{-1}) [*Feng and Fu*, 2013].

(See *Shuttleworth et al.* [2009] for numerous examples of fluxes observed over a wide range of r_s.) To calculate E_{ref}, the output (radiation, temperature, wind speed and humidity) from 27 different climate models (CMIP5, RCP8.5 scenario) was averaged and E_{ref} calculated at each grid-box [*Feng and Fu*, 2013; *Fu and Feng*, 2014].

The CMIP5 projections for *P* show increases in some regions with decreases in others with a global average increase in annual *P* onto land of around 50 mm by the year 2100 [*Feng and Fu*, 2013]. (Those results for change in global land *P* are more or less identical to summaries based on the earlier CMIP3 (A1B scenario) models [*Lim and Roderick*, 2009; *Nohara et al.*, 2006; *Roderick et al.*, 2014].) In contrast, E_{ref} is projected to increase nearly everywhere with a global average increase in E_{ref} over land of around 230 mm a^{-1} by the year 2100 (RCP8.5) [*Feng and Fu*, 2013]. In most regions E_{ref} is projected to increase much faster than *P* and the ratio P/E_{ref} is projected to decrease nearly everywhere. The relation between P/E_{ref} and aridity was assumed to be invariant (i.e., constant over time and space) and to follow the UNEP aridity classification [*Feng and Fu*, 2013] (Hyper-arid: $P/E_{ref} < 0.05$, Arid: $0.05 < P/E_{ref} < 0.20$, Semiarid: $0.20 < P/E_{ref} < 0.50$, Subhumid: $0.50 < P/E_{ref} < 0.65$). With P/E_{ref} projected to decrease, a greater fraction of the land area fell into classes with lower P/E_{ref} and this has been interpreted as a projected increase in global aridity [*Feng and Fu*, 2013; *Fu and Feng*, 2014; *Sherwood and Fu*, 2014].

3.2. Why Is E_{ref} Projected to Increase?

The reasons for the projected increase in global land *P* are now well understood in terms of projected changes in the surface energy balance over land and ocean [*Roderick et al.*, 2014]. Hence the key to understanding the "warmer is more arid" interpretation is to understand why E_{ref} is projected to increase. Fortunately, the research has been comprehensively documented and the underlying reasons for the projected increase in E_{ref} (Figure 1) are also understood. E_{ref} will be most sensitive to changes in R_n, r_a and *D* [*Donohue et al.*, 2010; *Fu and Feng*, 2014; *Roderick et al.*, 2007; *Scheff and Frierson*, 2014]. In terms of R_n, one anticipates an enhancement of the greenhouse effect to increase the incoming longwave irradiance at the surface. However, one poorly understood aspect of the climate model projections is that this increase is almost entirely dissipated at the terrestrial surface by a more or less equal increase in outgoing longwave irradiance (thereby increasing the surface temperature) with little change in either R_n or any of the other surface fluxes [*Roderick et al.*, 2014]. For example, in the CMIP3 archive, the change in R_n to the end of the 21st century (A1B scenario), when averaged over the land surface is around +3 W m^{-2} which equates to +39 mm a^{-1} (mm per annum) when expressed on a water-equivalent basis (Figure 1). The CMIP5 (RCP8.5) results for R_n are more or less identical [*Fu and Feng*, 2014; *Scheff and Frierson*, 2014].

In terms of the change in r_a, the CMIP5 ensemble average projects a slight decrease in wind speed over land (and therefore an increase in r_a [*Shuttleworth*, 2012]) that decreases E_{ref} but the overall effect on the E_{ref} projections is small [*Fu and Feng*, 2014; *Scheff and Frierson*, 2014]. The overwhelming reason for the projected increase in E_{ref} is the projected increase in *D* in the climate model output [*Fu and Feng*, 2014; *Scheff and Frierson*, 2014]. Earlier work with global (land plus ocean) averages noted that climate models project the relative humidity of near-surface air to remain more or less constant as warming proceeds [*Held and Soden*, 2000, 2006]. This requires an increase in *D* that follows Clausius-Clapeyron scaling of around 7% K^{-1}. Subsequent detailed investigation has revealed a land-ocean difference with slight increases in (global average) relative humidity projected over the ocean and an almost complementary decrease in relative humidity projected over land (~ 0.7% K^{-1}) with warming [*Fu and Feng*, 2014]. Hence over land the projected increases in D are slightly larger than Clausius-Clapeyron scaling and are around 7–9% K^{-1} and depend slightly on the initial relative humidity (see Appendix A). The larger increase in T over land relative to the ocean and the projected increase in D ($\sim$7–9 % K^{-1}) underpin the interpretation that an increase in aridity with warming is a simple thermodynamic consequence of warming [*Sherwood and Fu*, 2014].

