Initial Mediterranean response to major climate reorganization during the last interglacial-glacial transition

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Abstract

Millennial-scale Dansgaard Oeschger (DO) variability at northern high latitudes has influenced climatic and environmental conditions in the Mediterranean during the last glacial period. There is evidence that the hemispheric transmission of the DO variability occurred at the end of DO event 25; however, the exact timing and the trigger that activated the environmental response in the Mediterranean remains incompletely understood. Here, we provide evidence that the clear millennial-scale teleconnection between Greenland and the Mediterranean started at ~111.4 ka BP and was initiated by a sub-millennial scale cooling in Greenland (GI-25b). High-resolution sediment proxies and the pollen record of Lago Grande di Monticchio (MON), Italy, reflect climatic instability during the last millennium of the last interglacial, which was characterised by a first and short cooling episode (MON 1) at 111.44 ± 0.69 ka BP, coinciding with the Greenland cold sub-event GI-25b in duration and timing (within dating uncertainties). MON and Greenland (NorthGRIP ice core) also agree in recording a subsequent warm rebound phase that abruptly culminated in the stadial MON 2 / GS-25, marking the transition into the last glacial period. Our results show that the GI-25b triggered an early environmental response at MON to centennial-scale climate change in Greenland as a prelude to the millennial-scale teleconnection that was maintained during the glacial period.

Keywords: Palaeoclimatology, Last Interglacial-Glacial transition, millennial-scale variability, Mediterranean, varved sediments
1 Introduction

The transition from the Last Interglacial (LI) into the Last Glacial in the Northern Hemisphere (NH) developed differently from polar to subtropical regions. While the initial build-up of the NH ice sheets started ~122 thousand years before AD 2000 (ka b2k) (NGRIP MEMBERS, 2004; Rasmussen et al., 2014) and was associated with declining northern summer insolation, southern European climate remained warm and relatively stable until ~110 ka AD 1950 (BP) when a dramatic shift in the environment marked the end of the LI in this region (Woillard, 1979; Kukla et al., 1997; Shackleton et al., 2002; Shackleton et al., 2003; Tzedakis, 2003; Sánchez-Goñi et al., 2005; Brauer et al., 2007). This change corresponds to a biostratigraphic boundary that defines the transition to what is known as the Mélisey I stadial in central and southwestern European pollen records (Woillard, 1979; Kukla et al., 1997). The direct correlation between the European biostratigraphy and marine proxy-based stratigraphies (i.e. North Atlantic ice-rafted detritus (IRD), stable isotopes and sea surface temperature) developed by Sánchez-Goñi et al. (2005), Shackleton et al. (2002) and Shackleton et al. (2003) revealed that (1) the North Atlantic cold event C26 at ~119 ka BP was associated with a western European cooling marked by the increase of hornbeam trees within an oak-dominated forest at 42°N and probably the spread of cold-tolerant trees in Europe north of 50°N and (2) the Mélisey I stadial and the North Atlantic cold event C24 at ~110 ka BP occurred contemporaneously. Recently, Martin-Puertas et al. (2014) correlated the Mélisey I with Greenland stadial GS-25 (Greenland event stratigraphy, NGRIP members, 2004; Rasmussen et al., 2014). GS-25 (Greenland ice-core event stratigraphy), C24 (North Atlantic sea surface temperature stratigraphy) and Mélisey I (central Europe pollen stratigraphy) are the regional representatives of the first abrupt climatic oscillation within a series of millennial-scale climatic fluctuations in the NH during the last glacial period, known as Dansgaard Oeschger (DO) events (McManus et al 1994; Chapman and Shackleton, 1999; NGRIP MEMBERS, 2004, Oppo et al., 2006; Capron et al., 2012). DO events typically consist of an abrupt warming
followed by an interval of mild conditions (Greenland interstadial, GI) and a gradual cooling that culminates in a cold phase (Greenland stadial, GS). The duration of the GI and GS varies between a century and a few millennia (Sánchez Goñi and Harrison, 2010; Wolff et al., 2010; Rasmussen et al., 2014). This millennial-scale variability is linked to changes in the strength of the Atlantic Meridional Overturning Circulation, which influences the climate globally (Broecker, 1998, Barker et al., 2011). Prior to GS-25, a weak cold phase (GS-26) occurred between 119.14 and 115.37 ka b2k (NGRIP MEMBERS, 2004; Rasmussen et al., 2014) with presumed counterparts in the subtropical Atlantic Ocean (cold event C26) (Chapman and Shackleton, 1999) and in northern, western and central Europe (Chedaddi et al., 1998; Sánchez-Goñi et al., 2005). At centennial time scales, the updated INTIMATE event stratigraphy from the Greenland ice cores proposes a further sub-division of the following warm interval GI-25 into three sub-events: GI-25c (115.37–111.44 ka b2k); GI-25b (111.44–110.94 ka b2k); and GI-25a (110.94–110.64 ka b2k) (Rasmussen et al., 2014). GI-25c represents the main warm phase of DO 25, while GI-25b reflects a 500-year cooling episode, which was followed by GI-25a, a 300-year rebound preceding the stadial phase GS-25 that corresponds to the cooling stage of DO 25.

