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Of lakes and fields: A framework for reconciling palaeoclimatic drought inferences with archaeological impacts

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ABSTRACT

Quantitative estimates of climate variability are increasingly important in interpretations of archaeological turnovers in arid regions. Variations in lake levels or lake-water oxygen isotope ratios ($\delta^{18}O$) are often used to infer droughts or humid periods, along with speleothem $\delta^{18}O$, pollen, and windblown dust records. Key examples are the centennial-scale Holocene events associated with the end of the Bronze Age (~1200 BCE), the end of the Copper Age (~4000 BCE), and the onset of Neolithic expansion (~6200 BCE). Whether explicitly stated or only implied, causality between archaeological turnovers and inferred droughts is often ascribed to a disturbance to food resources, which means a disturbance to the agricultural potential of the study region. In the present study, a simple framework of equations is presented for evaluation of this causality. It quantitatively reveals significant complications. In one example, substantially improved crop-growing potential is found to coincide with dropping lake levels, which reflect significant net drought. The complications mainly arise from: (1) control of annually averaged climate conditions on lake changes *versus* control of seasonal conditions on the yield potential of fields; and (2) changes in the ratios between the overall catchment area of a lake or field, and the surface area of the lake or field itself. The results demonstrate that lake records *per se* do not satisfactorily reflect agricultural potential, but also that this gap may be bridged with targeted information collection about the regional setting. In particular, improved results may be obtained from detailed assessments of change in the catchment ratios of the lake(s) and field(s) that are being studied (e.g., using digital elevation models), along with expert opinions on field irrigation potential. The scenarios presented here then allow initial field-based assessments and hypothesis formulation to prompt more sophisticated modelling.

1. Introduction

There is a need for quantitative climate information in studies that assess potential climate impacts on early societies, especially in the lands around the eastern Mediterranean with their apparently close temporal coincidence between major archaeological transitions and sustained, centennial-scale climate events (e.g., Cullen et al., 2000; Weiss and Bradley, 2001; Brooks, 2006; Staubwasser and Weiss, 2006; Weninger et al., 2006, 2009; Clare et al., 2008; Berger and Guilaine, 2009; Clare and Weninger, 2010; Roberts et al., 2011; Langgut et al., 2013, 2014; Weiss, 2014; Clarke et al., 2016). In arid subtropical regions like the eastern Mediterranean borderlands, changes in the precipitation regime, and especially droughts, have long been the prime suspects for influencing archaeological/societal turnovers and/or migrations (e.g., Cullen et al., 2000; Weiss and Bradley, 2001; di Lernia, 2002; Hassan, 2002; Vernet, 2002; Brooks, 2006, 2012; Staubwasser and Weiss, 2006; Berger and Guilaine, 2009; Langgut et al., 2013, 2014; Clarke et al., 2016). Such studies regularly mention trends in food storage, dietary changes, development of irrigation systems, and competition for resources. Whether stated directly, or by implication only, the inferred link between drought and societal turnover concerns some aspect of perturbation to the region's agricultural potential (including pasture land), although intense debate remains around issues of: (a) temporal coincidence *versus* causation; (b) the temporal coincidence itself; (c) coeval impacts of perturbations other than drought; and (d) resilience or recovery potential with respect to drought (e.g., Hassan, 2002; Brooks, 2006, 2012; Wilkinson et al., 2007; Clare et al., 2008; Berger and Guilaine,

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2009; Weninger et al., 2009; Clare and Weninger, 2010; Maher et al., 2011; Schmidt et al., 2011; Riehl et al., 2012; Clarke et al., 2016; and references therein). In spite of such overprints, a close relationship appears to exist between politico-societal turmoil in the region and climate events – notably aridity – both in prehistoric times (Cline, 2014), and even today (Kelley et al., 2015).

Lake-level and lake-water $\delta^{18}O$ records, along with pollen and cave-speleothem-$\delta^{18}O$ records and wind-blown dust records, constitute the key information to palaeoclimatologists who try to ascertain whether sustained droughts occurred in the past. For examples, see Gasse and Van Campo (1994), Harrison et al. (1996), Landmann et al. (1996), Cheddadi et al. (1997), Naruse et al. (1997), Roberts et al. (1999, 2011), Digerfeldt et al. (2000), Gasse (2000), Hoelzmann et al. (2001), Weiss and Bradley (2001), Enzel et al. (2003), Migowski et al. (2006), Stevens et al. (2006), Rambeau (2010), Kuzucuoğlu et al. (2011), Finné et al. (2011), Frumkin et al. (2011), Rohling (2013), Clarke et al. (2016), and references therein. For applications that focus on potential archaeological implications, see – among many others – Weiss and Bradley (2001), Hoelzmann et al. (2001), Staubwasser and Weiss (2006), Clare et al. (2008), Berger and Guilaine (2009), Weninger et al. (2009), Langgut et al. (2013, 2014), Ackermann et al. (2014), Clarke et al. (2016), and the many references therein.

