Climate drivers for peatland palaeoclimate records

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Abstract

Reconstruction of hydroclimate variability is an important part of understanding natural climate change on decadal to millennial timescales. Peatland records reconstruct 'bog surface wetness' (BSW) changes, but it is unclear whether it is a relative dominance of precipitation or temperature that has driven these variations over Holocene timescales. Previously, correlations with instrumental climate data implied that precipitation is the dominant control. However, a recent chironomid-inferred July temperature record suggested temperature changes were synchronous with BSW over the mid-late Holocene. This paper provides new analyses of these data to test competing hypotheses of climate controls on bog surface wetness and discusses some of the distal drivers of large-scale spatial patterns of BSW change. Using statistically-based estimates of uncertainty in chronologies and proxy records, we show a correlation between Holocene summer temperature and BSW is plausible, but that chronologies are insufficiently precise to demonstrate this conclusively. Simulated summer moisture deficit changes for the last 6000 years forced by temperature alone are relatively small compared with observations over the 20th century. Instrumental records show the summer moisture deficit provides the best explanatory variable for measured water table changes and is more strongly correlated with precipitation than with temperature in both Estonia and the UK. We conclude that BSW is driven primarily by precipitation, reinforced by temperature, which is negatively correlated with precipitation and therefore usually forces summer moisture deficit in the same direction. In western Europe, BSW records are likely to be forced by changes in the strength and location of westerlies, linked to large-scale North Atlantic ocean and atmospheric circulation.

1. Introduction

Ombrotrophic peatlands developed during the Holocene in mid-high latitudes, achieving their greatest extent in regions with oceanic climates. Their moisture balance is solely dependent on precipitation and evapotranspiration as their surface profile is slightly convex or flat, such that they receive no surface runoff. Reconstructions of past bog surface wetness were used by Barber (1981) to refute the autogenic bog growth ideas formulated in the early 20th century and to show qualitative correlations with Lamb's (1977) High Summer Wetness and Winter Severity indices, correlations which were improved by the development of a Climate Response Model (Barber et al., 1994). A range of methods have been developed to reconstruct surface wetness changes over time in order to produce records of hydroclimate variability,including peat humification (Aaby 1976, Blackford and Chambers 1991), plant macrofossils (Barber 1981, Barber et al. 1994) and testate amoebae (Warner and Charman 1994). These techniques have been applied to a large number of sites from a range of geographical locations in Europe, the Americas and New Zealand (e.g. Wilmshurst et al. 2002, Barber et al. 2004, Booth et al. 2006, Charman et al. 2006, Chambers et al. 2007, Sillasoo et al. 2007, Swindles et al. 2007).

All reconstructions reflect the rather vaguely defined parameter of 'bog surface wetness' (BSW). Attempts to integrate BSW records with other proxy records and to provide data that are more directly useful in data-model comparisons, are hampered by poor understanding of the relationship between 'surface wetness' and simple climate parameters such as seasonal temperature and precipitation. Conceptually, it is known that during periods of positive precipitation-evapotranspiration the water table will rise to a maximum point beyond which excess moisture will be lost as runoff. Therefore the winter period (moist and cold in mid-high latitudes) is unimportant to long term surface wetness changes, except perhaps for snowmelt input in the spring which may extend the saturated season. It is the warm season, when evapotranspiration exceeds precipitation, that determines the length and severity of water table drawdown. There is thus broad agreement that it is the (extended) summer period that is reflected in the surface wetness proxies, but there are different views on whether temperature or precipitation is the primary variable driving the observed decadal-millennial scale BSW records. A recent debate concerning this issue has evolved based on evidence from two principal sources:

1. High resolution surface wetness reconstructions compared with instrumental meteorological records (Charman et al. 2004, Charman 2007) where correlations between reconstructions and meteorological records suggest strongest relationships of surface wetness changes with summer precipitation and, to a lesser extent with temperature. However, there is some suggestion that this relationship may vary spatially, with one record showing the strongest correlation with mean annual temperature rather than seasonal precipitation or temperature (Schoning et al. 2005).

2. A comparison between BSW records and mean July temperature for the mid-late Holocene from a chironomid-based reconstruction from a nearby lake (Barber and Langdon 2007). Comparisons between the reconstructions suggest that major shifts in surface wetness occurred simultaneously with changes in mean July temperature, supporting the hypothesis that temperature is a key driver of surface wetness changes.

In this paper we re-evaluate aspects of records with new data analyses and consider their wider 'distal' and 'proximal' climate drivers, in an attempt to present a consensus view of our current understanding of what drives the peatland surface wetness record.

2. A Holocene temperature reconstruction and bog surface wetness changes

Temperature records show spatial and temporal coherence over wide areas whereas precipitation displays much more variability (e.g. Croxton et al. 2006, Worral et al. 2006). This led to the hypothesis put forward by Barber et al. (2000) who postulated that lower temperature change (rather than precipitation) drove an almost identical response in peatlands 300km apart during the Little Ice Age. The observed clustering of the dates of prominent wet-shifts in European bogs also pointed in the same direction (Hughes et al. 2000, Barber and Charman 2003).

Barber and Langdon (2007) tested this hypothesis by comparing the first mid- to late-Holocene chironomid inferred temperature (CI-T) record for Northern Britain, from Talkin Tarn (Langdon et al. 2004), with two BSW records from Walton Moss (WLM17, covering the last 3000 years and WLM11, covering the last 7800 years (Hughes et al. 2000)). The CI-T record displayed significant fluctuations throughout the last 6000 years, with the highest temperatures around 5800 cal a BP, followed by a cooling trend, with specific cool events dated at around 5000, 4000, 3450, 2750, 2200 and 1800 cal a BP. There are other environmental factors which can effect changes in chironomid assemblages but these were discounted by other evidence at Talkin Tarn, and the fact that the reconstructed July temperature from the near-surface sample is within 0.2°C of the present (Langdon et al., 2004).