3.3. Interpreting the Projected Increase in E_{ref}

E_{ref} specifically refers to the evapotranspiration from a hypothetical well-watered crop that completely covers the ground with canopy properties including the surface resistance (= 70 s m^{-1}) that remain fixed over time. On that basis, we expect the results to apply to the evaporation from any moist surface whose surface resistance remains constant over time. For example, lake evaporation (whose surface resistance is zero and also fixed over time) is anticipated to increase more or less in line with E_{ref} [*Lim et al.*, 2013]. The consequence is that open water bodies at the surface are projected to evaporate faster and would therefore empty faster if all else (e.g., precipitation, runoff) were equal. However, that does not necessarily equate to an increase in E over the entire landscape which is usually (but not always) dominated by vegetation/soil and not by open water bodies. Vegetated (and soil) surfaces are different from open water bodies (and the hypothetical reference crop) because their surface resistance changes over time.

The key to understanding changes in aridity over the entire landscape is to consider the change in E in relation to the change in P and E_{ref}. The implicit assumption underlying the "warmer is more arid" interpretation is that E follows E_{ref} and E will therefore increase faster than P leading to a reduction in runoff. However, the climate model projections are for the opposite with E increasing more slowly than P leading to a (slight) increase in runoff (Figure 1). In these calculations the evaporative drivers (R_n, r_a, D, T) for E and for E_{ref} are the same which implies that the surface resistance retarding E must be increasing over time in the climate model projections.

The climate models being considered here do not as yet allow the vegetation canopies to adjust to the new atmospheric conditions. Hence they assume fixed canopies (e.g., a seasonally varying leaf area index that repeats from one year to the next) but the stomatal conductance of leaves within the canopy will decrease as atmospheric [CO_2] increases [*Betts et al.*, 2007; *Sellers et al.*, 1996] (also see Appendix B). With no capacity in the model to change the canopy leaf area, the increase in atmospheric CO_2 would by itself increase the surface resistance (in those models).

3.4. Space for Time Substitution

The ratio P/E_{ref} (or similar indices) has been widely used to assess *spatial* differences in aridity at a given instant of time. The UNEP 1992 Desertification Atlas [*Barrow*, 1992] is a classic example of the spatial approach where regions with lower P/E_{ref} values are considered more arid. Importantly, in those spatial comparisons, the atmospheric [CO_2] is fixed. However, climate change that is caused by changes in atmospheric [CO_2] introduces an additional source of variation not present in spatial comparisons which invalidates the space for time substitution approach [*Gerhart and Ward*, 2010].

4. Assessing Changes in Aridity Over Time

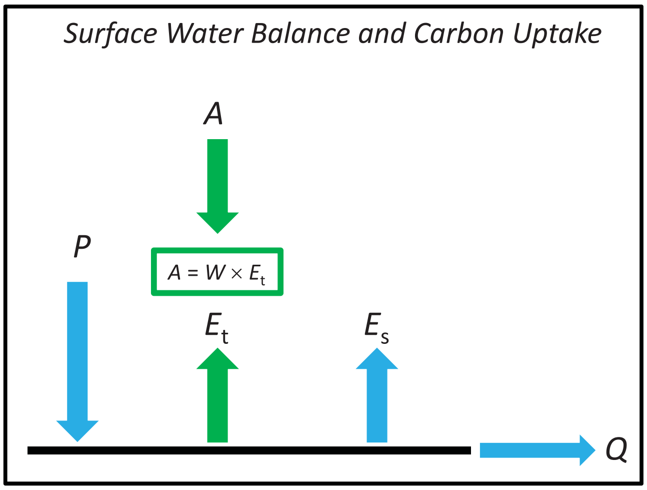
In this section, we first outline a flux-based approach for assessing aridity that incorporates meteorological, agro-ecological and hydrologic perspectives and avoids the above-noted problems with the space for time substitution. We then use that approach to assess changes in aridity over a large range of atmospheric [CO_2] and thereby address the global aridity paradox.

4.1. A Flux-Based Approach for Assessing Changes in Aridity

We begin with the usual water balance equation,

$$\frac{dS}{dt} = P - (E_t + E_s) - Q, \tag{2}$$

where the rate of change in water storage (dS/dt) is determined by inputs of precipitation (P) and outputs of evaporation (E) and runoff (Q). The total E is separated into two components, (i) transpiration (E_t) and (ii) a residual term that includes all other sources of evaporation (E_s). Note that E_s includes fluxes such as evaporation from soil and from wet canopies, open water bodies, etc. As a guide, the transpiration fraction (E_t/E)



	MEASURES OF ARIDITY
ΔP	change in meteorological aridity
ΔQ	change in hydrological aridity
ΔA	change in agro-ecological aridity

Figure 2. Steady state surface water balance (P, E, Q) and measures of climatic aridity. Evaporation (E) is split into two components, soil evaporation (E_s) and transpiration (E_t) with that latter flux coupled to the (net) photosynthetic uptake (A) via the water use efficiency (W).

roughly scales with vegetation cover and varies from near zero (e.g., in permanent snow/ice covered regions and extreme deserts) up to perhaps 80 or 90% in tropical evergreen forests. The global average transpiration has recently been estimated to be roughly 60% of total E [*Schlesinger and Jasechko*, 2014].