Despite the potential relationship between aridity inferred from palaeoclimate archives and production potential of either wild or managed agricultural land (including pasture land), there has been little attention to the question of how relevant information from palaeoclimate records is to an assessment of food-yield potential. In other words: how well – in at least a semi-quantitative sense – does a trend to 'aridity' as recorded in palaeoclimate archives translate to a limitation in the potential for foraging, crop-growing, or life-stock pasturing? There is an increasing realisation that such more quantitative interpretation methods are needed when assessing past water availability (Jones, 2013). The present study presents a simple generalised quantitative interpretative framework for arid regions, and applies this to the main example region of the Levant. First, however, the framework is validated against historical observations in the region of Canberra, Australia, which demonstrates the ubiquitous applicability of the scenarios to arid regions on a global scale. Application of this framework, potentially with region-specific adaptations, may offer basic quantitative assistance to future climate-archaeological studies, for initial field-based assessments and/or hypothesis formulation.

2. Methods and results

The scenarios developed rely on a couple of general principles. First, palaeoclimate archives reflect the net annual balance of precipitation + evaporation ($P+E$, where E is a negative term), in a manner that needs to include assessment of conditions in the catchment area, straight evaporation from open water, and evapotranspiration from vegetated soils. Second, lakes change in size as their levels rise or drop, which changes the surface area from which evaporative loss occurs, whereas the catchment area from which net precipitation is channelled into the lake depends on regional physiography, and is unlikely to change much over centuries to millennia. Third, the surface area of fields does not depend on $P+E$, and their catchment area may be (artificially) enhanced by irrigation.

As a result of the variety of processes and influences involved, substantial complexity may be expected in the relationship between the (multi-) annually averaged information contained in palaeoclimate archives, and the shorter-term controls that determine whether an agricultural field may have been productive during a regional crop-growing season. These complexities are commonly beyond the sampling resolution that can be achieved in palaeoclimate archives, but that does not need to limit the assessment. It has been shown that combined changes in lake-level and speleothem $\delta^{18}O$ data from closely spaced sites, but with different temporal resolutions, allow quantification of the magnitude of past climate fluctuations in considerable detail (e.g., Medina-Elizalde and Rohling, 2012).

The present study specifically investigates the impacts on lake levels (and to some extent lake $\delta^{18}O$) and field hydrology, due to: (*i*) control by $P+E$; (*ii*) seasonal variations of P and E; and (*iii*) natural and/or man-made changes in the ratio of the surface area of the total lake or field catchment relative to the surface area of the lake or field itself. To illustrate why this is important, imagine a situation in which sufficient P increase (or irrigation) occurs to achieve positive $P+E$ over a field in the crop-growing season, while regional evaporative loss (E) in the non-crop-growing season is so high that annually averaged $P+E$ becomes negative. Under such conditions, the situation is good for growing crops, even though lake levels in the region will be dropping over time. Hence, if lake level changes were used in this case to qualitatively assess the region's crop-growing potential, one would (correctly) infer a net drought that would be (incorrectly) considered to cause an agricultural crisis. A more quantitative perspective is required, and this study presents a basic and easily applicable framework for this.

The problem is reduced to the simplest relationships needed to achieve a more quantitative approximation of the relationships between climate conditions, lake levels (and $\delta^{18}O$), and potential field productivity for arid settings in which the groundwater table remains below the shallow root systems of crops (i.e., crops are fed only by rain and/or shallow soil moisture). This is not an attempt to model reality, but to present an idealised conceptual framework with hypothetical scenarios, aimed at evaluating: (1) whether lake levels (and $\delta^{18}O$) provide information about agricultural potential in sufficiently straightforward terms; and (2) what additional information would be most useful to achieve a better representation. I use the calculations mainly to assess scenarios in the climatic settings of the Levant (Jerusalem, Damascus), but also validate the solutions by investigating arid subtropical conditions in a completely different part of the world, namely southeastern Australia (Canberra).