The start of the WLM17 record is one of high BSW between *c.* 2800-2600 cal a BP which accords with many other European bog records (van Geel et al. 1996, Barber et al.

2004), and which coincides with the CI-T record reconstructing falling temperatures from around 14°C to 12°C, the coolest period of the whole 6,000 year CI-T record. Later wet-shifts in this core are seen at *c.* 1700, 1200 and 600 cal a BP, which correlate with lower CI-T (Fig. 4 of Barber and Langdon, 2007). The WLM11 record of BSW was compared with the full 6,000 years CI-T record (Fig. 5 of Barber and Langdon 2007) and this suggested correlations between cooler phases of CI-T and higher BSW, most notably around 4000, 2750, 2200 and 1800 cal a BP. Whilst aware of factors such as the low resolution of the CI-T record and the lack of an independent palaeo-precipitation record, Barber and Langdon (2007) concluded that cooling events were concomitant with increases in BSW, and that their evidence supported a link over centennial timescales between summer temperatures and the peat-based palaeoclimate records.

Testing synchroneity – a re-evaluation of the record

One problem with testing hypotheses by comparison of proxy records is that of chronology. Here we apply new methods (Blaauw et al. 2008) of estimating uncertainty in the timing and strength of proxy signals to test the synchroneity of the CI-T and BSW records of Barber and Langdon (2007).

The probability that an event took place in a proxy archive during a certain period depends on *i*) the archive's chronological uncertainty (Fig. 1), *ii*) the strength of the proxy evidence for an event (i.e., signal strength), and *iii*) the duration of the period or window width. Here we assess all of these uncertainty sources together by calculating the event probability within a time window for the individual archives. Plant macrofossil and testate amoebae-based BSW and CI-T records were re-sampled assuming normal errors, and a running median was calculated (smoothing 11 for WLM17, 3 for lower resolution Talkin Tarn). All depths *d* where the running median increased beyond the error threshold were "flagged". This process was repeated 1000 times, after which for all depths of the individual proxies, the ratio of flagged iterations over all iterations was plotted (Fig. 2 c,e). The mean ratio from both proxies estimates the probabilities of heightened BSW in WLM17, and the ratio estimates the probability of reduced temperature in Talkin Tarn. Depths with probabilities <5% were omitted. For all age-model iterations, we find those depths *d*, with event probabilities $p(e_d)$, that fall within a time-window with boundaries y_{max} and y_{min}. The probability that an event took place in an archive during this time-window is equal to $1-\Pi(1-p(e_d))$. These probabilities are plotted for a range of window sizes in Figure 3.

Both cores date from c. 3500-3000 cal a BP to recent times (Fig. 1). The high-resolution radiocarbon ages for WLM17 result in a more precise age-depth model (Fig. 1b) than the low-resolution Talkin Tarn (Fig. 1e). The age-model of Talkin Tarn shows several outlying radiocarbon ages (Fig. 1d). Both archives indicate a number of (possibly synchronous) changes in accumulation rate. From the proxies, several likely events of changes to wetter or cooler conditions can be inferred (Fig. 2). The most prominent wetness increases are around 176-172 and 74-71 cm for WLM17, and decreases in temperature at 210, 115, and 80 cm for Talkin Tarn. Both cores show a number of additional moisture increases/temperature decreases with lower probabilities.

Instead of aligning the wetness and temperature changes by eye, here we keep the time-scales of both proxy archives independent, and calculate for each core the probability of increased wetness/decreased temperature during a given time window (Fig. 3). The probability of synchronous changes to wetter/cooler conditions during a time window are the product of the individual core's probabilities during that time window (Blaauw et al. 2007, 2008; Fig. 3c). The high precision age-model of WLM17 causes more defined peaks of probable wetness increases (Fig. 3a), whereas the temperature decreases of Talkin Tarn are more spread out (Fig. 3b) owing to larger chronological uncertainties. Probabilities of synchronous changes are highest at c. 1600 cal a BP; here both cores react synchronously (at 68% confidence level) within a time-window of 150 years. All other peaks can only be said to have been synchronous when using much wider time windows (e.g. 1000 years for c. 2900 cal a BP).

The sensitivity of summer deficit to late Holocene summer temperature changes

A further aspect of the debate is whether temperature variability in the late Holocene was large enough to force major changes in water balance. Simulations of response surfaces to changes in precipitation and temperature suggest that quite large temperature change would be needed to drive observed BSW variability (Charman 2007). A simulation of changes in P-E arising from the CI-T temperature changes at Talkin Tarn for the last 6000 years was carried out using 1961-1990 meteorological data for Carlisle as a baseline (Figure 4). The surface sample was used as the present (C-IT = 14.6 ^{o}C compared to 14.8 ^{o}C for 1961-1990 mean at Carlisle) and ΔC-IT values were applied to monthly temperatures to simulate annual deficit changes over the last 6000 years, with precipitation held constant. This results in a maximum decrease of up to 22 mm yr^{-1} in summer water deficit. Because inferred T reaches its maximum in the uppermost sample, there are no periods when the simulated deficit is greater than today. For comparison, during the period AD 1901-1990, the wettest decade (the 1920s) had an average 29 mm yr^{-1} reduction in deficit and the driest decade (the 1970s) had an average 10 mm yr^{-1} increase in deficit compared to the 1961-1990 mean. Individual years show much greater variability. The relatively small changes in summer deficit that are produced as a function of ΔC-IT over the last 6000 years compared to the observed variability over the 20th century, suggests that the magnitude of temperature change over mid-late Holocene was not large enough on its own to force the large variations in BSW over the last 6000 years.

3. Instrumental records and bog surface wetness changes

Previous analyses of instrumental data and high resolution water table reconstructions showed correlations between both observed and reconstructed water table and precipitation and were stronger than those with temperature (Charman et al. 2004). Annual P-E deficit was later shown to be an improved explanatory variable for reconstructed water table changes in northern England (Charman 2007), where annual deficit is the sum of P-E for all months where P<E. Both the duration of the deficit period and its magnitude vary considerably over the period of instrumental observations. To test this idea further, we have calculated annual deficit for Männikjärve bog and correlated this with instrumental water table measurements. Because this site has actual rather than reconstructed water table measurements, it removes errors associated with the reconstruction method.