For the time scales of relevance to aridity ($>$ 30 years) we assume steady state conditions,

$$P \approx E_t + E_s + Q, \tag{3}$$

and equate fluxes in the water balance to various perspectives on aridity. We use P as the negative measure of meteorological aridity and Q for hydrologic aridity (Figure 2). For agro-ecological aridity we follow Thornthwaite's original concept of the effectiveness of P to support vegetation [*Thornthwaite*, 1948]. The key here is that aridity from the point of view of vegetation is determined by vegetation productivity. Here we use the (net) photosynthetic rate (A) (also sometimes called Gross Primary Productivity or GPP) as the flux that measures agro-ecological aridity (Figure 2). With the water use efficiency (W, see Appendix B for a more complete discussion) defined by,

$$W = \frac{A}{E_t}, \tag{4}$$

the steady state water balance can be rewritten as,

$$P \approx \frac{A}{W} + E_s + Q. \tag{5}$$

It may seem a little unusual to readers not familiar with carbon uptake to express the transpiration component of the water balance in terms of the carbon uptake but this formulation is actually widely accepted [*Berry et al*., 2010; *Monteith*, 1988; *Wong*, 1979]. Further, most process-based climate and vegetation models do in fact estimate E_t as a function of A [*Bonan*, 2008; *Leuning*, 1995; *Woodward*, 1987].

On this framework, an environment would become more arid from a meteorological viewpoint if P decreased. The same environment would be more arid from hydrologic and agro-ecological viewpoints if Q and A were to decrease.

4.2. Projected Changes in Aridity Over Glacial Time Scales

Simulating the climate response to CO_2-induced warming is computationally expensive making it less attractive to do the very long runs needed to examine glacial-inter-glacial cycles using comprehensive fully coupled climate models. Scientists from the NASA Goddard Institute for Space Studies (GISS) have recently developed a new approach than can rapidly estimate the *equilibrium climate response* to an imposed CO_2 forcing [*Russell et al*., 2013]. With substantially lower computational costs the GISS group have estimated the equilibrium climate response over an exceptionally large range of imposed atmospheric CO_2 (C_a) concentrations (C_a: 78–79,872 ppm). This range substantially exceeds the glacial ($C_a \sim 180$ ppm) to inter-glacial