Palaeotemperature records from the North Atlantic Ocean reflect a weak and short cooling episode (cold event C25) associated with a weak IRD event prior to cold event C24 (McManus et al., 1994; Chapman and Shackleton 1999). Assuming a correlation between cold event C24 and Mélisey I, Lehman et al. (2002) correlated the cold event C25 to an abrupt transition from deciduous trees to taiga forest in the pollen record of Grande Pile (France), known as the Woillard event (WE) (Woillard, 1979), suggesting that the WE was as a consequence of changes in the Atlantic temperature altering climate and vegetation in Europe. The WE is dated at ca. 111 ka BP in Grande Pile (Kukla et al., 1997) and lasted about 300 ± 150 years (Woillard, 1979; Kukla et al., 1997). In the Mediterranean, the pollen record of Lago Grade di Monticchio (MON) shows millennial- to centennial-scale variations, which
were correlated with the Grande Pile pollen record, adopting the same nomenclature (Brauer et al., 2007). The WE was identified in the MON pollen record centred at 111.16 ± 5.55 ka BP with an estimated duration of ca. 320 years (Brauer et al., 2007; Allen and Huntley, 2009), but it is represented by a single pollen sample only. Therefore, the duration estimate of the WE in the MON pollen record is limited by pollen data resolution, as we will discuss below. Most recently, a Mediterranean speleothem (BMS1) from Bue Marino Cave, Sardinia, Italy, provided evidence for a cool episode centred at 112 ±0.52/-0.59 ka followed by warmer conditions at 111.76 ±0.43/-0.45 ka, which were correlated with the Greenland sub-events GI-25b and GI-25a, respectively (Columbu et al., 2017). Both, the MON and BMS1 proxy records are independently dated, which allows reliable estimates of the duration of the climatic oscillations and comparison with the Greenland ice cores without a priori assuming synchronicity of the events. While the BMS1 stable isotope record reveals hydroclimatic and temperature information, the MON sediment sequence reflects climate-induced environmental changes and the preservation of varves in the lacustrine sediments allows us to develop a multi-proxy approach even at seasonal resolution (Brauer et al., 2007; Martin-Puertas et al., 2014).

Here we present an updated and detailed study of the final phase of the LI from the MON varved sediments. We have extended the high-resolution proxy records published by Martin-Puertas et al. (2014) from ca. 111.4 ka BP back to ca. 120 ka BP and evaluate sub-millennial-scale climate variability during the last glacial inception. Our objectives are to: (1) double check the occurrence of the WE in the MON proxy records; (2) evaluate differences in proxy response (pollen vs. sediment proxies) to sub-millennial-scale climate variability in order to define the change points more precisely; and (3) estimate the onset of the teleconnection between the initial build-up of the NH ice sheet and its environmental impact on the Mediterranean beyond the well-established correlation at millennial time scales. For this
particular objective, we have avoided wiggle matching of the climatic oscillations and have
compared high-resolution proxy records that are independently dated.

[Fig. 1]

2 Site description and Methods

2.1 Study site

Lago Grande di Monticchio (40°56' N, 15°35' E, 656 m a.s.l.) is a maar lake formed during the
final phreatomagmatic eruptions of Monte Vulture volcano (southern Italy) about 132,000
years ago (Brocchini et al., 1994; Stoppa and Principe, 1998). Maximum water depth is 36 m,
but much of the lake is a shallow shelf sloping from the shoreline towards 12 m depth from
where the cores were taken (Allen et al., 1999; Brauer et al., 2007). The lake area is 0.4 km²;
its catchment, which is mainly composed of K-alkaline phonotephrites, covers 2.37 km². The
maximum elevation in the catchment is 956 m a.s.l., with a maximum relief of 300 m. The lake
is spring fed and thus regulated by the local groundwater system and by rainfall and surface
runoff. Mediterranean summer drought might cause the lake level to lower (Hutchinson,
1970).