The new correlations between the instrumental water table record and annual deficit for Männikjärve bog, Estonia (Table 1) show improvement over previous published correlations with T and P separately and in multiple regression (Charman et al. 2004). This supports the idea that the deficit is a more accurate representation of actual water table changes than P or T alone or in linear combination. i.e. the influence of daylength on evapotranspiration loss is significant, as this is the only other parameter considered in the simple estimates of E based on the Thornthwaite formula (Charman 2007). Furthermore, including an input from late winter snowfall and melt reduces rather than improves the correlation with measured water tables, especially for the biologically active summer months (Table 1). These new analyses of instrumentally observed water table change confirm the hypothesis that peatland water tables are driven by warm season moisture deficit. Correlations between annual deficit, summer temperature and summer precipitation for the Carlisle area and Männikjärve bog again show that precipitation is more highly correlated with annual deficit than temperature (Table 2). However, there is a high degree of autocorrelation between summer T and P in both regions; 0.512 ($p<0.001$) for Carlisle, 0.290 ($p<0.05$) for Männikjärve bog.

4. Proximal and distal drivers of bog surface wetness

Correlations between different proxy records and between proxies and instrumental data can help in understanding the direct proximal climate causes of proxy responses, but larger scale spatial patterns of proxy records are needed to understand the changes in the climate system that ultimately drove local climate variability. The fact that bogs across a large area of NW Europe and eastern North America display BSW changes at similar periods in time argues for a driver of change that operates over a large part of the Earth (Barber and Charman 2003, Barber 2006, Hughes et al. 2006). Sutton and Hodson (2005) have shown that the Atlantic Multidecadal Oscillation (AMO) controls the summer climate of northwest Europe and of eastern North America, and the historical AMO index is correlated with Swiss glacier advances and retreats (Denton and Broecker 2008). A pseudo-AMO mechanism is one possible driver for the large-scale spatial patterns shown by BSW records in the North Atlantic region.

European bog records of wet-shift dates (Hughes et al. 2000, Barber et al. 2003, Barber and Charman 2003) demonstrate that, within the limits imposed by the chronologies, there were coherent periods of change across large areas as well as links to other climate proxies and to documented climate change. More precise chronologies generated by high precision ^{14}C wiggle-match dating (e.g. Mauquoy et al. 2008) support this conclusion, and Hughes et al. (2006) suggest that at least some of the major changes may also have occurred in eastern North America. It has even been suggested that some patterns find parallels in the Southern Hemisphere. Chambers et al. (2007) reported a marked change in mire surface wetness identified in an Argentinean peat bog and ^{14}C wiggle-match dated to ca. 2706 cal a BP, very similar to a key change in European bogs wiggle-match dated to ca. 2690 cal a BP by Mauquoy et al. (2008). This interesting teleconnection needs to be tested by further Southern Hemisphere records but if it is confirmed then it argues for the existence of a *distal* driver in the form of a continuously-acting, areally-coherent and perhaps globally-synchronous parameter such as temperature, that can drive quasi-synchronous wet-shifts in bogs over a large area by affecting air mass circulation which in turn gives us a *proximal* driver in the form of precipitation change.

One issue with the precise correlation of records required to investigate these hypotheses is the need for large arrays of expensive geochronological analyses (e.g. Yeloff et al. 2006). The increasing use of tephras as precise tools for correlating changes between bogs and as “pinning points” in age-depth models (Langdon and Barber 2004, 2005) can reduce this need, especially if the tephras are also more precisely dated through wiggle-matched AMS radiocarbon ages (Barber et al. 2008). Another promising approach to developing regional records that reduce site specific biases in the magnitude and timing of changes and provide a basis for comparison of large-scale spatial patterns of variability is that of “stacking” and tuning the records (Charman et al. 2006).

The drivers of observed change also may vary over time. For example, there is clear evidence that in the western Atlantic, the main drivers of palaeoclimatic change shifted during the Holocene. Before 6.8 cal ka BP, BSW variability in eastern Newfoundland was dominated by its proximity to meltwater discharges from the decaying Laurentide ice-sheet (Hughes et al. 2006). Cold meltwater plumes from lake Agassiz (Barber et al. 1999) and

Ungava (Jansson and Kleman 2004) probably had a direct impact upon local temperatures and the development of fog banks as well as disrupting local North Atlantic Deep Water (NADW) production in the Labrador Sea, which would in turn cause regional cooling through the impact on Gulf Stream heat transport and more widely through disruption of the Atlantic Meridional Overturning Circulation (AMOC). Following the end of significant meltwater discharges at 6.8 kcal BP (Carlson et al. 2008), ocean records suggest that sea surface temperatures and salinity stabilised in the Labrador Sea and western Atlantic and that subsequent climate might be expected to be relatively complacent (Solignac et al. 2004), in contrast to the continuing centennial-to-millennial-scale variability of the mid- and eastern Atlantic (Berner et al. 2008). However, the terrestrial palaeoclimate records from Newfoundland (Hughes et al., 2006) and continental N. America (e.g. Viau et al. 2006) indicate that centennial-to-millennial-scale palaeoclimatic variability persisted throughout the mid- to late Holocene, albeit with a lower magnitude than earlier events. The persistence of millennial-scale variability in mid to late Holocene BSW records of eastern Newfoundland supports the idea of a teleconnection with the eastern Atlantic driven by long-term trends in the AMO.