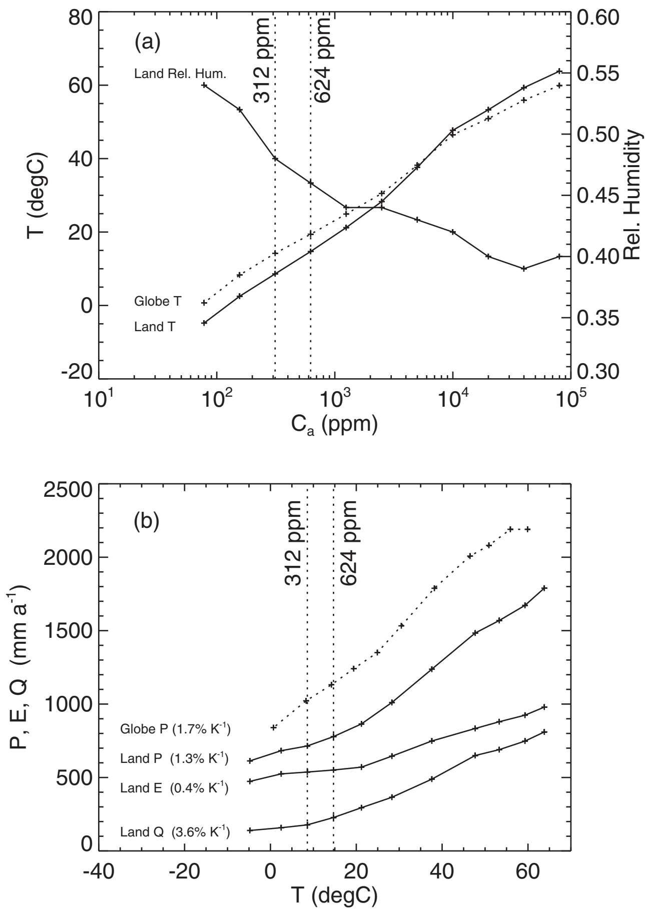


Figure 3. Equilibrium climates for a large range of atmospheric [CO_2] (C_a varies 78–79,872 ppm) as simulated by a modified version of the GISS climate model [*Russell et al.*, 2013]. Dashed vertical lines show C_a in the year 1950 (312 ppm) and for the subsequent doubling (624 ppm). (a) (left axis) Near surface air temperature for the globe/land and (right axis) relative humidity of the near surface air over land plotted as a function of C_a. (b) Water fluxes (P, E, Q) for globe/land as a function of near surface air temperature. Sensitivities (e.g., 3.6% K^{-1} for Q, etc.) are calculated for the change in C_a from 312 to 624 ppm. (All model outputs were provided by Dr Gary L. Russell from GISS.)

($C_a \sim$ 280 ppm) ranges experienced over the last million years [*Lambert et al.*, 2008] and also covers current (C_a $\sim$ 400 ppm) levels as well as near term projections (e.g., $C_a \sim$ 800–1000 ppm toward the end of the 21st century). As such, the output from these climate model runs offers a unique opportunity to investigate the global aridity paradox.

Unfortunately this version of the GISS model did not simulate A so we are only able to assess projected changes in meteorological (ΔP) and hydrologic (ΔQ) aridity with changes in global mean T and C_a. We also note that the model does not explicitly make the leaf-scale conductance for water vapour dependent on C_a (G. L. Russell, personal communication, 2015). The consequence is that any projected increase in runoff cannot be attributed to the direct biological effect of increasing C_a reducing leaf-scale transpiration [e.g., *Gedney et al.*, 2006]. In their original publication the GISS group concluded that the land surface generally becomes more arid as both C_a and surface T increases [*Russell et al.*, 2013]. The basis for that interpretation was their finding that over land, the relative humidity of the near-surface air decreases slightly with the C_a-induced increase in T (G. L. Russell, personal communication, 2013–2014). We use outputs from the same model runs and reinterpret their results in terms of meteorological and hydrologic aridity (Figure 2) as defined here.

The results show both land and global T increasing monotonically with C_a (Figure 3a). The change in equilibrium T going from a C_a of 312 to 624 ppm is 5.2 K for the globe and 6.1 K for the land putting the results at the upper end of current climate sensitivity estimates [*Otto et al.*, 2013; *Roe and Baker*, 2007]. Over land, the relative humidity of near-surface air declines as C_a increases (from 312 to 624 ppm) but that change is very small (Figure 3a) and D closely follows Clausius-Clapeyron scaling as noted previously for both CMIP3 and CMIP5 multimodel ensembles. In terms of the fluxes, P increases faster than E so that Q steadily increases as C_a and T increase. On that basis, we conclude for the global land average, that "warmer is less arid" from both meteorological and hydrologic perspectives.

As noted previously, the carbon uptake (A) is not available from the GISS model runs. Previous research using a suite of global vegetation models projected that A would increase by 42% from the year 2000 (when C_a is 390 ppm) to the year 2100 (when C_a was assumed to be 790 ppm) [*Cramer et al.*, 2001]. (That paper reported Net Primary Productivity (NPP) instead of A and we assumed NPP is half of A to calculate the above-noted percentage change.) This is equivalent to an increase in A of 41% for a doubling of C_a. In their study, the global mean T was assumed to increase by around 4 K to the year 2100 implying a sensitivity of (global land) A to the CO_2-induced increase in T (= 41%/4.0 K) of around 10% K^{-1}. That research also noted that the impact of climate change alone (i.e., holding C_a fixed) was for a small decrease in A which underlines the fundamental importance of changes in C_a on A [*Cramer et al*., 2001].

Subsequent research has not substantially altered that estimate. For example, a recent synthesis of the output from eight earth system models in the CMIP5 archive (of which six allowed the vegetation canopies to adapt to changes in C_a and other atmospheric conditions) projected an increase in A by 2100 of around 20% for an increase in C_a of around 49% (RCP4.5 scenario) [*Shao et al*., 2013]. This implies an identical increase in A of around 41% for a doubling of C_a.

We note that the response of A to increased C_a and the associated C_a-induced warming ($\sim$ 10 % K^{-1}) is much larger than the response for the water cycle fluxes to the same forcing (Figure 3b, P; 1.3% K^{-1}: E: 0.4 % K^{-1}; Q: 3.6% K^{-1}). The larger relative response in A is anticipated because C_a is a primary substrate for photosynthesis (see Appendix B for more details) and there is a direct biological response of A to changes in C_a [*Beerling and Woodward*, 1993; *Farquhar*, 1997; *Polley*, 1997; *Polley et al*., 1997]. With those results we conclude that "warmer is less arid" also holds on a global average basis from the agro-ecological perspective when that warming is induced by increases in C_a.