2.2 Sedimentological analyses

The MON sediment record is ~130 m long. We have worked on a sequence of 5 m length
between 79 and 84 m sediment depth of the composite profile J/M/O, which combines the
best-preserved sequences of sediment cores J, M and O (Fig. 1b) (Martin-Puertas et al., 2014).
Varve counting and microfacies analyses were done on sediment thin sections using a
petrographic microscope (Brauer et al., 2007; Martin-Puertas et al., 2014). The geochemical
composition of the sediments was obtained by micro X-ray fluorescence (XRF) scanning of the
10-cm long impregnated sediment blocks that were used to produce the thin sections
mentioned above, allowing direct comparison of element abundances with microfacies data.
The analytical system used for elemental scanning was an EAGLE III XL µ-XRF spectrometer deployed at the GFZ Potsdam (see details in Brauer et al., 2007).

2.3 Pollen analyses

Thirteen new pollen samples taken between 80.45 and 80.70 m sediment depth have been analysed following standard methods applied in previous pollen studies at the site (Allen et al., 1999; Brauer et al., 2007; Allen and Huntley, 2009) in order to increase the temporal resolution of the pollen record published by Brauer et al. (2007) between 111.25 and 111 ka BP, where the WE was assumed by Brauer et al. (2007). We integrated the new pollen samples into the previous pollen record in order to get a sample resolution higher than 20 years (according to the age-depth model). In the particular case of peak previously correlated to the WE, we have increased the sample resolution to 5 years (average sedimentation rate is 0.39 mm / year) by adding four contiguous samples below and four additional samples above the published sample that reflects the WE. We have also calculated pollen influx defined as the number of pollen grains accumulated per unit of sediment surface area and per unit of time (grains x cm\(^{-2}\) x year\(^{-1}\)) in order to get a semi-quantitative measure of both vegetation distribution and abundance (Davis et al., 1972; Hicks and Hyvärinen, 1999). Anomalies in the proxy records were considered significant where increases or decreases are greater than twice the standard deviation of the proxy dataset.

2.4 Chronology

The most recent version of the MON varve chronology for the interval comprising the transition from the LI into the LG (MON-2014; ca. 76 to 111.4 ka BP; 50.22–81.53 m sediment depth) was published by Martin-Puertas et al. (2014). Varves were counted on the composite profile J/M/O using an accurate counting method based on detailed sediment microfacies analysis (100x microscopic magnification) and the measurement of the thickness of the different sub-layers constituting the varves. In this study we extended the MON-2014 chronology down to 84 m sediment depth using the same counting method. In order to
provide a relative counting error as uncertainty range for the duration of the climatic phases that are discussed below, we compared the different varve counts available for the MON sediment record during the study interval between 79.00 and 84.00 m sediment depth. Those are MON-07 (79.00–84.00 m; Brauer et al., 2007), MON-2014 (79.00–81.53 m; Martin-Puertas et al., 2014) and the new count (81.53–84.00 m; this study). In terms of absolute age uncertainty, the MON varve chronology was validated by tephrochronology and its absolute uncertainty is up to 5% according to the 2σ error calculated by the discrepancy between the mean radiometric / radioisotopic and varve ages of tephra layers identified in the MON sediments (Wulf et al., 2012). As a new contribution to the MON chronology, we have applied Bayesian age-depth modelling for a more thorough integration of the varve and tephra records, aiming at minimising the absolute age uncertainty during the study interval (109–120 ka BP) as much as possible (<5%). For this purpose, we ran a P_Sequence deposition model implemented in OxCal 4.2 (Bronk Ramsey, 2008) and included recently updated 40Ar/39Ar dates for the Mediterranean marine tephra correlatives X-5 (106 ± 1.3 ka BP, Giaccio et al., 2012) and X-6 (109.5 ± 0.9 ka BP, Regattieri et al., 2017) for the dated points in the model and the last relative varve time scale as the Z value in the model (Bronk Ramsey, 2008; Blockley et al., 2008). As the model is based on relative varve numbers we have not included interpolation modelling (Table 1, Fig. 2).

