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NALPS: a precisely dated European climate record 120–60 ka

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Abstract

Accurate and precise chronologies are essential in understanding the rapid and recurrent climate variations of the Last Glacial – known as Dansgaard-Oeschger (D-O) events – found in the Greenland ice cores and other climate archives. The existing chronological uncertainties during the Last Glacial, however, are still large. Radiometric age data and stable isotopic signals from speleothems are promising to improve the absolute chronology. We present a record of several precisely dated stalagmites from caves located at the northern rim of the Alps (NALPS), a region that favours comparison with the climate in Greenland. The record covers most of the interval from 120 to 60 ka at an average temporal resolution of 2 to 22 a and 2σ-age uncertainties of ca. 200 to 500 a. The rapid and large oxygen isotope shifts of 1 to 4.5‰ occurred within decades to centuries and strongly mimic the Greenland D-O pattern. Compared to the current Greenland ice-core timescale the NALPS record suggests overall younger ages of rapid warming and cooling transitions between 120 to 60 ka. In particular, there is a discrepancy in the duration of stadial 22 between the ice-core and the stalagmite chronology (ca. 3000 vs. 3650 a). The short-lived D-O events 18 and 18.1 are not recorded in NALPS, provoking questions with regard to the nature and the regional expression of these events. NALPS resolves recurrent short-lived climate changes within the cold Greenland stadial (GS) and warm interstadial (GI) successions, i.e. abrupt warming events preceding GI 21 and 23 (precursor-type events) and at the end of GI 21 and 25 (rebound-type events), as well as intermittent cooling events during GI 22 and 24. Such superimposed Last Glacial events have not been documented in Europe before.

1 Introduction

In the Greenland ice cores drastic climate changes are documented during the Last Glacial period. The rapid and recurrent variations – known as Dansgaard-Oeschger

(D-O) events (Dansgaard et al., 1993; Grootes et al., 1993) – are expressed as relatively warm and humid Greenland Interstadials (GI) and relatively cold and dry Greenland Stadials (GS; see Lowe et al., 2008 for an event stratigraphical recommendation). These successions had a global effect on climate (e.g., Clement and Peterson, 2008). In the Austrian Alps, precisely dated speleothems were shown to mimic the Last Glacial oxygen isotope signal from Greenland ice cores (Spötl and Mangini, 2002). Large and rapid air temperature changes occurred within decades to centuries. The amplitude was largest in the N-Atlantic realm and reached 8–16 °C during rapid warmings in Greenland (Severinghaus et al., 1998; Lang et al., 1999; Huber et al., 2006; Capron et al., 2010a). The millennial-scale changes also affected the concentration of greenhouse gases (Grachev et al., 2007; Loulergue et al., 2008), as well as global sea-level and high-latitude ice-sheets (Lambeck and Chappell, 2001; Arz et al., 2007). Recently, the occurrence of three types of short-lived, sub-millennial climate events was discussed (Capron et al., 2010a). These events occurred within D-O cycles and consist of abrupt warmings either preceding individual GI (named precursor events) or occurring toward the end of some of the GI (named rebound events), as well as of abrupt coolings, e.g. during GI 24 (Capron et al., 2010a).

The D-O variations show a quasi-periodic occurrence of ca. 1470 a (Bond et al., 1997; Grootes and Stuiver, 1997; Rahmstorf, 2003), although recent studies suggest that the recurrence interval is not robust and thus not significant (Peavoy and Franzke, 2010). Different triggers and mechanisms have been invoked to explain these rapid climate changes. Among these are freshwater influx into the North Atlantic (Clark et al., 2001; Arz et al., 2007), solar variations (e.g. Braun et al., 2005), internal oscillations (Broecker et al., 1990; Birchfield et al., 1994; Rahmstorf, 2002) and stochastic resonance (Alley et al., 2001; Claussen et al., 2003; Ditlevsen and Johnsen, 2010). Moreover, changes and different states of the Atlantic Meridional Overturning Circulation (AMOC) are discussed as internal forcings (Ahn and Brook, 2008; Schmittner and Galbraith, 2008), including latitudinal shifts in the location of North Atlantic Deep Water production (e.g. Ganopolski and Rahmstorf, 2002). Others claim a tropical trigger

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of the D-O variability (Clement and Cane, 1999; Clement and Peterson, 2008), or a rapidly changing windfield due to the dynamics of continental ice sheets (Wunsch, 2006). Ditlevsen and Johnsen (2010) recently reported evidence that internal noise triggered D-O warmings and concluded that these events cannot be predicted.