The forcing driving the manifest changes in regional BSW patterns is still debated but two key factors are being considered. There is a long established suggestion that key changes in BSW reflect solar variability as the ultimate driver (van Geel et al. 1996), supported by precisely dated records (e.g. Mauquoy et al. 2002) and data compilations (Charman et al. 2006), suggesting this resulted in changes in the location and strength of westerly storm tracks. However, other studies have suggested that the relationship between BSW and solar forcing is weaker and oceanic forcing may be equally important (e.g. Blundell et al. 2008), as shown by changes in phase with measures of ice rafted debris (Bond et al. 2001) and Iceland-Scotland Overflow Water (Bianchi and McCave 1999). Further papers have examined these solar and oceanic forcing factors using both improved chronologies on BSW and other proxies and time-series analyses to explore periodicity of change (Turney et al. 2005, Swindles et al. 2007, Snowball and Muscheler 2007, Mauquoy et al., 2008). The evidence for cyclic change, first found by Aaby (1976) has also been found by other studies (e.g. Barber et al. 1994, Chambers and Blackford 2001) and linked to cycles in ocean cores of c. 550 and 1100 years (Hughes et al. 2000, Langdon et al., 2003). However, given the range of periodicities so far detected in different records, it is by no means clear that either solar or ocean-driven forcing is dominant in the BSW records. Reviews of the wider palaeoclimatic evidence suggest a more complex combination of both forcing and feedbacks were operating throughout the Holocene (Wanner et al. 2008, Charman 2009)

5. Conclusions: a consensus view?

There are several key conclusions from our re-analysis of the dataset of Barber and Langdon (2007), the instrumental data associated with British and Estonian peatlands (Charman et al. 2004, Charman 2007) and consideration of large-scale climate drivers and teleconnections.

- The suggested correlation between Holocene summer temperature variations and BSW is plausible, appears to vary through time and improved chronological control is required to test this further (Figure 3).
- The magnitude of temperature variability alone in the late Holocene is unlikely to have altered the annual deficit sufficiently to explain reconstructed BSW changes (Figure 4).
- Annual deficit is the best explanatory variable for measured water table changes (Table 1). Recent changes in annual deficit are more strongly correlated with precipitation than with temperature, in both Estonia and the UK (Table 2).
- Large-scale spatial patterns of change are needed to understand the relationship between distal and proximal drivers of BSW change. The logical progression here is that solar and/or ocean forcing are the ultimate drivers of regional atmospheric circulation, which then provide the proximal driver of water balance at individual bog sites.

The discussion in previous literature has commonly focused on whether temperature or precipitation is the key variable driving reconstructed BSW records. The results of Barber and Langdon (2007) and work on instrumental data (Charman et al. 2004, Charman 2007) can certainly be seen as contradictory in their conclusions on this issue and the analyses presented here provide at least some support for existing interpretations of the individual data sets. However, it seems much more likely that in practice, the P-E balance and BSW depends to some extent on both precipitation and temperature. In the current climate state of northwest Europe, precipitation clearly plays a more dominant role than temperature for driving BSW, as shown by the analysis of instrumental data of instrumentally observed and reconstructed water table changes. Furthermore, the magnitude of temperature variability over the mid-late Holocene only produces relatively small variations in annual deficit and can only partly explain reconstructed BSW changes. However, temperature is clearly correlated to some extent with observed summer deficit over the last 100 years (Table 2) and probably with BSW over the past 6000 years (Figure 3). The negative correlation between instrumentally observed summer temperature and summer precipitation shows that often both are driving the moisture deficit in the same direction at the same time. In other words, summers with low deficits are associated with both higher precipitation and lower temperatures. This relationship is stronger for the Carlisle area than for Mannikjarve bog, probably due to the east-west continentality gradient in Europe and the ratio of summer/winter precipitation (Zveryaev 2004).

Summer precipitation in western areas of the UK is mainly enhanced by stronger or more frequent westerly airflow, bringing cool moist air from the Atlantic, greater cloudiness and lower air temperatures. The effect of the westerlies is less important in eastern Europe, because of a higher proportion of precipitation from convective rainfall during the summer. The interpretation of European BSW records in terms of strength and position of westerly airflow is supported by previous work on peat and other proxies (e.g. Charman and Hendon 2000, Magny 2004, de Jong 2006, Blundell et al. 2008). New data on stable isotopes in *Sphagnum* peat also suggests significant shifts in air mass trajectories at times of major BSW changes in northern England (Daley et al. submitted). Thus the BSW records from Europe are perhaps best interpreted as an index of the strength and position of summer westerly airflow rather than as temperature or precipitation *per se*. Whilst at present the annual deficit and BSW are driven primarily by precipitation, it is possible that temperature played a more important role in the past, especially in the early-mid Holocene when summer isolation was stronger. Further independent records of summer temperature and precipitation based on secure chronologies are required to test this hypothesis further.

To make progress on applying BSW records to understanding Holocene climate variability, there should be a greater focus on temporal and large scale spatial variability in relation to atmospheric and ocean circulation changes. Coherent patterns of major phases of high/low water balance in Europe are emerging and these show interesting relationships with low latitude water balance, and North Atlantic ocean circulation and sea surface temperatures (Magny 2004, Verschuren and Charman 2008, Berner et al. 2008), interpretable as major reorganisations of the ocean-atmosphere system.

Our findings relate primarily to *Sphagnum* dominated peatlands in mid-latitude Europe where climate is relatively cool and moist and variability in P and T is strongly related to changes in westerly airflow. It is possible that in other regions, the sensitivity of P-E to temperature and precipitation is different and studies elsewhere should evaluate relationships between these parameters to help interpretation of Holocene BSW records.

Acknowledgements

The data for Männikjärve bog were provided by Tooma Mire Station, Estonian Meteorological and Hydrological Institute. The work on Talkin Tarn was funded by the Natural Environment Research Council (Grant No. GR9/04313 to KEB) and supported by radiocarbon age estimates from the NERC Radiocarbon Facility. EK was supported by Estonian Science Foundation grant No 6755.

