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Supplement of

A review of green- and blue-water resources and their trade-offs for future agricultural production in the Amazon Basin: what could irrigated agriculture mean for Amazonia?

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1 Global evapotranspiration

Estimates of global evapotranspiration (ET) over terrestrial surfaces since the early 1990s range from 62,262 km³ y⁻¹ to 67,900 km³ y⁻¹ (416 mm y⁻¹ to 650 mm y⁻¹) and differ based on the methodology used and the timeframe considered (Table S1). These values represent 64–69% of global terrestrial precipitation based on global assessments (Oki and Kanae, 2006). Climate change is expected to accelerate the hydrological cycle and increase evaporative demand due to increases in surface temperatures and radiative forcing (Jung et al., 2009). Analyses of time series from 1982 to 2008 show increases in global ET at a rate of 7.1 ± 1 mm y⁻¹ decade⁻¹ between 1982 and 1997, followed by a slowdown between 1998 and 2008 of -7.8 mm y⁻¹ decade⁻¹ attributed to water limiting conditions in Africa and Australia (Jung et al., 2009). Despite this slowdown, Jung et al. (2009) show a 26-year increase in global ET of 0.41 ± 0.20 mm y⁻², the smallest estimate in a series of assessments which include 0.72 ± 0.21 mm y⁻² for 1983–2006 (Zhang et al., 2010), 1.10 ± 0.20 mm y⁻² for 1982–2009 (Zeng et al., 2012) and 1.18 mm y⁻² (Mao et al., 2015) for 1982–2010. Despite refutation of the global decrease by Zeng et al. (2012), two of the three methods predict a significant increase in continental ET for South America between 1983 and 2006 at 3.21 ± 1.00 mm y⁻² (Zhang et al., 2010) and 2.33 ± 0.71 mm y⁻² (Zeng et al., 2012) (Table S2).

The apparent decline in global ET and the possible increase in South America raise questions about continental and regional ET responses, and drivers of the terrestrial contributions to the atmospheric water balance. Such a decline may be exacerbated by and lead to a drop in global precipitation should the amount of water vapour be sufficiently restricted. The four main drivers of this possible drop in global ET are: (1) the increase in atmospheric CO₂ concentration, (2) the effects of land use change, (3) an increase in nitrogen deposition, and (4) a drop in soil moisture stocks (Jung et al., 2009; Mao et al., 2015). Increased atmospheric CO₂ concentrations are known to increase water use efficiency in terrestrial ecosystems by affecting stomatal closure, thus reducing transpiration (Gedney et al., 2006). This effect, however, could be restricted according to this fertilization effect on leaf size which increases ET (Piao et al., 2007). Land use change effects might not be captured in global ET due to their regional nature and high inter-annual variability (Jung et al., 2009) or might show great uncertainty (Mao et al., 2015). Deforestation and irrigation expansion were found to have counteracting effects with a global difference of 400 km³ y⁻¹ (Gordon et al., 2005) considered signal noise when compared to the range of global ET assessments (Table S1). Nitrogen deposition can increase ET due to fertilization effects that increase leaf area index, but also decrease bare soil evaporation during the growing season (Mao et al., 2015). Finally, a drop in soil moisture stocks can reduce water vapour flows when ET is water limiting, particularly in arid and semi-arid regions, as well as highly seasonal environments also present on the South American continent. This effect requires further examination as it implies a change in the partitioning of local water resources.

Table S1. Estimates of global evapotranspiration (ET) over terrestrial land surfaces reported as volume (km³ y⁻¹) or depth (mm y⁻¹) through remote sensing and land-atmosphere models

Timeframe	Global ET		Method	Reference
		544 mm y ⁻¹	LSX model	Levis et al., 1996
	63,200 km ³ y ⁻¹		Area derived ET	Rockström and Gordon, 2001
1961–1990	63,906 km ³ y ⁻¹		LPJ model	Gerten et al., 2005
1961–1990	66,600 km ³ y ⁻¹		Potential ET and vegetation coefficient	Gordon et al., 2005
	65,500 km ³ y ⁻¹			Oki and Kanae, 2006
1971–2000	62,262 km ³ y ⁻¹		LPJmL model	Rost et al., 2008
1982–2008	65,000 km ³ y ⁻¹		FLUXNET upscaling	Jung et al., 2009
1986–1995	65,800 km ³ y ⁻¹		GSWP-2 model	Schlosser and Gao, 2010
1983–2006		539.3 ± 9 mm y ⁻¹	Empirical	Zhang et al., 2010
2003–2007	67,900 km ³ y ⁻¹	463 mm y ⁻¹	GLEAM	Miralles et al., 2011
2000–2006	62,800 km ³ y ⁻¹	416 ± 337 mm y ⁻¹	MOD16 ET	Mu et al., 2011
1989–1995		581 mm y ⁻¹	30 datasets	Mueller et al., 2011
1982–2009		558–650 mm y ⁻¹	Empirical	Zeng et al., 2012
		439–512 mm y ⁻¹	Various	Wang and Dickinson, 2012 ^a