All scenarios assume that the year is perfectly divided into two halves, a summer half and a winter half, as far as evaporation is concerned. This is actually quite representative of reality in Jerusalem today, with summer spanning April–September with daily evaporation rates of about −5 mm (equivalent to −1825 mm/y), and winter spanning October–March with daily evaporation rates of about −2 mm (equivalent to −730 mm/y) (IMS, 2015a). Seasonal distribution and calculated potential evaporation values for climatic values on Climatemps (2015) and Weatherspark (2015) are similar for Damascus. For the Canberra scenarios, annual mean evaporation rates are −1677 mm/yr, with summer rates some 4 times higher than winter rates (equivalent to about −2680 mm/yr in the summer half year, and −670 mm/yr in the winter half year) (BOM, 2016).

As per data compilations presented on IMS (2015a), Climatemps (2015), and Weatherspark (2015), nearly all P is set to occur in the winter half year in the Levantine scenarios. There, the typical fraction of the winter-half year over which rain actually falls (α) is about 0.25 (i.e., 44 days in the Jerusalem region; IMS, 2015b). If precipitation becomes more focussed (extreme) in a shorter time, then α reduces. If precipitation gets more evenly spread out, then α increases. Note that for different systems in the same region, α will be approximately the same for all systems, as it represents a characteristic of the regional climate conditions. Given that the data compilations indicate no significant rainfall in the summer half year, the Levantine scenarios use very low β values close to 0. In the Canberra scenarios, P is evenly distributed through the year, with α and β both at around 0.3 (108 days per year; BOM, 2016).

Evapotranspiration from a field with grasses or crops of wheat is related to potential ('pan') evaporation for the selected growth season, using relationships outlined in Allen et al. (1998). That study relates specific crop evapotranspiration to reference (grass) evapotranspiration via coefficient K_c, so that $ET_c = K_c\ ET_0$. The coefficient is typically close to 1.05 for short crops, and increases to mean values of 1.2–1.25 for tall crops (see Figs. 20 and 21 of Allen et al., 1998). From this, it would seem that a reasonable approximation for K_c in the case of intermediate-size plants such as wheat would be 1.1. However, soil conditions come into play as well. When the soil surface is dry, K_c will be small and may drop to as low as 0.1, but it still approaches 1 for fully grown crops on dry soils (Fig. 22 of Allen et al., 1998). Taking this into account, I use $K_c = 1$ in this arid-region study, which means that crop evapotranspiration is equal to reference evapotranspiration. If anything, the soil-humidity dependence (Fig. 22 of Allen et al., 1998) and crop-stage dependence (Fig. 24 of Allen et al., 1998) suggest that $K_c = 1$ may be an overestimate when integrated over a crop lifecycle, and that it is more likely to be <1. Working with a value of 1 therefore assumes a worst-case scenario of water loss.

Given that the evaporation rates used in this study effectively are pan-evaporation rates, a conversion is needed from these to reference evapotranspiration rates in the same settings (ET_0). Allen et al. (1998) state that $ET_0 = K_p E_{pan}$, where E_{pan} is the pan-evaporation rate, and K_p the conversion constant that is needed here. Values for this constant are found to be between 0.6 and 0.7 over a variety of scenarios in Table 6 of Allen et al. (1998), for a sunken pan (most similar to a lake) that is far removed from cropland, and for typical mean wind speeds of 3–7 m s^{-1}. I use a mean value of $K_p = 0.65$. Finally, then, evapotranspiration from the crop fields for the growth season is given by $ET_c = K_p K_c E_{pan}$, or $ET_c = 0.65 E_{pan}$.

The calculations allow evapotranspiration from soil as soon as there is moisture, and ramp evapotranspiration down to zero if there is no soil moisture. From the saturated surface of a lake, however, there is always evaporative loss. The scenarios calculate steady state solutions, and do not consider recharge of, or release from, aquifers (that is, I consider the most basic of configurations that are simply set to be sealed with respect to aquifer interaction – such processes can easily be added in a case-specific manner if 'real' lakes were investigated). All calculations are set up with non-dimensionalised areas, which is done through normalisation relative to the lake or field area (i.e., the lake or field area is set to 1, and catchment areas are expressed as multiples of that unit).

To account for the fact that the lake catchment area will consist of similarly vegetated soil as the field, evapotranspiration needs to be considered in the catchment area (as for the field), while total evaporation affects the lake surface itself. Here, evapotranspiration uses $K = K_p K_c = 0.65$, as discussed above, given that even rainfall soaked into the soil remains accessible for wheat due to root systems that can grow down to 1.5–2.0 m (FAO Water, 2013), i.e., enough to penetrate the entire typical <1 m depth of soil in the Levant (Henkin et al., 1998).