3 Results, age modelling and discussion of the presumed Woillard Event in the Lago Grande di Monticchio sediment record

A total of 8500 varves were counted from 81.53 to 84.00 m sediment depth, which extends the previously published MON-2014 chronology back to ca. 120 ka BP. The maximum cumulative counting error (i.e. the differences between the three varve time scales where they overlap) at the base of the study interval was ± 386 varves. As this counting error is lower than the modelled uncertainty at tephra layer X-6 (± 697 years) (Table 1), we have considered
the latter as the maximum absolute age uncertainty for the entire study interval, and the age
given by the varve time scale as the most reliable date for the observed oscillations. Thereby,
it has been possible to reduce the absolute age uncertainty from 5% down to 1.2%.

[Fig. 2]

Figure 3 shows a set of sedimentological, geochemical and pollen data from the MON
sediment sequence as indicators of environmental and climatic processes affecting the
lacustrine system, and changes in the surrounding landscape, from 120 to 109 ± 0.69 ka BP.
The rapid reduction in woody taxa pollen (60–40%) at 110.43 ± 0.69 ka BP represents the
onset of the Mélisey I stadial and the biostratigraphic end of the L1 in the MON sediment
record (Brauer et al., 2007; Allen and Huntley, 2009). The WE at MON was associated with a
rapid but transient decrease in woody taxa pollen (from 95% to 70%), with grassland
expanding on some part of the catchment, that occurred towards the end of pollen sub-zone
LPAZ 21a and ca. 700 years before the onset of Mélisey I / LPAZ 20 (the MON pollen
stratigraphy is fully described in Allen and Huntley [2009]).

[Fig. 3]

Although the reduction in woody taxa pollen is less pronounced than during Mélisey I, the WE
fluctuation in the pollen record was interpreted as a cool and dry episode within a warm
period (Allen and Huntley, 2009). In line with our first objective, new pollen samples were
added to verify the signal of that fluctuation, which is so far based on one single pollen sample
only (Fig. 4). The analyses of the new pollen samples confirm the existence of a change in
pollen data which is represented by four contiguous pollen samples that indicate a minor
reduction in the amount of tree pollen taxa from 111.185 to 111.150 ± 0.69 ka BP. This is
supported by a negative oscillation in woody taxa pollen influx and woody taxa percentages (Fig. 4a). Both the transitions at the beginning and end of this fluctuation occurred in less than 8 and 13 varve years, respectively (Fig. 4a). This fluctuation was, however, much shorter than previously assumed, i.e. 40 ± 5 years instead of the ca. 320-year oscillation reported by Allen and Huntley (2009). Considering this new information, we suggest that this fluctuation was too short to encompass a transition from forest to grassland and a subsequent return to forest. It is more likely that the pollen data reflect a short episode of reduced tree pollen production. Such an episode may reflect a short-lived shift in climatic conditions, sufficient to reduce the vigour of the trees and hence also reduce their flowering, but not sufficient to cause their death before climate conditions reverted once again to favour the trees.

Another reason to revise the correlation of this episode with the WE defined in Grande Pile (France) is based on chronology. The WE in Grande Pile preceded the transition to Mélisey I by four millennia (Kukla et al., 1997), while the observed tree pollen decline at Monticchio occurs only ca. 700 years before Mélisey I was only ca. 700 years (Brauer et al., 2007; Allen and Huntley, 2009). As both chronologies are based on annual layer counts, these two events in the pollen records are therefore unlikely to reflect the same climatic signal.