^aAverage of 17 global ET results

Table S2. Changes in global and South American (SAm) evapotranspiration trends (ΔET) as evaluated by four different studies for the 1983–2006 period (Zeng et al., 2012).

Timeframe considered for global mean	Global mean ΔET (mm y⁻²)	Global ΔET (1983–2006) (mm y⁻²)	SAm ΔET (1983–2006) (mm y⁻²)	Reference
1982–2008	0.41 ± 0.20	0.48 ± 0.13	0.27 ^a ± 0.21	Jung et al., 2009
1983–2006	0.72 ± 0.21	0.72 ± 0.21	3.21 ± 1.00	Zhang et al., 2010
1982–2009	1.10 ± 0.20	1.12 ± 0.25	2.33 ± 0.71	Zeng et al., 2012
1982–2010	1.18 ^b			Mao et al., 2015

^aonly non-significant trend, all other results are significant (99% confidence); ^bmedian value provided

Table S3: Evapotranspiration of biomes, cropland (green and blue water) and pasture derived from remote sensing as described in the references

Humid tropical forest	Deciduous broadleaf forest	Savannah	Cropland	Grassland	Reference
<i>mm y⁻¹</i>					
1245	792	882–1267	1438	599	(Rockström et al., 1999)
800		280–1200			(Falkenmark and Rockström, 2004)
1310–1675	635 ± 200	676 ± 183	507 ± 157	311 ± 193	(Brujinzeel, 1990; Zhang et al., 2010)
1182	200–900	806	542	462	(Matyas and Sun, 2014; Miralles et al., 2011)

2 Precipitation in the Amazon region

Mean precipitation varies within the Amazon Basin (Figure 1 in the main document) between 1400 and 3000 mm y⁻¹ (ANA, 2011), occurring over a gradient of wet (North-western Amazonia, 1.25–1.75 mm d⁻¹) to dry climates (south-eastern Amazonia, 0.25–0.50 mm d⁻¹) (ANA, 2011; Davidson et al., 2012). Along with this gradient is a difference in wet and dry seasons: northern Amazonia's shorter wet season peaks between March and May compared to southern Amazonia (December and February). The dry season in Amazonia typically occurs between May and October and is more extended in southern Amazonia and into Mato Grosso (Restrepo-Coupe et al., 2013; Arvor et al., 2014). Rainfall in the basin results from two interconnected processes: two thirds of precipitation is sourced from influx of water vapour from evaporation in the Atlantic Ocean, while as high as 30% of precipitation is sourced from evaporation occurring within the Basin (Table 3 in the main document) with variations observed between wet and drought years (Bagley et al., 2014).

The input of water vapour from the Atlantic Ocean acts as an initial water vapour source into the region (Marengo, 2006). Possible impacts on this input have been proposed by Lewis et al. (2011) due to consequences from the rise in sea surface temperatures resulting from global climate change. Northern Amazon rainfall is correlated to El Niño Southern Oscillation (ENSO) events while southern Amazonia's rainfall appears to be more influenced by the south Atlantic multi-decadal oscillation (Marengo, 2004). During El Niño events (e.g. 1998, 2003, 2005, 2010), the northern region can experience drought, as opposed to the southern region which may see excess precipitation during ENSO-positive wet seasons (Marengo, 2004). To ENSO cycles is also superimposed a 28-year precipitation cycle which may intensify rainfall effects in the basin when in phase with El Niño or La Niña (e.g. 1970s wet years) (Davidson et al., 2012). Time series analysis shows transitions in last century's precipitation regimes through the mid-1940s and 1970s with more intense El Niño events occurring since 1976 which created drier conditions in the north and wetter conditions in the south (Marengo, 2004). Since then, tropical north Atlantic sea surface temperatures have increased gradually with spikes observed in 1985, 1998, 2005 and 2010 which have also coincided with drought years (Marengo et al., 2011). In addition to increases in sea surface temperatures since the "climate shift" of the 1970s, a recent increase in the length of the dry season, especially in the Southern region of the basin, has been observed particularly in the most recent drought of 2010 (Marengo et al., 2011).