With reference to the terms in Table 1, conditions for a lake may then be summarised as follows. Net water budget in the dry part of the summer half year: $X_{s.d} = 0.5E_s(1-\beta)$. Net water budget in the wet part of the summer half year: $X_{w.w} = (P_s + 0.5E_s\beta K)(\Phi - 1) + (P_s + 0.5E_s\beta)$. Net water budget in the dry part of the winter half year: $X_{w.d} = 0.5E_w(1-\alpha)$. Net water budget in the wet part of the winter half year: $X_{w.w} = (P_w + 0.5E_w\alpha K)(\Phi - 1) + (P_w + 0.5E_w\alpha)$. Solving these equations to give P_0, the critical mean annual ($P_s + P_w$) value for which lake levels remain unchanged, uses the steady state solution $X_{s.d} + X_{s.w} + X_{w.d} + X_{w.w} = 0$. This gives:

$$P_0 = \left\{0.5\frac{E_w(1-\alpha K) + E_s(1-\beta K)}{-\Phi} - K(\alpha E_w + \beta E_s)\right\} \tag{1}$$

For a field, the relevant relationships are as follows. Net water budget in the dry part of the summer half year: $X_{s.d} = 0.5E_s(1-\beta)K$. Net water budget in the wet part of the summer half year: $X_{w.w} = (P_s + 0.5E_s\beta K)\Phi$. Net water budget in the dry part of the winter half year: $X_{w.d} = 0.5E_w K(1-\alpha)$. Net water budget in the wet part of the winter half year: $X_{w.w} = (P_w + 0.5E_w\alpha K)\Phi$. This can be used to solve the annual amount P_0 and the separate winter amount P_{0w} for a field, according to $X_{s.d} + X_{s.w} + X_{w.d} + X_{w.w} = 0$, and $X_{w.d} + X_{w.w} = 0$.

It is relevant to separately assess the critical values of precipitation for the winter half year, and the annual total, because winter is the key growth season in the arid regions considered. In Israel today, plowing typically begins in September/October, planting in October to December, and harvesting starts in February (flax), March (barley), and May (wheat). Despite the more even distribution of rain through the year in the Canberra region, winter remains the key season for grain crops, as it is throughout southeastern Australia because of extreme summer evaporation (CelciusPro, 2010). Note that the Canberra scenarios are presented only to validate the methods, since evaluating constraints on past wheat production in the Canberra region has very limited value by itself, given that wheat was only introduced to Australia in 1788. We find:

$$P_{0w} = 0.5K\left\{\frac{E_w(1-\alpha)}{-\Phi} - \alpha E_w\right\} \tag{2a}$$

$$P_0 = 0.5K\left\{\frac{E_w(1-\alpha) + E_s(1-\beta)}{-\Phi} - (\beta E_s + \alpha E_w)\right\} \tag{2b}$$

The annual mean $\delta^{18}O$ mass balance for a lake can be evaluated in similar ways as the argument for Equation (1), and then reduces to a steady state (no $\delta^{18}O$ change) solution of:

$$P_0 = 0.5\left\{\frac{E_w(1-\alpha K) + E_s(1-\beta K)}{-\Phi}\frac{\delta_{ev\_am}}{\delta_{p\_am}} - K(\alpha E_w + \beta E_s)\frac{\delta_{e\_am}}{\delta_{p\_am}}\right\} \tag{3}$$

Here, δ_{p_am} stands for the annual weighted mean rainfall $\delta^{18}O$. Over the year, rainfall δ_p in the Levant typically swings between about −2 and −6‰, with occasional values of 0 to −10‰ (Gat, 1996; Gat et al., 2005; El-Asrag, 2005). The annual weighted mean value (δ_{p_am}) is about −5‰ around the Levantine coast (±1‰) (Gat, 1996; Gat et al., 2005), and also close to −5‰ (±1‰) for Canberra (IAEA, 2016). In both cases, δ_{e_am} relates to the annual weighted mean surface-water $\delta^{18}O$ via a −10‰ equilibrium fractionation relative to the evaporating water bodies; potentially stronger offsets due to kinetic fractionation are not considered (cf., Rohling, 1999; Rohling et al., 2004). With respect to the annual weighted mean $\delta^{18}O$ of evapotranspiration (δ_{ev_am}), plants typically bring groundwater up to the leaves with little fractionation (commonly remaining close to δ_{p_am}), and then experience strong evaporation fractionation in the leaves. As the plant approaches steady state, $\delta_{ev\_am} \approx \delta^{18}O$ of groundwater (so commonly close to δ_{p_am}) (e.g., Gat, 1996; Griffis et al., 2010).

For a closed lake to achieve mass balance (Equation (1)) as well as isotopic mass balance, a unique solution is required, in which $\delta_{e\_am} \approx \delta_{ev\_am} \approx \delta_{p\_am}$ (Equation (3)). Given the weighted mean annual δ_{p_am} of −5‰ in the Levant and Canberra regions, this

Table 1
Parameters used in the scenarios.