Martin-Puertas et al. (2014) reported that the assumed WE in the MON pollen record coincided with the first cooling period at the end of the last interglacial identified in the sediment proxies, i.e. MON 1 (111.23–111.01 ± 0.69 ka BP; Martin-Puertas et al., 2014). These proxies indicate stadial-like (cold) conditions through the deposition of thick siderite varves (microfacies 2b in Martin-Puertas et al., 2014), while thin organic-diatomaceous varves (microfacies 1a in Martin-Puertas et al., 2014) occurred during interglacial (warm) and interstadial (mild) conditions. The downcore extended dataset presented in this study,
however, revealed a short preceding phase of thick siderite varves, which appeared already ca. 220 years earlier at 111.45 ka BP and lasted until 111.39 ± 0.69 ka BP (Fig. 3, 4b). Therefore, we must re-define MON 1 as a slightly longer cool period with a duration of 440 ± 9 years, lasting from 111.45 to 111.01 ± 0.69 ka BP compared to earlier reports defining MON 1 from 111.23 – 111.01 ± 0.69 ka BP (Martin-Puertas et al., 2014). The brief 40 ± 5 year decline in tree pollen that we constrained by our new high-resolution pollen analyses occurred in the middle of MON 1 and consequently we interpret this phase as an integral part of MON 1. The first truly stadial episode recorded in the sediments occurred about 1000 years later and is termed MON 2 (110.43 – 108.63 ± 0.69 ka BP). MON 2 agrees in timing and duration with Mélisey I in the MON pollen record (Fig. 3; Fig. 6 in Martin Puertas et al., 2014) and marks the end of the last interglacial as defined by biostratigraphy (Brauer et al., 2007). Both MON 1 and MON 2 are characterised by an increase in varve thickness and the occurrence of an annual authigenic siderite layer (represented by higher Fe/Ti ratios, Fig. 4). However, MON 1 and MON 2 are geochemically different. The Fe/Ti ratio and Ti counts are considerably higher during MON 2, indicating detrital input (Ti) as the main control of varve thickness variability (Martin-Puertas et al., 2014). Ti counts do not show a noticeable increase during MON 1 and Fe/Ti ratios mirror the changes in varve thickness (Fig. 3). Siderite precipitation in both MON 1 and MON 2 is interpreted as a response to colder and drier climate (i.e. stadial-like conditions). Lower evapotranspiration under stadial-like conditions might increase the water budget around the lake, favouring Fe input by groundwater. Additionally, cooler conditions might favour water column stratification, i.e. anoxic conditions at the lake bottom, by either relative lake level rise or reduction of lake water circulation due to winter ice cover. The combination of Fe input and hypolimnic anoxia under cooler conditions could have triggered siderite precipitation (Ohlendorf et al., 2000). A likely explanation for higher detrital input during MON 2 could be the opening of the forest cover that characterised the Mélisey I stadial and could have led to increased soil erosion. In contrast, MON 1 was apparently not associated with a persistent
large-scale reduction in tree pollen abundance, presumably because this cooling episode occurred during still prevailing interglacial conditions when the forest remained dense. This is well expressed by the lack of a Ti increase even during the short-lived fluctuation in pollen discussed above (Fig. 3), showing that soil erosion did not increase during that period. This evidence supports the idea that the pollen fluctuation most likely represents only a temporary decrease in woody taxa pollen production rather than a catchment-scale shift in vegetation structure. In addition, the fact that MON 1 occurred about 1000 years before MON 2/Mélidey I, a stadial period that can be unequivocally linked to the Grand Pile pollen record, supports the conclusion that MON 1 reflects a different climatic oscillation than the WE, which occurred much earlier in Grand Pile (ca. 4000 years before Melisey 1). Examining the MON pollen record for the interval around 4000 years before the transition to Mélidey I reveals a minor single sample fluctuation dated to 113.70 kya BP (Allen & Huntley, 2009) that may correspond to the WE fluctuation at Grand Pile (Fig. 3). This possibility requires further investigation.

During the ca. 585-year interval between the end of MON 1 and the onset of MON 2, interglacial-type varves were again deposited, but these are slightly thicker than during full interglacial conditions (Fig. 3). We call this interval from 111.01 to 110.43 ± 0.69 ka BP ‘MON-rebound 1’ and assume that it reflects a moderate environmental improvement after the cool period that, however, did not reach full interglacial conditions. Apparently, the climate-proxy relationship was not stationary during the study interval. The interpretation of the sedimentological proxies reveals shifting varve facies and varve thickness variability as a response to changes in temperature and lake circulation prior to the transition to Mélidey I. But after that time varve proxies were more sensitive to soil erosion and land surface processes. The pollen response to climatic conditions also seems to be different depending on the time scales of the forcing mechanisms. Whereas pollen variability generally reflects changes in vegetation composition and structure, most often in response to climatic changes at centennial to millennial time scales (e.g. interstadial-stadal variability), at annual
to decadal time scales, pollen may reflect variations in pollen production of different taxa resulting from climatic variability (Huusko and Hicks, 2009). The fact that the observed short-term decrease in pollen productivity at 111.18 ± 0.69 ka BP coincided with the most persistent peak in siderite precipitation during that interval (Fig. 4) suggests that it was caused by cooler climatic conditions. This further shows that, at MON, lake circulation was more sensitive to low-amplitude climate change during the final stage of the LI than the composition of catchment vegetation. For this reason, we defined the boundaries of decadal to centennial climatic oscillations in the MON sediment record based on sediment proxies and not pollen.