Accurate and precise chronologies are fundamental to improve our understanding of the enigmatic D-O pattern and its global teleconnections. The existing chronological uncertainties during the Last Glacial, however, are still significant (Svensson et al., 2008) and various timescales are available for Greenland ice cores alone (Meese et al., 1997; Johnsen et al., 2001; Svensson et al., 2008). The multi-parameter, annual layer-counted GICC05 timescale (Svensson et al., 2008) covers the past 60 ka and age uncertainties are on the order of 2.5 ka at 60 ka. In the first half of the Last Glacial (ca. 118–60 ka) the ss09sea timescale is regarded as the most reliable (NGRIP members, 2004). A comparison of the ice-core chronologies with those from other Last Glacial archives, e.g. N-Atlantic deep-marine sediments, suffers from the radiocarbon dating limit and uncertainties associated with the ^{14}C-calibration and the marine reservoir effect. U-Th-dated speleothem chronologies are promising and help to reduce the dating uncertainties substantially. Previous studies showed that speleothems indeed capture the D-O pattern (e.g., Wang et al., 2001; Spötl et al., 2006; Genty et al., 2003; Drysdale et al., 2007; Fleitmann et al., 2009; Asmerom et al., 2010). Valuable contributions come from Asian (e.g. Dykoski et al., 2005; Wang et al., 2008; Cheng et al., 2009) and Brazilian caves (Cruz et al., 2005; Wang et al., 2007).

In the Alps, which are known to be a climatically sensitive region (Casty et al., 2005; Auer et al., 2007), the speleothem O isotopic composition constitutes a climate proxy that allows for a direct comparison with the Greenland O isotope records. In particular, the Alps and Greenland share a dominant Atlantic influence and the common O isotopic signal allows comparison of the chronology of the ice-cores to that of radiometrically dated speleothems. In addition, the regional impact, similarities and differences can be evaluated.

In this study we present a record consisting of several precisely dated stalagmites from caves located at the northern rim of the European Alps (NALPS). This region is exposed to a strong Atlantic influence via northwesterly winds thus favouring a comparison with climate in Greenland. Moreover, the cave host rock favours deposition of speleothems with excellent geochemical dating characteristics. All Last Glacial samples exhibit sharp O isotope transitions highly reminiscent of the D-O pattern seen in the ice cores. Our record covers most of the first half of the Last Glacial period (118–64 ka) at high temporal resolution.

2 Cave sites and stalagmite samples

Four cave sites were selected for this study (Fig. 1). Beatus Cave is located in central Switzerland and samples were collected in a gallery at 875 m a.s.l. The cave air temperature is ca. 8 °C and mean annual precipitation is 1258 mm (2004–2008; meteorological station Interlaken, ca. 8 km from the cave). Baschg Cave is located in the most western part of Austria and the cave entrance is at 780 m a.s.l. The cave air temperature is ca. 10 °C and mean annual precipitation is 1231 mm (1971–2000; meteorological station Feldkirch, ca. 5 km from the cave). Klaus-Cramer and Schneckenloch Cave are located close to each other on the margin of the high-alpine Gottesacker karst plateau in western Austria (Fig. 1). Klaus-Cramer Cave is a shallow cave and the entrance opens at 1964 m a.s.l. The cave air temperature is only ca. 1–2 °C. The entrance of Schneckenloch Cave is at ca. 1270 m and the cave air temperature is ca. 6.5 °C. Mean annual precipitation at both sites is 1908 mm (1971–2000; station Schoppernau, 835 m a.s.l. and ca. 7 km from the caves). All four caves have small and well-defined catchments (a few km^2 only) and developed in the same carbonate host rock (Lower Cretaceous Schrattenkalk Formation). This host rock provides favourable geochemical conditions for U-Th dating, i.e. high U (0.5–2 ppm) and low detrital Th concentrations (typically 0.2–6 ppb ^{232}Th). Therefore, no significant correction of the U-Th ages is needed and the resulting ages are precise and accurate. Two stalagmites were recovered from

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Beatus, three from Baschg, one from Klaus-Cramer, and one from Schneckenloch Cave. Some samples were found broken while others were still in growth position. The stalagmites are typically small in size, i.e. between 11 and 38 cm high and near-equal in diameter along their growth axes. The samples consist of dense calcite and some show distinct lamination. Minor portions of some samples (typically near the base) consist of impure calcite due to clay and organic inclusions. These sections were not used in the combined record. Interestingly, speleothem growth locally also occurred during cold stadial conditions. This places tight constraints on the minimum temperatures of these alpine caves.