Parameter	Description
E_s	Summer evaporation (a negative value, in mm/y)
E_w	Winter evaporation (a negative value, in mm/y)
E_{pan}	Pan-evaporation rate (a negative value, in mm/y)
K_c	Coefficient to convert reference evapotranspiration to specific crop evapotranspiration
K_p	Coefficient to convert (pan) evaporation to evapotranspiration
ET_0	Reference evapotranspiration rate (a negative value, in mm/y)
ET_c	Specific crop evapotranspiration rate (a negative value, in mm/y)
P	Annual precipitation (in mm/y)
α	Fraction of winter during which it rains ($0 \leq \alpha \leq 1$). Value of 0 means dry winter
β	Fraction of summer during which it rains ($0 \leq \beta \leq 1$). Value of 0 means dry summer
Φ	Area ratio $A_{catchment}/A_{lake\ or\ field}$
P_0	Critical P value to keep water level at steady state (in mm/y)
X	Net water gain or loss for lake or field/groundwater (in mm/y)
δ_p	$\delta^{18}O$ of precipitation (in ‰)
δ_e	$\delta^{18}O$ of evaporation (in ‰)
δ_{ev}	$\delta^{18}O$ of evapotranspiration (in ‰)
s	Subscript indicating summer
w	Subscript indicating winter
$\_am$	Subscript addition indicating annual weighted mean
$w.d$	Subscript indicating dry portion of winter; i.e., portion ($1-\alpha$)
$w.w$	Subscript indicating wet portion of winter; i.e., portion (α)
$s.d$	Subscript indicating dry portion of summer; i.e., portion ($1-\beta$)
$s.w$	Subscript indicating wet portion of summer; i.e., portion (β)
lake	Subscript to identify use of a parameter specifically for a lake
field	Subscript to identify use of a parameter specifically for a field

implies that the $\delta^{18}O$ of lake surface water would need to approach $-5+10 = +5$‰ (this agrees with more detailed assessments and observed values around the eastern Mediterranean; Jones and Imbers, 2010). In all other cases, either more positive or more negative, mass balance and isotopic mass balance for a lake are achieved at different E and P values. However, under constant forcing, a closed lake system will eventually approximate the steady state solution. This is because any change in δ_{p_am}, which is not necessarily related to a change in P (e.g., a change to more negative values due to a shift from annual rains to winter-dominated rains) will result in rapid adjustment of δ_{ev_am}. The water flowing into the lake will be more negative as well, so that lake-water $\delta^{18}O$ will shift to a more negative value; slowly in a deep lake, and more rapidly in a shallow lake. In turn, this results in a more negative δ_{e_am}, until eventually a new steady state is achieved.

Fast (quasi-)steady state isotopic mass balance adjustments that do not seriously lag behind mass balance adjustments may be expected only in shallow closed lakes with short residence times. Isotopic mass balance of deep lakes adjusts over longer timescales, and thus will lag behind mass balance adjustments. This is because, relative to the residence time (τ), any perturbation will cause property (e.g., isotope) changes in the reservoir so that $C(t) = C_{init}\, e^{-(t/\tau)}$ where C_{init} is the initial property value, and $C(t)$ the property value at time t. Hence, (nearly) complete adjustment of the reservoir's property value will take about 3 times the residence time, although most will be completed within about two times the residence time (see also Jones and Imbers, 2010). In contrast, any mass-balance change will be visible immediately in rising or falling lake levels. Fully resolving the isotopic solutions requires a series of assumptions, as well as time-dependent solution of Equation (3), which goes beyond the scope of the present study and is best done in a comprehensive, system-specific manner. If anything, however, note that lake-isotope Equation (3) is sufficiently similar in structure to lake mass-balance Equation (1) that any P_0 value determined with it will be much closer to that from Equation (1) than from winter-specific field Equation (2a). When considered over timescales of 2 times the studied system's residence time or longer, the solutions from Equations (3) and (1) will be nearly identical.