4 Regional and hemispheric comparison: palaeoclimatic implications

The presence of the short-lived interval MON 1 at 111.44 ± 0.69 ka BP is the first signal of environmental change at the very end of the LI in MON that we assume to be triggered by the climatic deterioration. A detailed comparison of the MON record with the independently dated high-resolution BMS1 speleothem from Bue Marino Cave in Sardinia (Columbu et al., 2017), 550 km west of MON, provides further insights into the origin of the succession of a cool (MON 1) and a mild (MON-rebound 1) interval (Fig. 5). The speleothem BMS1 is a three millennia long record from ~113 – 110 ka with a growth hiatus at 111.76 ±0.43 / -0.45 ka. The BMS1 stable isotope data are interpreted as a proxy for precipitation and temperature with higher δ18O / higher δ13C values indicating cool-dry conditions (Columbu et al., 2017). Both records describe synchronous (within age uncertainties) sub-millennial-scale variability, which strongly support the inference that the MON lake system was very sensitive to regional climate change. Figure 5 also shows a recently published δ18O isotope record of a speleothem from Corchia Cave in Italy, 580 km northwest of MON, which is interpreted in terms of rainfall amount (Drysdale et al., 2007; Tzedakis et al., 2018). The cold-dry (MON 1) oscillation and the subsequent rebound, as well as the abrupt end of the LI (MON 2, yellow dashed line in Fig. 5) are not recorded in the Corchia Cave speleothem. Instead, the Corchia Cave isotope data seem
to respond to a long-term gradual trend toward drier conditions at the end of the LI, suggesting that either this proxy-record might not be sensitive to sub-millennial-scale climate variability at that time or that climate change was regionally different.

On a hemispheric scale from Northern to Southern Europe, Columbu et al. (2017) correlated the two rapid and short-lived climatic oscillations recorded in Bue Marino Cave with the Greenland sub-events GI-25b and GI-25a described by Rasmussen et al. (2014), based on the agreement between the independent dating of the isotopic records of Bue Marino Cave, the NGRIP ice core and the NALPS speleothem record (Boch et al., 2011). The onsets of GI-25b in Greenland and its counterparts in the Mediterranean date to 111.44 ± 2.78 ka b2k (111.39 ± 2.78 ka BP) in NGRIP, 111.44 ± 0.69 ka BP in MON, and 112.40 ± 0.52 / ±0.59 ka in BMS1. Within the inherent absolute age uncertainties associated with each chronology, one could assume a quasi-synchronous proxy response to climate change in these three records; however, possible discrepancies in the timing of the climatic response of less than ca. 3000 years (2.5% age uncertainty in NGRIP) cannot be excluded. In order to avoid the effects of the relatively large uncertainties in absolute ages, we apply an alternative approach for the hemispheric-scale comparison. Instead of absolute ages, we compare the duration of the climatic periods in the proxy records from MON, Bue Marino Cave and Greenland, assuming that uncertainties are lower. We consider this as a more suitable approach to identify possible regional differences between Greenland and the Mediterranean. The duration of GI-25b (500 ± 40 years), the cold-dry phase in BMS1 (480 ±730 /-380 years) and MON 1 (440 ± 9 years) are very similar; however, the following warm-wet phase in BMS1 (1080 ±850 /-630 years) and the ‘MON rebound-1’ (550 ± 35 years) are significantly longer than GI-25a (300 ± 40 years) (Fig. 5; Table 2). A shorter climate rebound in Greenland might suggest that the impact of the cooling at the
onset of GS-25 on the Mediterranean region was delayed by at least 250 years given minimum

difference between the duration of the rebound phase in Greenland (ca. 300 years) and MON

(ca. 550 years calculated in MON). However, we are aware that the sub-events within GI-25

in NGRIP are only defined by δ¹⁸O data and the estimated durations were considered as

preliminary (Rasmussen et al., 2014). Therefore, the speculation of a possible time lag needs

to be tested by other proxy records and improved dating.