3 Methods

The stalagmites were cut in half and a 0.5–1 cm-thick slab was cut from the axial part and polished. Subsamples for radiometric U-Th dating (typically 0.05 to 0.1 g) were obtained using a dentist drill. After dissolving the powders in nitric acid and adding a mixed ^{229}Th-^{233}U-^{236}U spike, U and Th were separated from each other and from matrix elements using an ion-exchange resin in Teflon columns (procedure similar to Edwards et al., 1986). The isotopic composition of U and Th of the majority of the samples was analysed using a ThermoFinnigan NEPTUNE multi-collector inductively-coupled plasma mass spectrometer (MC-ICP-MS; Cheng et al., 2009). The remaining samples were analysed on a ThermoFinnigan ELEMENT single-collector ICP-MS (Shen et al., 2008). In both cases a spiked NBL-112A standard solution was measured before and after the sample runs and blank measurements were conducted to correct for the U and Th backgrounds. Isotopic activity ratios were calculated using the new decay constants of Cheng et al. (2008). The final ages were corrected for detrital Th using an initial ^{230}Th/^{232}Th activity ratio of 0.8 (cf. Richards and Dorale, 2003). Ages are quoted in “a BP” (years before 1950 AD; see Table 1 in Supplement). The U-Th-based age models (cf. Supplement Fig. S1) were calculated using the open-source R statistics software (version 2.10.0; R Development Core Team, 2009) and an algorithm

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optimised for speleothems (Scholz and Hoffmann, accepted). The age-depth function and the corresponding 95%-confidence intervals were calculated by superposition of ensembles of piecewise linear fits. In addition to the U-Th data points and corresponding errors the algorithm also uses stratigraphic information, i.e. the age of the speleothem must increase with increasing distance from top.

Stalagmite slabs and thin sections were investigated using transmitted-light, epi-fluorescence, as well as reflected-light microscopy. Subsamples for stable oxygen and carbon isotopic analysis were micromilled at 0.15 to 0.25 mm resolution along the central stalagmite growth axes. The isotopic compositions were analysed using a ThermoFisher Delta^{plus}XL isotope ratio mass spectrometer coupled to a ThermoFisher GasBench II. Results are reported relative to the VPDB standard and the precision of the δ^{18}O and δ^{13}C values is 0.08 and 0.06‰ (1-sigma), respectively (Spötl and Vennemann, 2003). In this paper, we focus on the palaeoclimatic application of the O isotopic signal.

4 The NALPS stalagmite record

The new O isotope record from the northern Alps covers the time interval from ca. 120 to 60 ,ka, i.e. D-O 25 to D-O 18. Interstadials dominate the record, reflecting favourable climate conditions with regard to speleothem growth. Speleothem formation, however, continued at least during some of the stadials, indicating that the caves were not frozen during these times. The record from seven stalagmites is temporally constrained by 154 U-Th ages (20–30 per stalagmite). Typical relative 2σ-uncertainties range from 0.2–0.6 %, i.e. average uncertainties associated with the timing and duration of rapid climate changes range from ca. 200 to 500 a (stalagmites KC1 210 a, BA1-clean section 410 a, BA1b 350 a, BA2 340 a, EXC3 450 a, EXC4 400 a, SCH7 530 a; cf. Fig. 2).

The NALPS stable isotope curves consist of ca. 8200 individual analyses and show rapid and large isotope shifts of up to 4.5‰ underscoring the high sensitivity of these cave sites. The average temporal resolution ranges from 2 to 22 a depending on the