5. Discussion

5.1. Reassessment of Previous Results

The basis for the previous interpretation that aridity will increase with future CO_2-induced warming [*Sherwood and Fu*, 2014] was that the projected increase in potential evaporation (E_{ref}) was substantially larger than the projected increase in precipitation (P) over land (Figure 1). The implicit assumption in that interpretation was that the (total) evaporation (E) would follow E_{ref} and increase substantially faster than P leading to a more arid environment with less runoff (Q). The major problem with that interpretation is that the same climate models project global land P to increase faster than E leading to more Q (Figure 1). In short, in the climate model output, E does not follow the increase in E_{ref}.

It is important to fully trace the implications of the model projections back to the coupled surface water-energy balance. In particular, evaporative fluxes cool the surface (by removing heat) and the small increase in E (Figure 1) means a small cooling effect. The consequence is that the projected increase in incoming longwave radiation at the surface (i.e., the enhanced greenhouse effect) has to be dissipated by other pathways. In climate model output the dissipation pathway over land is almost entirely due to an increase in outgoing longwave radiation (with an associated increase in surface temperature) leaving little change in net radiation at the surface [*Roderick et al*., 2014]. The consequence of the small projected increase in E coupled with a much larger increase in outgoing longwave irradiance (and hence an increase in surface temperature) is important to fully comprehend. For example, the view often communicated to other scientists and to the wider public is that an increase in near-surface air temperature will cause increased evaporation. This is an unfortunate starting point for communicating hydro-climatic projections because it is not always true. (What is true is that as temperature increases, the saturated vapour pressure increases but that is a state variable, and not a flux, like evaporation.) The climate model projections are much easier to interpret and communicate by starting with the more general proposition (i.e., energy balance) that an increase (decrease) in E will cool (warm) the surface. The resulting interpretation is that increased warming projected over land relative to the ocean is not a cause of increased aridity but is instead the result of the smaller increase in E over land relative to the ocean [*Boer*, 1993; *Roderick et al*., 2014; *Sutton et al*., 2007].

Most of the increase in E_{ref} was projected to occur because of an increase in the vapour pressure deficit (D) of the near-surface air. Interestingly, if one follows the original Budyko scheme [*Budyko*, 1958] by using net radiation as the energetic constraint (instead of E_{ref}) on E then the aridity index approach does make predictions for partitioning the change in P (between E and Q) that more or less replicates the climate model output [*Roderick et al*., 2014]. On that basis it seems likely that the projected increase in D (that causes the projected increase in E_{ref}) should not be considered a forcing by the atmosphere on the surface but should perhaps be considered a feedback from the surface to the atmosphere. It may eventually prove possible to directly separate the forcing from the feedback perhaps using approaches similar to those recently developed for the analysis of forcing-feedback relations during transient meteorological droughts [*Yin et al*., 2014].

5.2. Comment on the Aridity Index Approach

The utility of the aridity index approach is not based on the numerical value of the index. On the contrary, the utility is derived when the index is used for the intended purpose, i.e., partitioning the mean annual P between E and Q [e.g., *Budyko*, 1958; *Milly*, 1994; *Roderick and Farquhar*, 2011]. The aridity index approach is currently in widespread use because comprehensive measurements of key water balance components (e.g., evaporation, runoff, soil moisture, etc.) are generally not available [*Blöschl et al.*, 2013]. In the absence of measurements the aridity index approach has a tendency to become an end in itself where the aim is to predict the aridity index and the change in that index. When using climate models, it is important to remember that one does not need to use an index (of any kind) to partition P between E and Q because that partitioning has already been done within the model. Of course the climate model partitioning may not be accurate but that is a separate question requiring rigorous and comprehensive model evaluation.

5.3. Resolving the Aridity Paradox

Long-term ($>$ 30 years) climatic aridity has been assessed here by asking three questions, (i) how much precipitation?, (ii) how much runoff?, and (iii) how productive is the vegetation? To address those questions we equated lack of P with meteorological aridity and lack of Q with hydrologic aridity. Vegetation productivity is equivalent to the carbon gain and we measured agro-ecological aridity using the lack of (net) photosynthetic rate (A) that is the lack of Gross Primary Productivity.

To evaluate the aridity paradox we used previously published model runs that specifically examined the equilibrium hydro-climatic response to an exceptionally large range of atmospheric [CO_2] (C_a) [*Russell et al.*, 2013]. We found that P ($+1.3\%$ K^{-1}) and Q ($+3.4\%$ K^{-1}) both increase more or less monotonically with T as C_a increases (Figure 3b). The sensitivity of that response is broadly similar to that previously documented using CMIP3 and CMIP5 models [*Lim and Roderick*, 2009; *Nohara et al.*, 2006; *Roderick et al.*, 2014; *Zhang et al.*, 2014]. On that basis we conclude that in terms of the global average, that "warmer is less arid," at least from meteorological and hydrologic perspectives. We note that this conclusion is based on the output from one model and we await confirmation from a larger range of models.