From 111.4 ka BP (MON 1) back to 120 ka BP, our proxies do not show any significant

fluctuations in MON (Fig. 3) that might correlate with GS-26 at 119.14 b2k (Rasmussen et al.,

2014). A lower amplitude of this stadial could explain the lack of signal in the MON sediment

record as the interglacial environment around the lake was most likely too stable to respond

to it. Consequently, the short-lived cooling phase GI-25b recorded in NGRIP at 111.44 ± 2.8 ka

b2k was the first of a series of millennial- to centennial-scale high-latitude cold phases that

caused environmental responses in the MON record. We suggest this as the beginning of

teleconnection between climate variability at northern high latitudes and environmental

changes in the Mediterranean that persisted during the entire following glacial period (Allen

et al., 1999). This agrees with the evidence for the onset of the global climatic re-organisation

at GI-25a (Capron et al., 2012). In contrast to our finding, it has been reported that millennial-

scale variability persisted also during the last interglacial in the North Atlantic Ocean

(McManus et al., 2002; Wang and Mysak, 2002; Oppo et al., 2006) and in southern Europe

(Couchoud et al., 2009; Tzedakis et al., 2018). It has been therefore proposed that DO-type

variability during the last glacial period could have been a continuation of last interglacial

millennial-scale variability (e.g. Mokeddem et al., 2014 and Tzedakis et al., 2018). Since these

regional comparisons are generally based on proxy wiggle matching possible, regional

differences might have been overlooked. Nonetheless, the lack of signal in MON might be

explained by the location of Lake Monticchio in a sheltered location within a volcanic crater

with favourable local climate conditions. Thus, the amplitude of climatic oscillations during
the last interglacial might have been too low to affect the stable forested environment during
the last interglacial in southern Italy, or else as during the Holocene (Allen et al, 2002) these
oscillations may have acted as triggers for vegetation changes rather than driving fluctuations
in the vegetation. It further remains elusive why also the Greenland ice core records do not
show such a pronounced millennial-scale variability.

5 Conclusions
This study updates the palaeoclimatic and environmental interpretation of the sediment
record of Lago Grande di Monticchio, southern Italy, at the transition from the last interglacial
to the early glacial.
We re-define the short-lived cold phase MON 1 from 111.44 to 111.01 ± 0.69 ka BP. This
interval of 440 ± 9 years is characterised by increased varve thickness and formation of siderite
varves and includes a brief decline in tree pollen.
In contrast to the previous assumptions, we propose that the brief decline in tree pollen at
this point in the MON record does not correspond to the Woillard Event. Instead, we consider
this abrupt and short fluctuation from 111.177 to 111.150 ± 0.69 ka BP as an episode of low
tree pollen production during a more persistently cold episode within the cold phase MON 1.
Comparison of the high-resolution sediment proxies and pollen data from the record of Lago
Grande di Monticchio has enabled us to identify the onset of environmental response to rapid
climatic variability of the last glacial period in the Mediterranean. The first sign of climatic
instability in the final phase of the last interglacial appears at 111.44 ± 0.69 ka BP with the
onset of MON 1, when siderite varves start forming in response to cooler and drier conditions.
MON 1 coincides with the sub-event GI-25b at 111.44 ka b2k with a duration of 500 ± 40 years.
Subsequent climate and environmental change in the Mediterranean corresponded to
Greenland climate variability on millennial to centennial time scales. Before MON 1, no clear
proxy evidence for marked environmental fluctuations is found in the MON sediment record
of the Ll. North Atlantic marine records and a Mediterranean speleothem show, however, climatic oscillations also during the last interglacial.

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Table 1: P_Sequence parameters linked to the age model shown in Figure 2. Modelled ages are expressed as maximum (from) and minimum (to), confidence intervals (%) are 68%, and mean ages (μ) and errors (σ) are also shown.