For conditions representative of Jerusalem/Damascus in the Levant, Equations (1), (2a), and (2b) give the results shown in Fig. 1. This shows that, in a given climate regime (i.e., using the same α and β for the three plots), a field with a given catchment area ratio (Φ) requires less precipitation to remain at steady state than a lake with the same catchment area ratio, and this difference is especially pronounced when focussing on the precipitation amount required

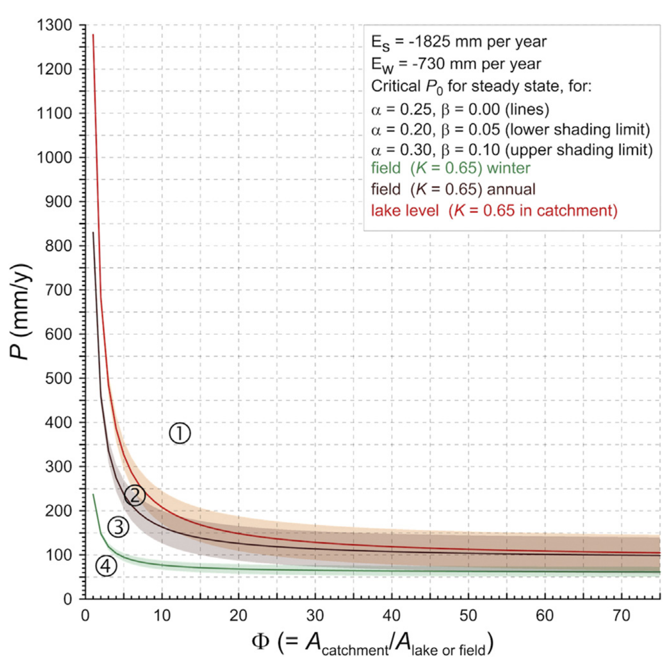

Fig. 1. Levantine scenarios. Critical P_0 values needed to maintain steady state in a field (green for winter conditions; brown for annual mean conditions), and lake level (red), as a function of catchment ratio (Φ). These scenarios are developed using input values and ranges that are reasonable for places like Jerusalem and Damascus in the Levant. The circled numbers indicate distinct sectors in the diagram, as discussed in the text. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

to maintain steady state in the field during the winter growth season. Also, the sensitivities are very different, which demonstrates that lake level offers a distinctly non-linear measure of the conditions that would affect farming. Similar observations, if more pronounced, apply to the Canberra scenarios (Fig. 2).

In these evaluations, the catchment area ratio (Φ) is a function of: (*i*) the area of the catchment, which for lakes (until recent centuries with pipelines, tunnels, and aquaducts) was always limited by physiography and would not likely change much over the centennial to millennial timescales considered; and (*ii*) the area of the lake itself, which may be quite variable, especially for lakes in regions of flat topography. When $P > P_0$, the lake would rise, and accordingly grow in size. A doubling of lake size would half the value of Φ. At that lower Φ, the lake would need more P to remain at steady state (Figs. 1 and 2). This negative feedback keeps a check on lake growth as long as P remains close to critical values, and until the lake size would reach the appropriate steady-state size for the available P (if P became excessive, then the lake would eventually overflow and drain, and the extra loss term of drainage would then help to establish a new steady state).

For a field, what matters is the catchment area for groundwater that resides shallow enough to be available to shallow crop roots. Typically, fields in arid regions have small catchments relative to the field area, because the resident groundwater does not reach shallow enough for the shallow-penetrating roots. This is why people develop irrigation, which effectively is an artificial increase in the field catchment. So fields in arid regions typically have very low Φ, and development of irrigation is a means to increase the Φ value of the field. A field at steady state means that it can be productive, as it has sufficient soil moisture to sustain the required evapotranspiration in the growth season (this assumes that there is no 'leakage' into long-term groundwater, which would increase P_0 from the values calculated here).

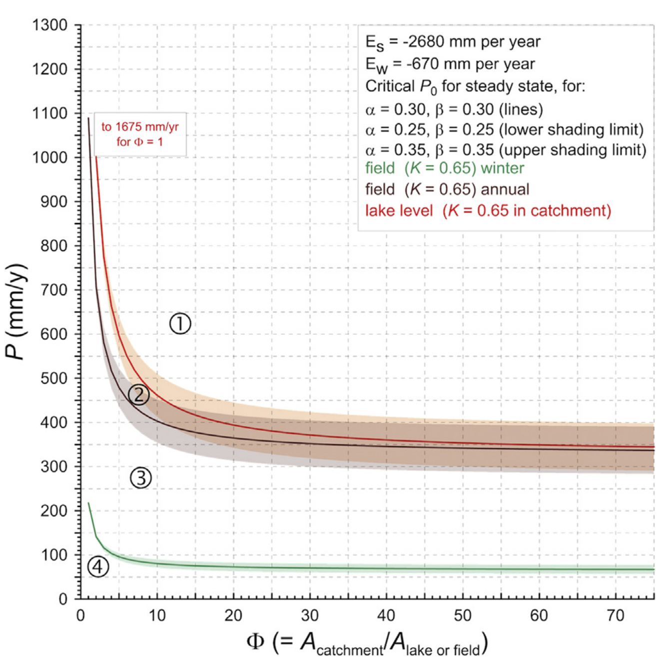

Fig. 2. Canberra scenarios. Critical P_0 values needed to maintain steady state in a field (green for winter conditions; brown for annual mean conditions), and lake level (red), as a function of catchment ratio (Φ). These scenarios are developed using input values and ranges that are reasonable for the Canberra region today. The circled numbers indicate distinct sectors in the diagram, as discussed in the text. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