The 2005 and 2010 droughts were caused by higher than usual warming of the tropical north Atlantic (with 2010 warmer than 2005) and the north-western displacement of the intertropical convergence zone, rather than ENSO (Marengo et al., 2008; Lewis et al., 2011; Marengo et al., 2011). These two droughts showed temperatures 3–5°C higher than usual (Davidson et al., 2012) and saw large areas affected (1.9 and 3.2 million km² in 2005 and 2010 respectively (Lewis et al., 2011)) with southern and western Amazonian regions experiencing a greater reduction in precipitation (Marengo et al., 2008). Drops in regional precipitation by as much as 25% from normal values for the December 2004 to February 2005 period (Marengo et al., 2008) had significant impacts on the river discharge, repeated in the 2010 dry season which was considered the worst on record (Marengo et al., 2011). Such droughts have also been described as connected to anthropogenic climate change (Lewis et al., 2011) with possible feedbacks on the water cycle through a reduction in precipitation recycling.

The 1970s climate shift identified from long term precipitation and reanalysis data (Marengo et al., 2008, 2011) is echoed by another phase in the Amazon's climate which includes possible feedbacks from land use change (Davidson et al., 2012). Increases in temperatures from global climate change can greatly affect tropical forests in the Amazon with consequences on ecological functions, especially in southern and south-eastern Amazonia which seem more susceptible to predicted declines in regional rainfall (Coe et al., 2013). Such feedbacks make the region vulnerable to increased occurrences of natural and human-made fires with smoke and aerosol releases possibly contributing further to ecological stress by reducing ET via diminishing incoming radiation, preventing droplet formation within the basin, and thus cutting short the precipitation recycling process (Davidson et al., 2012; Coe et al., 2013). Some models predict a so-called "savannization" process in Amazonia in which humid tropical forest could be replaced by savannah ecosystems partially as a result of reduced precipitation and with important consequences for the region (Nepstad et al., 2008). New openings and fragmentation in the forest landscape can also increase drying of the forest floor which creates ideal conditions for fires to spread with dry litterfall acting as fuel. Fires, in turn, create new conditions which favour invasion of grasses susceptible to further increase risks of fires while changing the landscape (Silvério et al., 2013). Deforestation and land use change, either through agricultural expansion, road building, human-made or natural fires, change the surface energy balance with consequences on the amount of water vapour returning to the atmosphere to recycle precipitation (Davidson et al., 2012; Bagley et al., 2014). Such recycling processes have become important in the dry season as the landscape transitions to a more significant source of water vapour to the atmosphere with greater precipitation recycling ratios in drought years and further feedbacks on dry season rainfall and drought conditions (Marengo et al., 2011; Bagley et al., 2014). During those same periods, the fraction of green to blue water resources drops by half from the wet to dry season, from 0.6 to 0.3 (Ellison et al., 2012).

20 **3 Green and blue water resources in the Amazon Basin**

Early basin wide estimates from 18 different studies in the Amazon region assessed partitioning of 2000–3664 mm y⁻¹ of precipitation into 992–1905 mm y⁻¹ of green water, and 400–1800 mm y⁻¹ of blue water (as runoff) (Marengo, 2006) with roughly 3255 mm y⁻¹ and 2350 mm y⁻¹ respectively of water input and output into the region (Costa and Foley, 1999). More recent biome specific reports from in-situ measurements from the Large-Scale Biosphere-Atmosphere Experiment in Amazonia network (LBA) (Keller et al., 2004) and modelling results are shown in Table 2 in the main document.

The almost 7 million km² Amazon Basin contains 7 large river basins (Figure 1 in the main document), namely Solimões, Negro, Amazonas, Madeira, Tapajós, Xingu and Tocantins (Coe et al., 2009; ANA, 2011). The Amazon River typically shows discharges of 130,000 m³ s⁻¹ (September–October) to 240,000 m³ s⁻¹ (May–June) and some tributaries with stages of 2.1 m and 2.7 m (Rio Negro at Manaus, September–October and June–July), 1.3 m and 2.0 m (Solimões River at Fonte Boa, September–October and May–June), and 0.3 m and 1.1 m (Rio Branco at Rio Branco, August–September and February–March) (Marengo et al., 2008). Other tributaries on the right bank of the river flow north from south and south-eastern Amazonia such as the Xingu (8548 m³ s⁻¹), Tapajós (12,434 m³ s⁻¹) or Madeira (33,602 m³ s⁻¹) (ANA, 2011). Many wetlands in the Amazon Basin are temporary blue water reservoirs in the wet season when river systems experience a food pulse that