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List of tables

	Annual WT	*JJA WT*
Annual Temperature (T)	*0.015	
JJA T		-0.369
Annual Precipitation (P)	0.666	
JJA P		0.811
JJA P and JJA T (multiple r)		0.823
Annual deficit	0.771	0.885
Annual deficit plus snow melt	0.727	0.691
Annual total deficit	0.651	0.529

Table 1: Linear correlations between instrumental water table measurements and meteorological data for Männikjärve bog, Estonia 1951-2000. Annual deficit – sum of months in moisture deficit (c.f. Charman, 2007), Annual deficit plus snow melt – as above plus total late winter (Jan to March) snowfall, Annual total deficit – sum of moisture balance for all months. All significant at p<0.01 except *NS.

Climate variable	Carlisle region, UK	Mannikjarve, Estonia
JJA Precipitation	0.835	0.858
JJA Temperature	-0.611	-0.434*
Combined (multiple r)	0.862	0.880

Table 2: Correlations between climate variables and the annual deficit for Carlisle (1901-1995) and Männikjärve bog (1951-2000). All correlations significant at p<0.001, except *, which is significant at p<0.005.

List of figures

Figure 1. Bayesian age-depth models for the two cores (WLM17: a-c, Talkin Tarn: d-f). The radiocarbon (black circles in a,d) dates were matched against the IntCal04 calibration curve (Reimer et al. 2004). Red circles show ages rejected through Bayesian outlier detection and down-weighting. The age distributions are graphed as grey-scales (Blaauw et al. 2007) (b,e). Besides the depths themselves, their proxy values can also be plotted against calendar age (c,f). Dark areas indicate secure sections of the age-models, while lighter grey areas warn us of sections of a core where the chronological uncertainty is large.

Figure 2. WLM17 (a-c) and Talkin Tarn (d-e) proxies plotted against depth (black lines). Probabilities of increasing moisture were calculated, through finding those depths with major decreases in DCA (a), or major increases in either testate-amoebae derived water levels (b) or chironomid inferred temperature from WA_PLS_3 (d). Probabilities of moisture increases are plotted as histograms (c, e). For WLM17, the means of these proxy increases estimate the probabilities of moisture increases (histograms in c).

Figure 3. Events of increased moisture in WLM17 (a), Talkin Tarn (b), and both records combined (c). Probabilities of increased moisture were calculated using a time-window approach (Blaauw et al. 2007, 2008). Time windows moved at 5 year jumps. Window sizes were 10 to 1000 years in steps of 50 years, shown as rainbow colours in the probability histograms. The probabilities of simultaneous events between WLM17 and Talkin Tarn within time-windows are shown in (c).

Figure 4. Simulated changes in summer deficit (b) from temperature changes inferred from Talkin Tarn (a), assuming constant precipitation at 1961-1990 mean values. Summer deficit is the sum of the deficit of all months in the year where P<E. Dashed lines show the summer deficit for the wettest (1920s) and driest (1970s) decades from 1901-1990.