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stalagmite and time interval. With regard to the transitions, a sharp (rapid) central portion is often flanked by more gradual progressions towards the isotopic maxima and minima. This could either be an expression of the regional climate, reflect hydrological processes in the karst aquifer, or could in part be a smoothing artifact of the applied age model. Regarding the climatic interpretation of the alpine speleothem O isotopic signal, the pattern is highly reminiscent of Greenland, i.e. high $\delta^{18}O$ values represent warm interstadials and low values cold stadials (Fig. 2). The speleothem O isotope values primarily reflect the O isotopic composition of regional meteoric precipitation and in the Alps this variable is strongly correlated with air temperature (Rozanski et al., 1992; Humer et al., 1995; Kaiser et al., 2001). Moreover, the temperature of carbonate precipitation in the cave is directly related to the outside air temperature and determines the O isotopic fractionation between drip water and speleothem calcite (Friedman and O'Neil, 1977; Kim and O'Neil, 1997; Lachniet, 2009). Therefore, the air temperature has a major, combined effect on the speleothem O isotopic composition. Seasonality, however, might also affect the isotope signal in the Alps (Humer et al., 1995), as the seasonal amplitude in $\delta^{18}O$ of meteoric precipitation is around 10‰ (IAEA/WMO, 2010). The distribution of precipitation during the warm and cold season might thus exert a major influence on the overall isotopic composition. This could partially explain the large shifts at the D-O transitions. Major changes in the source region of precipitation, e.g. Atlantic vs. Mediterranean (cf. Spötl et al., 2010), are considered negligible on the northern rim of the Alps, whereas changes in the trajectories of moisture across the European continent might have some influence (cf. Sodemann and Zubler, 2010). Replication of intervals in different stalagmite samples suggests neither cave-specific nor drip site-specific effects had a major influence on the O isotopic composition. The unsaturated zone above the studied caves ranges from ca. 10 to 200 m allowing for fast transmission of the meteoric O isotopic signal by the seepage water.

In general, the rapid transitions of 1 to 4.5‰ in $\delta^{18}O$ took place within decades (central shift) to several centuries (whole transition) in our samples. The central shift encompasses the abrupt and also largest part of the amplitude of a D-O transition. The

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whole transition also includes the less abrupt, i.e. more gradual flanks and the entire amplitude of a transition. In particular, the transition from Marine Isotope Stage (MIS) 5 to 4, i.e. the D-O 19 cooling, lasted 400 a in its central portion (2.8‰ shift) and 950 a for the whole transition (4.3‰; Fig. 3; stalagmite KC1). The D-O 20 warming lasted 110 a (2.6‰, stalagmite BA1b) and D-O 21 cooling 20 a (2.7‰, BA1b; the event, however, is not recorded completely). The central portion of the D-O 21 rapid warming (2.3‰; BA1) lasted for 60 a, while the whole transition took place within 550 a (4.0‰, BA1). The cooling of D-O 22 occurred within 60 a (2.2‰; stalagmite BA2) and the cooling of D-O 24 lasted 120 a in its central part (1.2‰) and 1050 a for the whole transition (3.3‰, EXC4). Stalagmite EXC3 recorded the warmings of D-O 23 and 24: the central shifts of 2.2‰ each occurred within 340 and 170 a, respectively, while the whole transitions of 3.6 and 3.5‰ lasted 1400 and 590 a, respectively. The minor D-O 25 warming (1.5‰) lasted for 20 a in our record (Fig. 3; stalagmite SCH7).

5 Chronological implications

Significant chronological implications arise for the current Greenland ice-core timescale. The MIS 5/4 transition, i.e. the D-O 19 cooling transition, is recorded in stalagmite KC1 in great detail (Fig. 2). D-O 19 started with a rapid increase towards positive O isotope values, followed by a gradual cooling over several centuries and a rapid drop towards negative values into MIS 4, which reflects the typical pattern associated with D-O fluctuations (e.g. Rahmstorf, 2002; Peavoy and Franzke, 2010). The GI 19 isotopic maximum occurred at 71 690 ± 220 a BP (average 2σ-error at the isotope maximum), compared to 72 900 a BP in the NGRIP record (ss09sea timescale; Fig. 2; Johnsen et al., 2001; NGRIP members, 2004). The two tie points (isotopic maxima) thus exhibit a difference of 1210 a, i.e. NALPS suggests a significantly younger age (Fig. 2). In the East Asian cave records (e.g. Wang et al., 2001; 2008) the MIS5/4 transition is not as pronounced as in the NALPS record. D-O transitions in the monsoonal records are generally more gradual and the cold/dry stadials are smoother than in alpine samples. Going further back in time, the rapid transition from GS 21 to GI 20 is constrained to 75 860 ± 300 a (mid-point of the transition; Fig. 2) compared to 77 100 a in the ice-core record, i.e. NALPS is younger by 1240 a. The mid-point of the transition from GI 21 to GS 21 is located at 77 580 ± 240 a in NALPS and at 78 500 a in NGRIP, i.e. a difference of 920 a.