inundate river floodplains and forests (Junk et al., 2014). Groundwater resources are vast in the Amazon Basin, mostly in the Óbidos, Purus and Amazonia sub-basins as well as in the headwaters of the Xingu Basin with capacity of 3.0 to >10.0 m³ cap⁻¹ month⁻¹ (ANA, 2010). The Guaraní aquifer system, which extends 1.8 million km² and is shared between Brazil (71%), Argentina (19%), Paraguay (6%) and Uruguay (4%), does not intersect the Amazon Basin directly; rather it extends
5 into the Cerrado biome (OAS, 2009). Surface water is also a key component of power generation in the region with a surge in dam projects given the regional projected hydroelectric capacity (ANEEL, 2002). Recent accounting listed 154 hydroelectric dams in the region with 21 currently under construction and over 200 projects registered in the Brazilian national inventory. Included on the list is the Xingu's Belo Monte project in Altamira, Pará, with an expected capacity of 11,000 MW in a close to 440 km² reservoir (ANEEL, 2015; Macedo and Castello, 2015).

10 Recent droughts had significant impacts on discharge of the Amazon River system. During the 2005 drought, the Rio Negro dropped 3–4 meters below normal stage levels (22.30 meters) in September and Rio Branco dropped to a level of three meters, greater than one standard deviation below the long term mean (Marengo et al., 2008). In the more pronounced 2010 drought, the Rio Negro dropped closer to four meters below the stage historical average (Marengo et al., 2011), also the lowest ever recorded water level at Manaus (Davidson et al., 2012). Pokhrel et al. (2014) tested how groundwater in Amazonia might help
15 terrestrial ecosystems build resilience to future climate change with temperatures possibly increasing by 8°C in the September to November period. Simulations showed large reductions in discharge of the Tapajós and Xingu Basins with the worse changes in discharge of 59% and 32% with overall ET in the Xingu dropping by 110 mm month⁻¹ when removing the groundwater buffer effect (Pokhrel et al., 2014). These results demonstrate the importance of this blue water stock in providing additional ecosystem resilience to climate change in Southern Amazonia.

20 Green water resources are essential for terrestrial ecosystems in the region. Recent reviews of long term eddy covariance (Restrepo-Coupe et al., 2013; Christoffersen et al., 2014; Rodrigues et al., 2014) and Bowen ratio measurements (Biudes et al., 2015) across the biome have been reported (Table 2 in the main document). Precipitation partitioning varies with ecosystem ET across the Northern to Southern Amazonia ecotone and Cerrado (Restrepo-Coupe et al., 2013; Christoffersen et al., 2014; Rodrigues et al., 2014; Biudes et al., 2015). All values reported in Table 2 (main document) represent the spatial and temporal
25 variability in precipitation partitioning across the biome with the fraction of ET representing 0.45–0.77 of precipitation from Southern (Sinop, Mato Grosso) to equatorial Amazonia (Santarem, Pará; Manaus, Amazonas). Differences in ET are due to the precipitation gradient as forest ecosystems exhibit similar green water consumptive uses in non-water-limited conditions (Rodrigues et al., 2014). The North to South ecotone follows the precipitation gradient from evergreen humid tropical forest (Manaus, Amazonas, 964 mm y⁻¹ and 1123 mm y⁻¹ at two sites) through to semi-deciduous and deciduous transition forest
30 (Sinop, Mato Grosso, 965 mm y⁻¹ and Alta Floresta, Mato Grosso, 1100 mm y⁻¹), Cerrado and finally the Pantanal wetland in Southern Mato Grosso (Rodrigues et al., 2014; Biudes et al., 2015). While these values are similar on an annual basis, there are important differences in ET seasonality in the region which also follows a north to south gradient with both biotic and abiotic controls that impact the amount of green water consumed in the region, and consequently the partitioning of precipitation. Modelling results for soybean in Mato Grosso show values of ET averaging 446 mm y⁻¹ for soybean,

276 mm y⁻¹ for maize and 856 mm y⁻¹ for pasture (crop year considered) (Lathuillière et al., 2012). A change in precipitation partitioning with land use was apparent in a simulation by Dias et al. (2015) performed with empirical data collected on a farm in south-eastern Amazonia: 1301 mm y⁻¹ of precipitation became 1025 mm y⁻¹ of green water and 323 mm y⁻¹ of blue water for tropical forest landscapes and 679 mm y⁻¹ and 636 mm y⁻¹ respectively for soybean (Cerrado was similar to tropical forest) (Dias et al., 2015).

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