While projections for A were not available from the same model, previous modeling has projected a substantial increase in A ($+40\%$) with a doubling of C_a [*Cramer et al.*, 2001; *Shao et al.*, 2013]. At face value, those results suggest that the increase in C_a over the coming century will likely result in a substantial decrease in the globally averaged agro-ecological aridity. Indeed, there is already direct evidence for a CO_2-induced greening of global vegetation [*Donohue et al.*, 2013].

5.4. Comment on the Agro-Ecological Aridity Assessment

The use of A to measure agro-ecological aridity will implicitly include a great many processes that go well beyond hydro-climatology but are more familiar in agriculture, ecology and forestry. For example, on the basis of the definition we have used, any management practice that increases transpiration (and therefore A) at the expense of soil evaporation (e.g., mulching, minimum till, etc.) will reduce agro-ecological aridity. The photosynthetic uptake is also sensitive to soil properties and nutrient availability. In that sense, agro-ecological aridity is not solely determined by hydro-climatic factors and one can readily incorporate soil factors, nutrient availability and other management activities [*Le Houérou*, 1984, 1996; *Reynolds et al.*, 2007] within the same aridity framework.

The above discussion highlights the potential role of nutrient availability in moderating the response of A to C_a. Many earth system models do not yet include nutrient cycling and it has been argued that the response of A to increasing C_a may well be restricted by nutrient availability [*Hungate et al.*, 2003]. We emphasize that to date, there is widespread evidence for the stimulation of A by the ongoing increase in C_a [*Drake et al.*, 1997; *Frank et al.*, 2015; *Franks et al.*, 2013; *Hungate et al.*, 2013; *Norby and Zak*, 2011]. However, we cannot exclude the possibility that in future, nutrient constraints will limit the response of A to increasing C_a [*Peñuelas et al.*, 2011]. In our experience there appears to be some confusion about the role of nutrient constraints and the associated hydrologic implications. It is important to understand that a nutrient constraint that limits the response of A does not also mean that changes in C_a will have no effect on aridity. On the contrary, there will still be a large impact on aridity but a nutrient constraint on A will alter how the change in C_a is ultimately expressed.

To understand the possibilities we consider two extremes. For the first extreme we follow the model projections with A increasing by around 40% for a doubling of C_a. The water use efficiency increases with C_a but decreases with D. For a doubling of C_a, the projected relative increase in D is much smaller ($\sim$+30%) than the relative increase in C_a (+100%) and we estimate a substantial increase in water use efficiency (see Appendix B). The increase in A and in the water use efficiency, are, at least very roughly, of the same order and we would expect a smaller change in transpiration which would ultimately translate into a smaller change in runoff for a given P. Hence a strong response of A to C_a will substantially reduce agro-ecological aridity and on the above assumptions have a smaller impact on hydrologic aridity. At the other extreme, assume that nutrient constraints are so complete that A does not respond to a doubling in C_a. (We emphasize that there is limited empirical support to date for this possibility and we use it for illustrative purposes only.) However, the future fluxes (photosynthesis, transpiration) will occur in an atmosphere that has much higher C_a and there remains a substantial increase in water use efficiency. The combination of no change in A with a large increase in C_a and hence water use efficiency would substantially reduce transpiration and ultimately lead to a substantial increase in runoff for a given P. The key point is that the possibility of a nutrient constraint on A does not mean that there is no change in terrestrial aridity as C_a increases. On the contrary, nutrient constraints will shift the response to increasing C_a between A (agro-ecological aridity) and Q (hydrologic aridity) via changes in transpiration.

5.5. Weaknesses of the New Aridity Assessments

One obvious weakness with the approach used here is that we focus solely on the globally averaged mean annual water balance and have ignored the critical regional, seasonal and inter-annual perspectives. For example, in many circumstances it is often more important to know the seasonal timing of P and Q [e.g., *Kumar et al*., 2014] and a seasonal perspective (that includes variations in soil moisture) will need to be added to the scheme proposed here. Thinking even further ahead, an assessment of changes in extremes (that includes soil moisture variations) resulting from inter-annual variability would also be welcome. Regional assessments are also needed and we already know that the projections for many areas will reveal both increases/decreases in aridity.