Figure 1 colour version

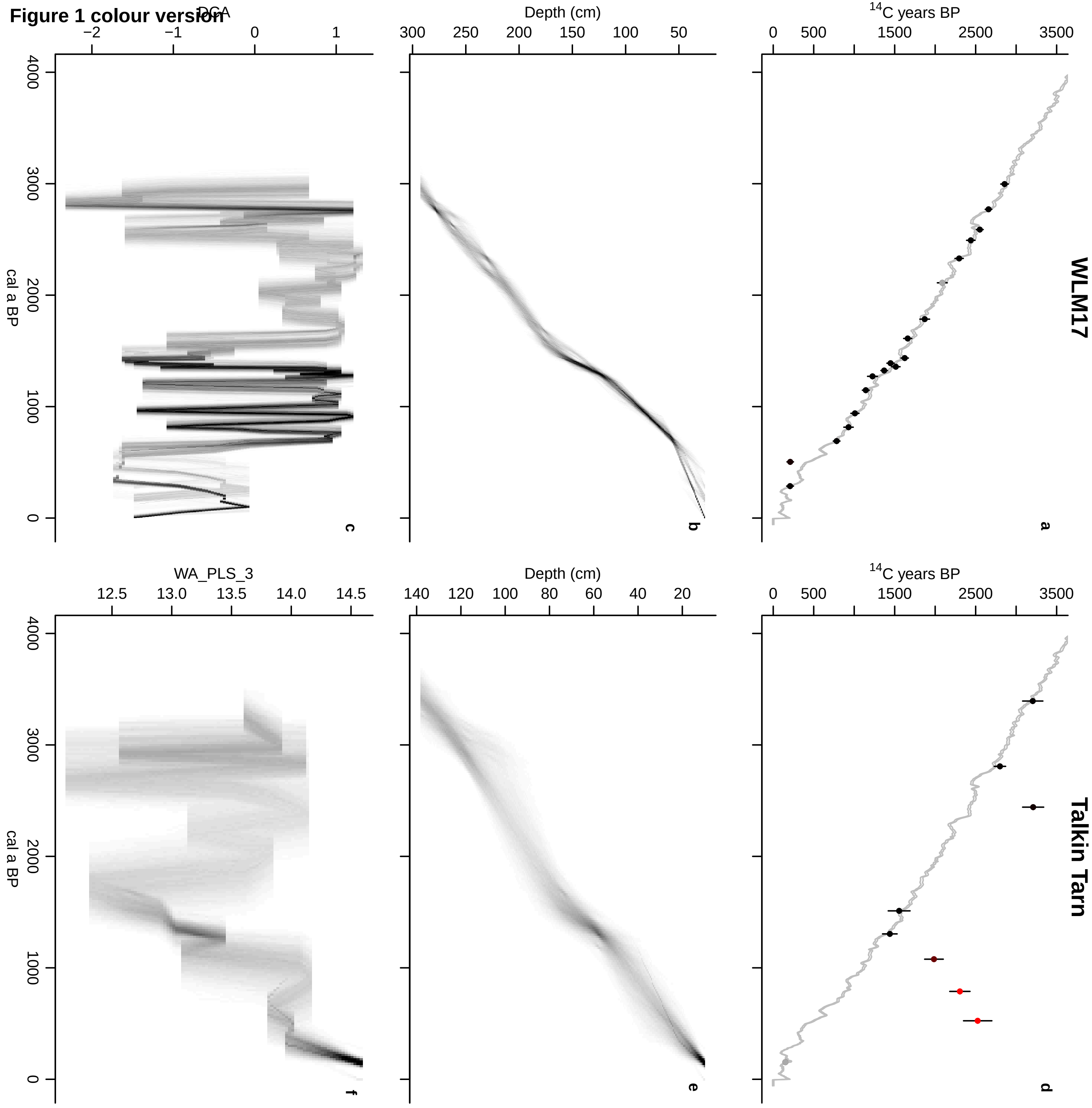


Figure 1 greyscale version

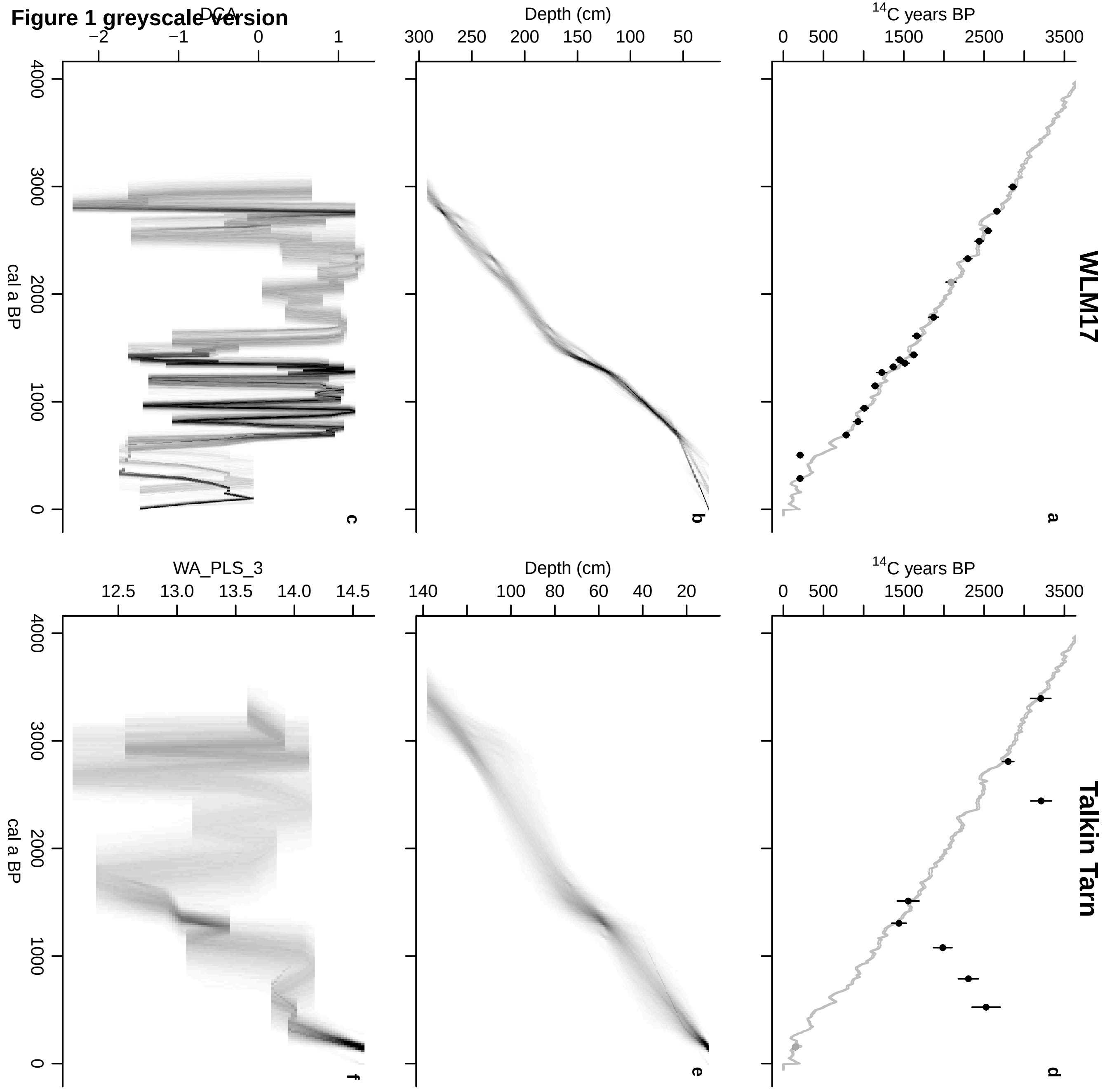


Figure 2

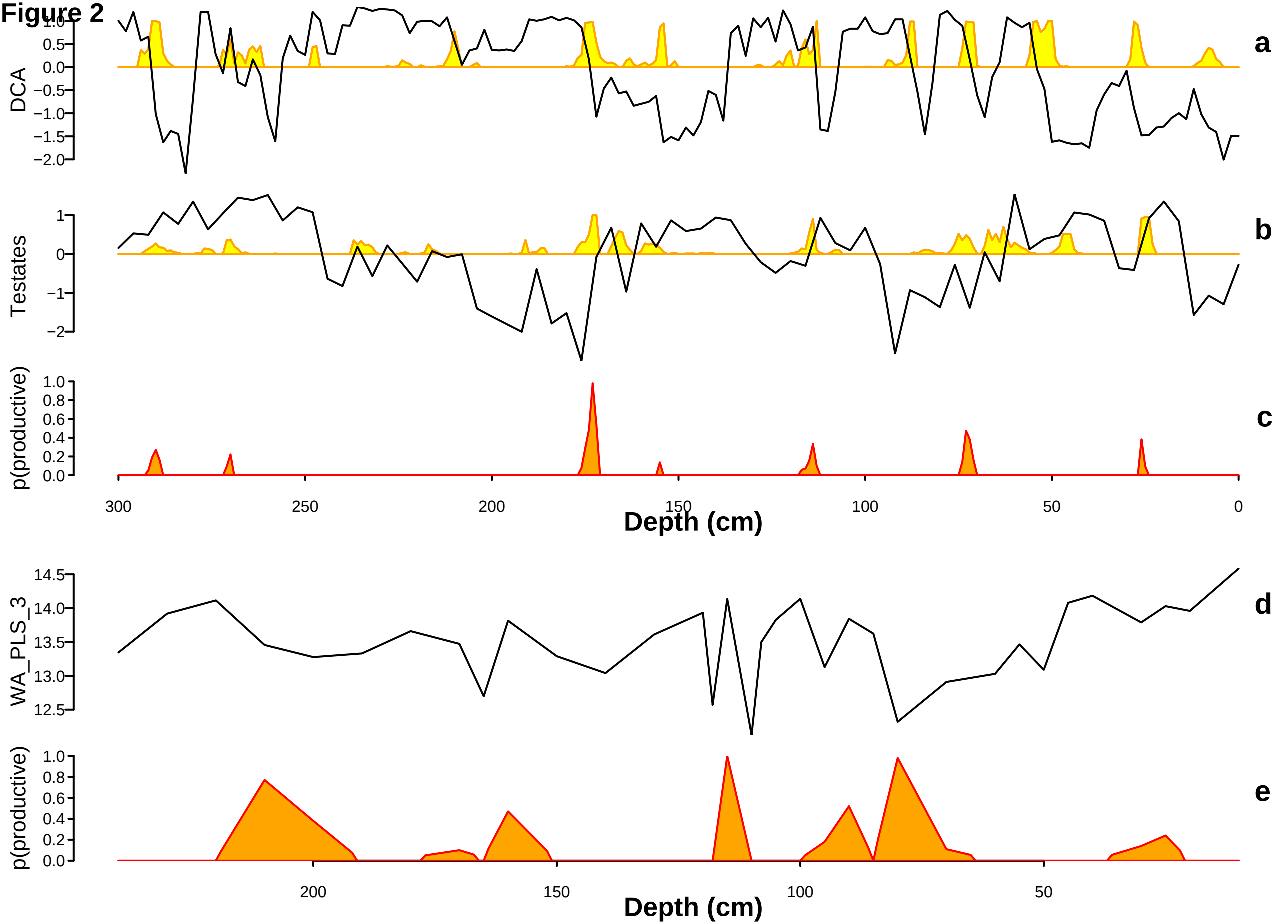

DCA
1.0
0.5
0.0
−0.5
−1.0
−1.5
−2.0
a
Testates
1
0
−1
−2
b
p(productive)
1.0
0.8
0.6
0.4
0.2
0.0
c
300
250
200
150
100
50
0
Depth (cm)
WA_PLS_3
14.5
14.0
13.5
13.0
12.5
d
p(productive)
1.0
0.8
0.6
0.4
0.2
0.0
e
200
150
100
50
Depth (cm)