None of the idealised systems considered here allows exchange with aquifers. For assessments where such processes are a factor, a case-specific scenario would need to be made. Regardless, the idealised 'sealed system' approach used here suffices already to illustrate the profound and fundamental complexity that is involved in comparisons between lake-level and field moisture conditions.

Within the calculated scenarios (Figs. 1 and 2), four distinct sectors can be recognised (see numbers in the figures). In sector 1, fields receive abundant net moisture and thus have the potential to be productive throughout the year. Lakes also receive a net moisture influx, so lake levels are rising, at least until drainage out of the lake develops. In sector 2, fields receive enough net moisture to be productive throughout the year, while lakes are insufficiently refilled so that lake levels are gradually dropping. In sector 3, fields still receive sufficient net moisture to be productive in winter, but not throughout the year. Lakes experience a considerable net loss of water and lake levels will be dropping relatively fast. In sector 4, fields are parched throughout the year, and lakes are dropping very fast.

In short, the wide range of values covered by sectors 2 and 3 represent environmental conditions during which lake levels would drop, while (winter) cropping remains unaffected. For development of the kind of regional aridity that may cause societal crises due to persistent crop failures, regional environmental parameters would have to slip into the very harsh conditions of sector 4 in Figs. 1 and 2.

3. Validation

The Dead Sea, a well-known arid lake/sea in the Levant, has a Φ of about 40, and more sophisticated calculation for the Dead Sea suggests that P_0 = ~110 mm/y (Rohling, 2013). Given the great simplicity of the scenarios presented here, the range of the red curves in Fig. 1 at $\Phi = 40$ is in decent agreement with that more accurate P_0 value. Note that freshwater evaporation rates (or pan-evaporation rates) in the Dead Sea region are twice the rate used here, but that evaporation from the Dead Sea itself occurs at only half the pan-evaporation rates due to the sea's extreme salinity (Rohling, 2013). This makes evaporation rate values for Dead Sea waters similar to the values considered here.

The values obtained from the simple calculation scheme presented here can be further validated by comparison with detailed modern observations available for the Canberra region. First, the field calculations are assessed against agricultural data. Next, the lake calculations are compared with data for Lake George, directly northeast of Canberra.

In the Canberra region, average rainfall in the two mid-winter months of August and September is 55 mm, and minimum rainfall over that period is 14 mm; below 40 mm, drought assistance is arranged for farmers (CelciusPro, 2010). The minimum equates to roughly 40 mm of total winter rainfall, the average to about 165 mm, and the drought threshold to about 120 mm. The natural (pre-irrigation) minimum rainfall line for cropping is known as the Goyder line (Australian Government, 2016), and corresponds to 300 mm of annual rainfall; just under half of that would fall in winter, which supports the 120 mm drought threshold mentioned before. The P_{0w} value calculated here for a non-irrigated field, with no extended catchment so $\Phi = 1$, is about 220 mm. The observed average condition of 165 mm winter precipitation provides enough moisture for fields with $\Phi \geq 1.5$; i.e., fields with some irrigation. At the drought threshold of 120 mm winter precipitation, the calculated P_{0w} lines suggest a necessity for heavy irrigation to the level of $\Phi = 3$. Overall, the calculated scenarios for fields reasonably approximate the observed conditions. If anything, the evapotranspiration factor (K) may have been slightly overestimated for the field scenarios, so that the calculations require a fraction more rainfall than observed drought limits suggest.

Lake George is a closed lake that is E−P dominated, as it has no other significant recharge mechanisms (Fenby and Gergis, 2012). It has a small catchment, with $\Phi = 6.25$ today (NSW Department of Primary Industries, 2016), and observed long-term average P is between 50 and 55 mm per month, or about 630 mm per year. The lake is located in a flat plain (ancient lake bed), and any refill or reduction is associated with large changes in Φ. Today the lake is small and roughly at steady state. The Canberra scenarios in Fig. 2 estimate mean annual P_0 for Lake George at 550 ± 50 mm, which (within assumptions) agrees well with the modern observed conditions. The small offset may be due to a considerable capacity for water-uptake by soil in the basin before lake levels are affected − this process is not included in the P_0 calculations.