Regarding the long GS 22, there is a mismatch between the ice core and the speleothem record (Fig. 2). The transition from GS 22 to GI 21 (mid-point) is constrained to 85 400 a in NGRIP and to 85 030 ± 410 a in stalagmite BA1, i.e. NALPS is only 370 a younger, which is within the error of the associated U-Th ages. The transition from GI 22 into GS 22, however, reveals an age of 88 300 a in NGRIP versus 88 690 ± 330 a in NALPS, i.e. this is the only transition in the first half of the Last Glacial, where NALPS suggests an older age than Greenland (390 a). The sharp transition into GS 22 is recorded in the two stalagmites BA2 and BA1, but is only constrained precisely in BA2 (Fig. 2). For that reason, the age model of BA1 in the overlapping section was adjusted to the age model of BA2 using the software AnalySeries (Paillard et al., 1996). Based on several precise U-Th ages, however, there is a distinct discrepancy in the duration of GS 22: it was approximately 3000 a long based on the current ice core chronology, while it lasted ca. 3650 a according to the NALPS record. A relatively long GS 22 is also supported by East Asian stalagmite records (e.g. Sanbao Cave; Wang et al., 2008).

The D-O transition from GS 24 to GI 23 is constrained to 104,700 a in NGRIP compared to 103 550 ± 375 a in NALPS, i.e. the latter is 1150 a younger (Fig. 2). The rapid cooling from GI 24 to GS 24 occurred at 106 000 a in the ice core and at 105 210 ± 450 a in stalagmite EXC4, i.e. a difference of 790 a. The latter transition is also recorded in stalagmite EXC3, although the isotope drop is less pronounced there. A stalagmite record from Corchia Cave, Italy (Drysdale et al., 2007) also covers the time interval from 118 to 96 ka. In this record, GS 24 appears rather long and the transition into GI 23 occurs somewhat late compared to NALPS, although it is within the 2σ-age uncertainty (0.8–1.0 ka; Drysdale et al., 2007). The progression of GI 23 is very similar

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in both speleothem records, i.e. showing a long-term cooling trend. The mid-point of the transition from GS 25 to GI 24 is constrained to 108 650 a in NGRIP and to 108 300 ± 450 a in NALPS; the small discrepancy of 350 a is within the errors of the U-Th age model. Similarly, the transition into GI 25 is constrained to 115 450 a and 115 320 ± 500 a in NGRIP and NALPS, respectively, i.e. a minor difference of 130 a in this oldest portion of the Last Glacial period. Compared to NALPS, the stalagmite record from Corchia Cave shows a more gradual and relatively early transition into GI 24 and also a more gradual shift into GI 25. The latter, however, is better developed in the Italian record. In Sanbao Cave (Wang et al., 2008) the maximum of GI 25 occurred relatively late and the transition is much more gradual.

Taken together, the NALPS record suggests overall younger ages for the rapid stadial and interstadial transitions compared to the current NGRIP timescale (ss09sea) between 120 and 60 ka. This observation is consistent with East Asian stalagmite records (Xia et al., 2007), as well as with the layer-counted GICC05 ice-core timescale (Svensson et al., 2008), which is also systematically younger than ss09sea in the time interval from 60 to 40 ka. Moreover, stalagmites from Kleegruben Cave in the Alps suggested a shift towards younger ages in the ss09sea timescale in the interval of GI 15 to 14 (Spötl et al., 2006). Based on our U-Th data and age models, this shift towards younger ages is relatively large (800–1300 a) in the interval 106 to 70 ka and small (a few hundred years) in the oldest part of the Last Glacial (118–106 ka).