<table>
<thead>
<tr>
<th>Name</th>
<th>Unmodelled (BP)</th>
<th>Modelled (BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Unmodelling</td>
<td>Modelled</td>
</tr>
<tr>
<td></td>
<td>from to</td>
<td>from to</td>
</tr>
<tr>
<td></td>
<td>% μ σ</td>
<td>% μ σ</td>
</tr>
<tr>
<td>Outlier Model</td>
<td>-167 171</td>
<td>-0.002315</td>
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<tr>
<td>General</td>
<td>68.2 1.29081</td>
<td>1.18295</td>
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<tr>
<td>T (5)</td>
<td>3.99E-17 68.2 2</td>
<td>5.32E-17 68.2 2</td>
</tr>
<tr>
<td>U (0.4)</td>
<td>1.1431</td>
<td>1.14351</td>
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<td>P_Sequence</td>
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<tr>
<td>Boundary</td>
<td>120308 119508</td>
<td>119903</td>
</tr>
<tr>
<td>Base varve</td>
<td>120288 119501</td>
<td>68.2 386</td>
</tr>
<tr>
<td>X-6 (40Ar/39Ar)</td>
<td>110419 108583</td>
<td>68.2 900</td>
</tr>
<tr>
<td>X-5 (40Ar/39Ar)</td>
<td>107327 104675</td>
<td>68.2 1300</td>
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<tr>
<td>Boundary</td>
<td>107379 105611</td>
<td>68.2 106496</td>
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</tbody>
</table>


**Table 2:** Absolute dating and duration of the climatic oscillations discussed in this study given by the independent chronologies associated with the sediment record of Lago Grande di Monticchio, the Bue Marino Cave speleothem BMS1 and NGRIP ice core (GICC05modelext).

<table>
<thead>
<tr>
<th>Climatic oscillation / event</th>
<th>Lago Grande di Monticchio Italy, Mediterranean</th>
<th>Bue Marino (Columbu et al., 2017) Sardinia, Mediterranean</th>
<th>NorthGRIP (Wolff et al., 2010; Rasmussen et al., 2014) Greenland</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Age [ka] and 2σ uncertainty</td>
<td>Duration [years]</td>
<td>Age [ka] and 2σ uncertainty</td>
</tr>
<tr>
<td>MON 1 / GI-25 b</td>
<td>111,446 ± 697</td>
<td>440 ± 9</td>
<td>112,480 ± 210/250</td>
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<tr>
<td>MON rebound 1 / GI-25 a</td>
<td>111,013 ± 697</td>
<td>550 ± 35</td>
<td>111,760 ± 210/240</td>
</tr>
</tbody>
</table>

**Figure captions:**

**Figure 1.** (a) Location of Lago Grande di Monticchio and other sites mentioned in the text. (b) Bathymetry of the lake and coring sites.
Figure 2. Age model for Lago Grande di Monticchio during the interval 120–106 ka BP based on varve counts and reported ages for the MON tephra layers X-5 and X-6. All data were incorporated into a Bayesian P.Sequence age model.
with a fixed $k$ value of 0.3 (the model rigidity constraint) and $Z$ defined as relative varve years.

**Figure 3.** Environmental and climate proxies from the MON sediment record versus age scale.

Microscope images of the varved sediments of MON (to left) showing interglacial and glacial varve types; (a) sedimentary units and microfacies; (b) varve thickness; (c) Fe/Ti ratio (log scale); (d) Ti counts; (e) pollen assemblage. Yellow shading indicates the interval MON 1
discussed in this study. The dashed line in the middle of the yellow shading shows the position of the pollen anomaly previously identified as the WE. This line also marks the onset of MON 1 published by Martin-Puertas et al. (2014), which has been extended in this study.

**Figure 4.** High-resolution pollen and sediment proxies records during MON 1. (a) Environmental proxies plotted against depth. From left to right: thin section images along the MON 1 oscillation (core M59-u, 17-52 cm); Fe/Ti ratio is plotted on the thin sections images showing an increase in this ratio when siderite varves (light sediments) occur; Woody taxa pollen percentages and woody taxa pollen influx. (b) Environmental proxies plotted against
age. From left to right: varve thickness (log scale); Fe/Ti ratio (log scale); and pollen
percentages of woody taxa. Blue points indicate pollen samples published previously in Brauer
et al. (2007) and Allen and Huntley (2009), the yellow point corresponds to the WE sample
discussed in this study. Red points are the new samples included in this study. Yellow shading
indicates the interval MON 1 and grey shading shows the pollen anomaly discussed in the text.
Figure 5. Regional and hemispheric-scale context. Comparison of the MON varve thickness record (centre, c) with the isotope records (δ¹⁸O and δ¹³C) of the Mediterranean speleothems from Corchia Cave (top, a) and Bue Marino Cave (top, b) and the NGRIP ice core δ¹⁸O record (bottom, d). Each record is plotted using its own chronology. Durations of the relevant intervals are shown.