The approximate agreements between observed conditions and the scenarios calculated here indicate that the calculations provide a sufficient representation of reality.

4. Discussion

The simple scenarios of this study allow several intriguing observations. First among these is that a lake needs more P to achieve steady state than a field in the same environment, considering both systems at the same Φ. However, lakes typically have large Φ (for an example in addition to the two lakes discussed, the West Nubian Palaeolake in the eastern Sahara had Φ values between 14 and 200; Hoelzmann et al., 2001). Fields in arid regions typically have very small Φ. It is therefore illustrative to look across the graph at an equal P_0 value.

In Fig. 1, a lake with $\Phi_{lake} = 40$ at $\alpha = 0.25$ needs $P_0 = {\sim}120$ mm/y to maintain a steady level. At the same critical P_0 value, productive winter fields can exist as long as $\Phi_{field} \geq 3$. If people were to achieve the marginal balance possible in a field with $\Phi_{field} = 3$, then there would be both steady lake levels as well as productive farming. A slight reduction in P (or a slightly more even spread of rain through winter, making α go up) would cause the lake to start dropping toward a new steady state with a smaller lake area (hence a higher Φ_{lake}). To achieve a certain area reduction, the level of a lake in a shallow and flat topography will have to drop only little, but in a steep-sided configuration the lake level would need to drop greatly. Hence, lake levels by themselves are highly non-linear measures of the amount of net drought; better estimates would require assessment from palaeo-data of changes in the lake's shape and catchment ratio (Φ_{lake}) that resulted from reconstructed changes in its level. For the field, the pertinent question is whether the small P reduction (or the more even distribution of P through winter) would mean that farming comes to an end. The answer is yes if irrigation was maximally developed already (Φ_{field} cannot be increased beyond 3), or no if some irrigation potential remained (to increase Φ_{field}).

Second among the observations from Fig. 1 is that continuous improvement of irrigation can sustain farming only so well; the critical P value has a distinct asymptote. Once irrigation reaches an effective value of $\Phi_{field} = {\sim}15$, which is a massive irrigation value already, the resilience to P changes does not appreciably improve even for titanic irrigation efforts. Moreover, any lakes that existed in the area would by then have experienced rapid net drawdown, as their P_0 values were no longer met at all, and this would adversely affect any irrigation prospects.

The third observation comes from another look at the systems for equal Φ values (for example, ~20), and considering settings where P is close to the respective values of P_0. In that case, a field could still be productive if winter $P > {\sim}70$ mm/y, but lake levels would be extremely low, and dropping, as this P value is well below the lake P_0 of ~150 mm/y. Even for a sizeable increase in P up to almost 150 mm/y (a doubling that would be great news for farmers in such marginal settings), lake levels still would be dropping or at best stabilise at a very low position. Very low and/or fast dropping lake levels would traditionally be interpreted as indicators of severe drought, yet farmers at the time might celebrate a doubling in rainfall, which would greatly increase their marginal fields' potential for production.

Finally, I emphasise a caveat to all mention above of field productivity. All of the argument is based uniquely on the moisture balance. Physical and heat damage to soils, and nutrient-related issues, are not included in these considerations and would need to be separately evaluated.

5. Conclusions

Despite their basic formulation, the scenarios presented here clearly illustrate that changes in lake level (and $\delta^{18}O$) exhibit very different sensitivities to precipitation changes than farming fields, in arid regions. Threshold values are different, and the relationship between changes in lake level (and $\delta^{18}O$) and field moisture is distinctly non-linear. It is demonstrated that conditions are possible in marginal settings for which lake levels may have been dropping, while fields were getting wetter and more productive. At face value, lake-level reconstructions therefore offer poor measures of agricultural potential, but there is a way forward. Lake-level (and $\delta^{18}O$) information may be much more effectively utilised in quantitative climate interpretations of a region's agricultural potential, if combined with estimates of: (1) the catchment ratios of the lakes and fields under investigation; (2) changes in the surface areas (hence catchment ratio) of the lakes due to lake-level variations; and (3) changes in the catchment ratios of the fields through irrigation. Much of this information may be obtained from comprehensive use of digital elevation models for the study regions, along with the application of expert opinions on irrigation potential (including transparent uncertainty estimates). In addition, hard-to-constrain changes in the seasonality of the hydrological fluxes and their isotopic composition can play an important role. Finally, case-specific assessments of exchange with deeper aquifers may also be added. Thus, the scenarios presented here may guide initial field-based assessments and formulation of hypotheses for testing with more sophisticated modelling.

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