6 Details recorded in NALPS

A comparison of the Greenland ice-core records with NALPS reveals a great deal of agreement but also some significant differences. Next to the difference in the duration of GS 22 in NGRIP and NALPS (see above), the short-lived D-O events 18 and 18.1 are not recorded in the O isotopic composition of the speleothem record (Fig. 2), although stalagmite KC1 is obviously sensitive to rapid climate changes (it recorded D-O

19) and the mean temporal resolution in the corresponding stalagmite section is high (11 a). In the NGRIP record the maximum amplitude of 4.5‰ of D-O 18 is located at 64 700 a and the interstadial lasted from 64 850 to 64 450 a (400 a). The maximum amplitude of 5.2‰ of D-O 18.1 is located at 70 250 a and this warm episode lasted 450 a (70 350–69 900 a). Interestingly, there is also a lack of clear evidence for D-O 18 in the East Asian monsoon records from Hulu and Sanbao Cave, although the temporal resolution is high enough to resolve the short interstadial during MIS 4 (Wang et al., 2001, 2008; Xia et al., 2007). With regard to D-O 18.1 there is evidence from new data of Hulu Cave (R. L. Edwards, unpubl. data). The observations in the Alps and East Asia provoke questions regarding the nature of some of these short-lived D-O interstadials, in particular with respect to their regional impact.

The NALPS record further resolves other short-lived details also found in the Greenland ice-core records, i.e. recurrent sub-millennial climate changes within the well-known D-O stadial and interstadial successions recently discussed by Capron et al. (2010a). The intermittent climate swings consist of short and abrupt warming events preceding GI 21 and 23 (termed precursor-type events) and of rapid warming events at the end of GI 21 and 25 (termed rebound-type events). Moreover, distinct transient cooling events are observed during GI 22 and 24. Such superimposed and rapid Last Glacial events have not been documented in Europe before. In the NGRIP record, a precursor-type climate event preceded GI 21 (Fig. 3) and shows a 2.2‰ variation in δ^{18}O of ice within 200 a (Capron et al., 2010a). The actual onset of GI 21 coincided with a 4.2‰ increase in δ^{18}O of ice, following a 100 a-return to cold conditions. NALPS provides evidence of two short-lived events preceding GI 21: the first one is centred at 85,360 a and lasted for 190 a (85 440–85 250 a). It is characterized by a rapid increase of 1.7‰ and a subsequent rapid decrease of 2.2‰ (Fig. 3). The maximum of the second event is centred at 84 990 a and its duration was 100 a (85070–84970 a). This event consists of a rapid 2.4‰ increase followed by a rapid 1.0‰ decrease and the final transition into GI 21. At the end of the gradual cooling interval of GI 21 the Greenland O isotope values increased by ca. 2‰ in less than 100 a, before they returned to stadial conditions (Capron et al., 2010a). In the NGRIP record this rebound-type event shows two major positive peaks centred at 79 700 and 78 650 a (Fig. 2). The latter, shorter event directly preceding GS 21 is also recorded in NALPS: the δ^{18}O maximum occurred at 77 590 a and this distinct rebound of 1.2‰ lasted for 150 a (77 730–77 580 a), followed by a rapid transition into the stadial. Capron et al. (2010a) noted that this rebound pattern is similar to GI 22 with regard to its δ^{18}O magnitude, duration and structure. In spite of this, however, it is not counted as a GI. GI 22 is characterized by two distinct cooling events centred at 88 950 and 89 610 a (recorded in stalagmite BA2). The O isotope values show decreases of 2.1 and 1.8‰ and the anomalies lasted for 250 a (89 130–88 880 a) and 210 a (89 700–89 490 a), respectively. The younger of the two events is also recorded in stalagmite BA1, although the associated U-Th age errors are significantly larger. There is also evidence for these cooling anomalies during GI 22 in the NGRIP record (Fig. 2).

Another precursor-type structure is evident immediately prior to GI 23 in Greenland: δ^{18}O increased by 3.8‰ within 130 a and subsequently dropped by 3.6‰ in 100 a. The latter transition back to stadial conditions lasted for 300 a before the δ^{18}O values increased again by 3‰ at the onset of GI 23 (cf. Capron et al., 2010a). In the NALPS record this precursor event is centred at 104 050 a (Fig. 3) and is characterized by a positive shift of 1.2‰ followed by a negative shift of 0.8‰ in δ^{18}O and lasted for about 400 a (104 170–103 770 a). Going further back in the Last Glacial period, GI 24 shows two distinct negative (cooling) peaks of 100 to 200 a duration in NGRIP. Based on the ss09sea timescale the two minima occurred at 107 300 a and 106 800 a, respectively. NALPS shows the first, more pronounced minimum in two stalagmites and the event is constrained to 107 470 a based on the age model of stalagmite EXC3 (higher resolution and better chronology in this section than sample EXC4). The negative O isotope anomaly of 0.7‰ magnitude lasted for 380 a (107 620–107 240 a) and its progression is characterized by a rapid decrease and more gradual increase in both stalagmites. Capron et al. (2010a) reported that no other interstadial is interrupted by comparable, short, cold events. In the NALPS record, however, two stalagmites document a distinct negative peak during GI 22 (see above). In this context the question whether GI 22 should be considered an interstadial or (only) a rebound-type structure following the long GI 23 (cf. Capron et al., 2010a) can be raised again. At the end of GI 25 there is evidence of a rebound-type event in the NALPS record comparable to NGRIP (cf. Capron et al., 2010a). The event lasted for ca. 180 a (111 780–111 600 a) and the maximum positive shift of 0.7‰ occurred at 111 660 a based on NALPS (Fig. 3).