Further weaknesses are more related to the underlying modeling. The first is that many (but not all) climate models do not allow the vegetation to adjust to changing climate or atmospheric conditions, e.g., change leaf area index, etc. This is a clear shortcoming as highlighted by recent modeling [*Schymanski et al*., 2015] and observational studies [*Donohue et al*., 2013; *Frank et al*., 2015]. Models also do not yet incorporate major anthropogenic impacts such as irrigation and groundwater extractions. Perhaps they need not. But, we note that current groundwater extractions [*Yoshihide and Bierkens*, 2014] and other human-induced perturbations to the surface hydrology are in many cases already larger than many of the regional changes being projected by climate models [*Grafton et al*., 2013]. Those existing perturbations must always be kept in mind when evaluating the functional significance of projected hydro-climatic changes.

5.6. Concluding Remarks

Our conclusion is that in terms of the global average, "warmer is less arid" from meteorological, hydrologic and agro-ecological perspectives, at least when that warming is induced by elevated CO_2. Further analysis using a greater range of models is recommended.

What has proved a little surprising is that the projections of global average meteorological and hydrologic sensitivity to warming are rather small (Figure 3b, P; $+1.3\%\ K^{-1}$: E; $+0.4\%\ K^{-1}$: Q; $+3.4\%\ K^{-1}$). The source of that low sensitivity can be traced to the small projected changes in the surface net radiation [*Roderick et al*., 2014]. Whether that holds in reality is a separate (and critically important) question but for the moment let us assume it to be true. Let us further assume a doubling of C_a results in warming of say 3 K (IPCC range: $\sim$ 2–5 K) [*Otto et al*., 2013; *Roe and Baker*, 2007]. Using the above sensitivities implies that the globally averaged changes in P and Q over land for a doubling in C_a would be around +4% and +10% respectively. The concurrent increase in A is projected to be much larger at around +40%. With that contrast in mind, the large increases in dust that have previously been interpreted to indicate greater aridity during colder glacial periods [*Muhs*, 2013] may be more a function of changes in vegetation productivity and abundance due to the direct biological impact of changes in atmospheric CO_2 [*Beerling and Woodward*, 1993; *Franks et al*., 2013; *Gerhart and Ward*, 2010; *Prentice and Harrison*, 2009] than to changes in vegetation caused by changes in the hydro-climate [*Yung et al*., 1996]. Time will tell if this interpretation proves to be correct.

Appendix A: Understanding Changes in D

D is defined as the difference between the saturated vapour pressure at air temperature (e_s) and the (actual) vapour pressure (e_a) of the air,

$$D = e_s - e_a. \tag{A1}$$

Using the relative humidity h ($= e_a/e_s$) we rewrite equation (A1) as,

$$D = e_s - h\, e_s. \tag{A2}$$

The change in D with T can be expressed in a relative form by,

$$\frac{dD}{dT}\frac{1}{D} = \frac{de_s}{e_s dT}\frac{e_s}{D} - \frac{de_s}{e_s dT}\frac{h\, e_s}{D} - \frac{dh}{dT}\frac{e_s}{D} = \frac{de_s}{e_s dT} - \frac{dh}{dT}\frac{1}{(1-h)}. \tag{A3}$$

The first term on the right side $(de_s/(e_s dT))$ is the standard Clausius-Clapeyron scaling ($\sim$7 % K^{-1} at current earth T) while the second term $(dh/(dT \times (1-h)))$ describes how D changes with h. The CMIP5 climate model projections are for h to decrease slightly over land ($\sim$ 0.7% K^{-1}) [*Fu and Feng*, 2014]. Using equation (A3), for a typical arid environment (e.g., assume $h = 0.20$) we expect a relative change in D (= 7 % K^{-1} – (−0.7 % K^{-1} × 1/(1–0.2))) of $\sim$ 7.9 % K^{-1}. For a typical humid environment (e.g., assume $h = 0.7$) we expect a slightly larger change in D (= 7 % K^{-1} – (−0.7% K^{-1} × 1/(1–0.7))) of around 9.3 % K^{-1}. This brackets the 7–9% K^{-1} range for D given in the main text.

Appendix B: Water Use Efficiency of Photosynthesis

The leaf scale (net) photosynthetic rate (A_L) can be expressed as the product of the conductance g_c for CO_2 and the difference in CO_2 concentration between the ambient air (C_a) and the inter-cellular air spaces (C_i) inside the leaf,

$$A_L = g_c(C_a - C_i). \tag{B1}$$

An expression for leaf scale transpiration ($E_{t,L}$) can be written in the same form using the conductance for water vapour (g_w) and the difference in vapour pressure between the inter-cellular spaces ($e_s(T_L)$, which is assumed to be saturated at leaf temperature, T_L) and the ambient air (e_a),