Figure 3 colour version

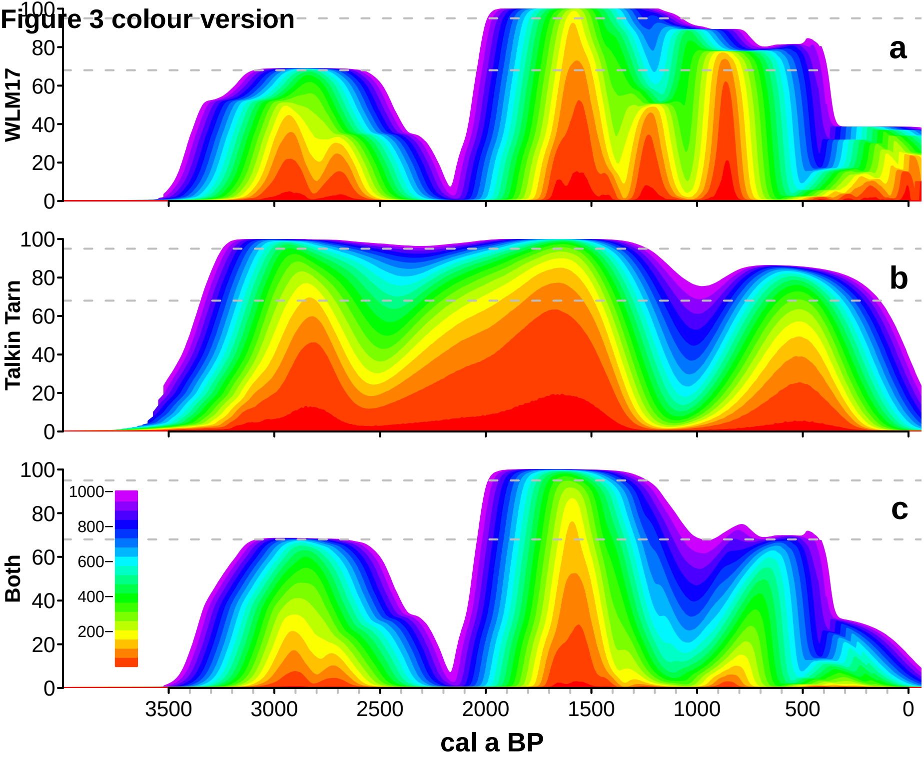


Figure 3 greyscale version

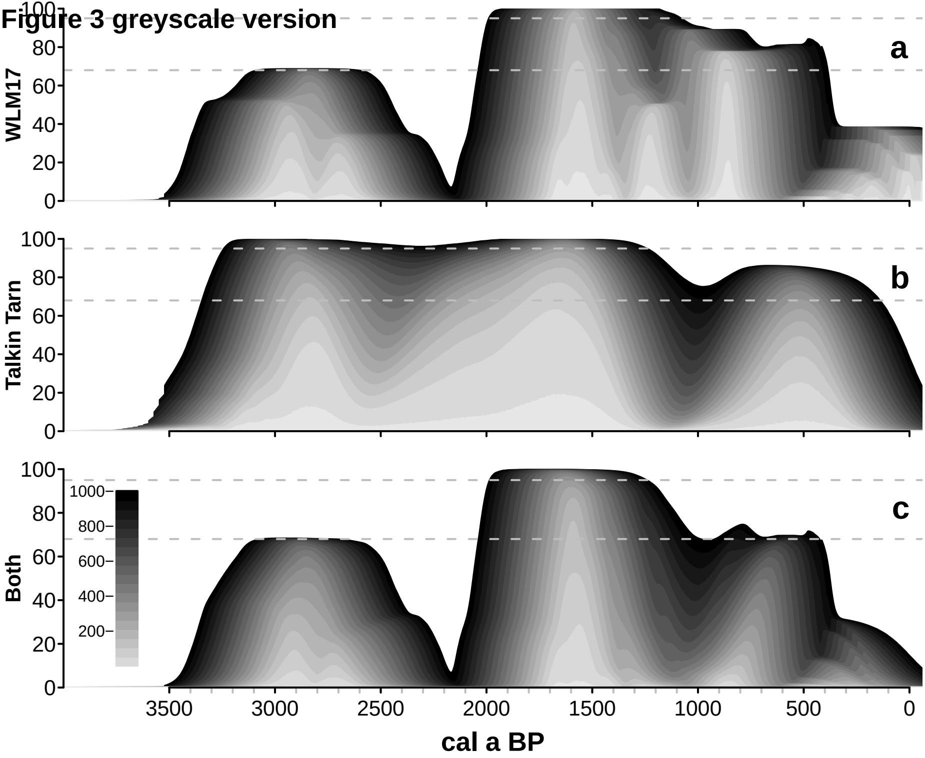
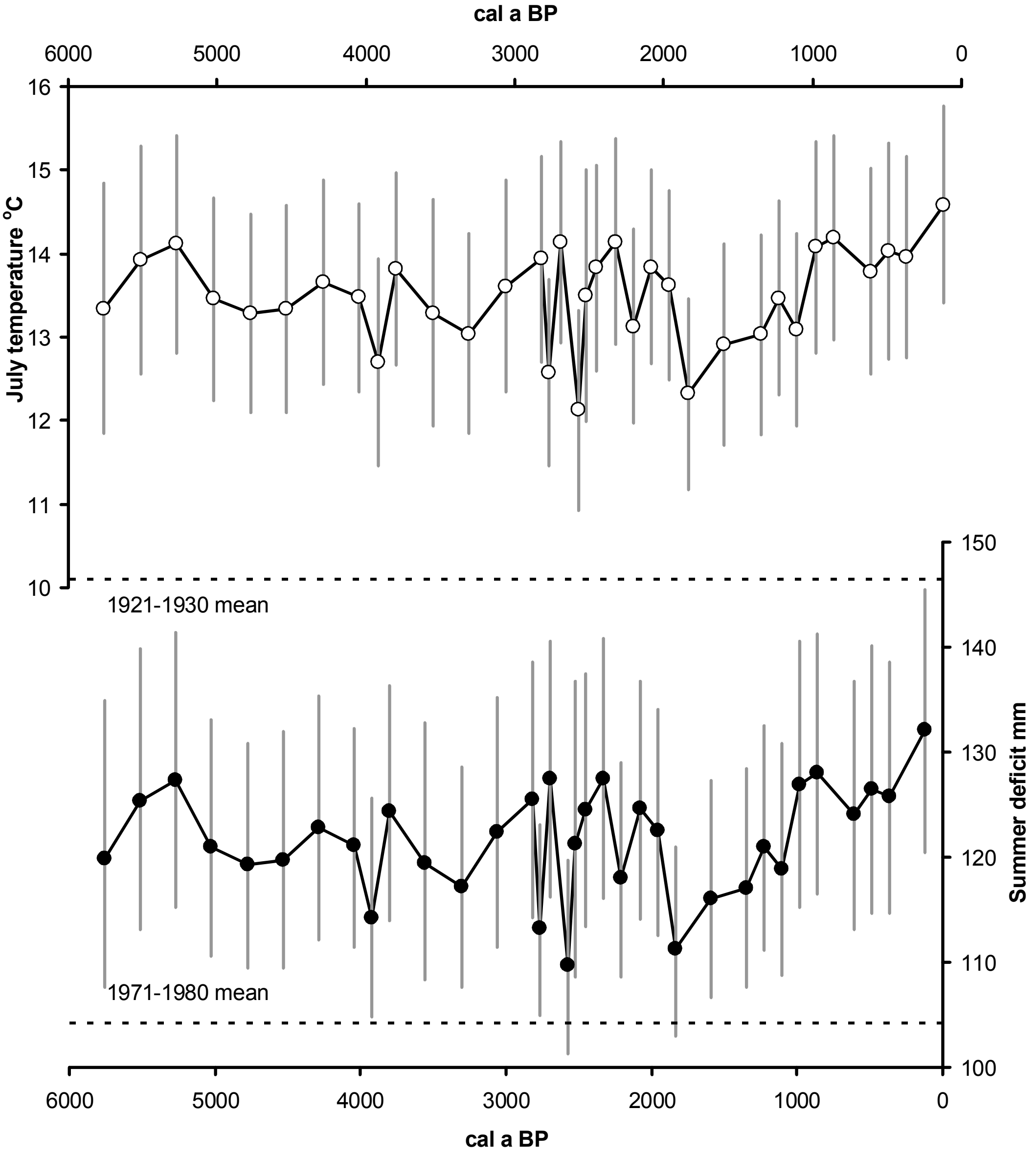


Figure 4



cal a BP
6000
5000
4000
3000
2000
1000
0
16
15
14
13
12
11
10
July temperature °C
1921-1930 mean
150
140
130
120
110
100
Summer deficit mm
1971-1980 mean
6000
5000
4000
3000
2000
1000
0
cal a BP