$$E_{t,L} = g_w(e_s(T_L) - e_a). \tag{B2}$$

In air, water vapour diffuses faster than CO_2 and g_w and g_c are related by the ratio of their diffusivities ($g_w = 1.6\, g_c$) [*Cowan*, 1977]. On that basis the leaf scale water use efficiency (W_L) is given as [*Wong et al.*, 1979],

$$W_L = \frac{A_L}{E_{t,L}} = \frac{C_a}{1.6\, v}\left(1 - \frac{C_i}{C_a}\right), \tag{B3}$$

with v ($= e_s(T_L) - e_a$) the leaf-air vapour pressure difference. We assume the leaf-to-air vapour pressure difference can be approximated by the vapour pressure deficit of the air ($D = e_s(T_a) - e_a$) which is equivalent to assuming the leaf is at air temperature (but see *Helliker and Richter* [2008] for an alternate view). With that approximation, the relative change in W_L is given by,

$$\frac{dW_L}{W_L} \approx \frac{dC_a}{C_a} - \frac{dD}{D} + \frac{d(1-\frac{C_i}{C_a})}{(1-\frac{C_i}{C_a})}. \tag{B4}$$

The ratio, C_i/C_a tends to be conservative for a given photosynthetic pathway (C_3 plants: $C_i/C_a \sim 0.7$; C_4 plants: $C_i/C_a \sim 0.3$) [*Wong et al.*, 1979; *Jones*, 1992]. Observations show a slight increase in ($1 - C_i/C_a$) with an increase in D that has been modeled as [*Farquhar et al.*, 1993],

$$\left(1 - \frac{C_i}{C_a}\right) \propto \sqrt{D}, \tag{B5}$$

in agreement with measurements on whole trees [*Wong and Dunin*, 1987]. With that approximation, equation (B4) is rewritten as [*Donohue et al.*, 2013],

$$\frac{dW_L}{W_L} \approx \frac{dC_a}{C_a} - \frac{1}{2}\frac{dD}{D}. \quad \text{(B6)}$$

This represents a useful summary provided the photosynthetic pathway (C_3, C_4) of the vegetation remains unchanged.

The key point is that water use efficiency increases with C_a but decreases with D. Hence there is a competition and the net effect will depend on the magnitudes of the changes in C_a and D. On a year to year basis there are small changes in C_a and the larger changes in D will dominate. However, over a longer term, e.g., decadal to century time scales and beyond, the changes in C_a are expected to dominate over changes in D. To take a typical example using data from the main text, assume a doubling of C_a leads to a T increase of say 3 K. With the projected sensivity of D over land ($\sim$ 7–9% K^{-1}, see Appendix A) assumed to be 9% K^{-1}, the relative increase in water use efficiency is ($=$ 1.0 – (0.5 $\times$ 3 K $\times$ 9% K^{-1}) $=$ 1.0 – 0.14 $=$ 0.86) around 86% for a doubling of C_a. Even if one were to assume a 6 K increase for a doubling of C_a, the water use efficiency would still increase substantially (for 6 K, we expect $+$73%). (Note that the magnitude of those changes are larger than can be accurately calculated with the first-order formulation used here. In that sense, the estimates are indicative only.) The key point is that based on current underdstanding of the various sensitivities, water use efficiency is expected to increase substantially from glacial ($C_a \sim$ 180 ppm) to inter-glacial ($C_a \sim$ 280 ppm) periods and is also expected to continue to increase from the current $C_a \sim$ 400 ppm well into immediate future.

The photosynthetic pathway is important because the lower C_i/C_a for C_4 plants (that are mostly tropical grasses) confers an advantage in terms of higher water use efficiency (compared to C_3 plants) and this is thought to be a major reason for the increased abundance of C_4 plants during glacial periods when C_a is low [*Bond*, 2008; *Ehleringer et al*., 1997; *Prentice and Harrison*, 2009; *Sage and Stata*, 2015; *Street-Perrott et al*., 1997]. Along similar lines, the increase in C_a over the last century has been implicated as one reason for the observed increase in woody plant cover (C_3 plants) in many arid regions [*Berry and Roderick*, 2002; *Bond and Midgley*, 2000; *Donohue et al*., 2013] although the ecological dynamics occurring during that grass-to-tree transition are complex and still not completely understood [*Beerling and Osborne*, 2006; *Bond and Midgley*, 2012; *Higgins and Scheiter*, 2012; *Scholes and Archer*, 1997].

Scaling up the leaf scale physiological relations to a ground area basis requires a simultaneous understanding of how the leaf area per ground area changes. This is a complex problem [*Bonan*, 2008; *Woodward and Lomas*, 2004; *Woodward*, 1987] that is well beyond the scope of the research in the current manuscript. However, in sparse canopies typical of arid environments there is minimal self-shading and the relations expressed on a ground area basis will more or less follow the leaf-scale relations [*Donohue et al*., 2013].

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