Regarding the secondary, sub-millennial events a relationship with the summer insolation at 65° N in connection with variable ice-sheet extensions is discussed (Capron et al., 2010a). The rebound-type events are typically associated with relatively low summer insolation at the end of particularly long cooling phases during the GI progression. In contrast, the precursor-type events might be linked to insolation maxima; Capron et al. (2010a) reported an in-phase relationship of the GI 21 precursor event with a relative maximum in 65° N summer insolation and a delay of ca. 2.5 ka of the GI 23 precursor relative to the preceding insolation maximum (after Laskar et al., 2004). Based on the NALPS chronology the precursor of GI 23 (maximum at 104 050 a; cf. Supplement Fig. S2) is delayed by ca. 1 ka when compared to the adjacent insolation maximum calculated by Laskar et al. (2004). The two precursor events of GI 21 recorded in NALPS (at 85 360 and 84 990 a; Fig. 3) preceded the insolation maximum by ca. 1 to 1.5 ka (Fig. S2). Using the NALPS chronology in relation to the summer insolation after Laskar et al. (2004) therefore suggests a shorter delay (ca. 1 ka) of the GI 23 precursor event as compared to Capron et al. (2010a). Our data support a preferential occurrence of the precursor events at times of maximum Northern Hemisphere summer insolation during the Last Glacial.

Supplementary material related to this article is available online at: http://www.clim-past-discuss.net/7/1049/2011/cpd-7-1049-2011-supplement.zip.

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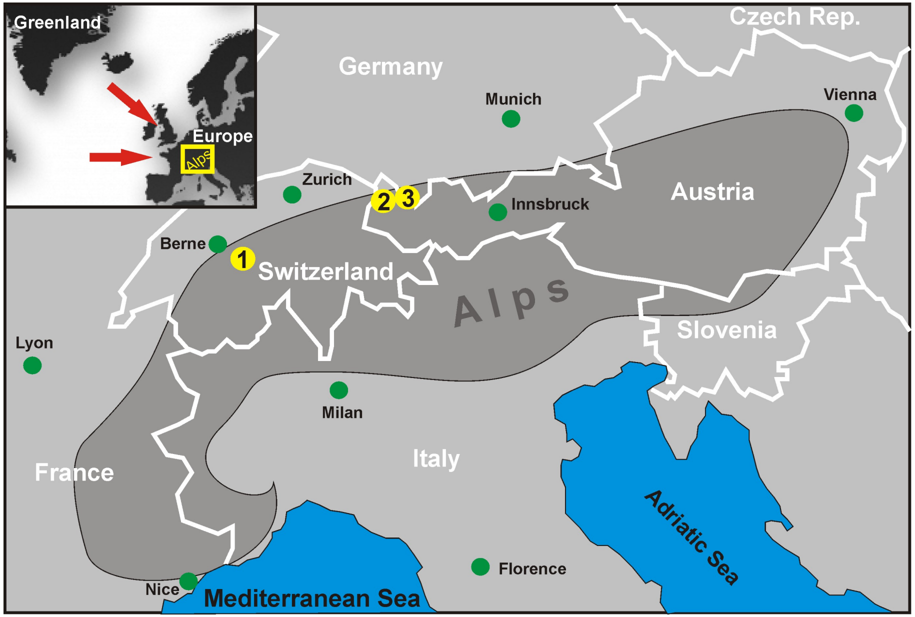

Fig. 1. Location of the selected cave sites at the northern rim of the Alps: (1) Beatus Cave (Switzerland); (2) Baschg Cave (Austria); (3) Klaus-Cramer- and Schneckenloch Cave (Austria). The northern rim of the Alps is dominated by moisture advection from the Atlantic Ocean (red arrows).

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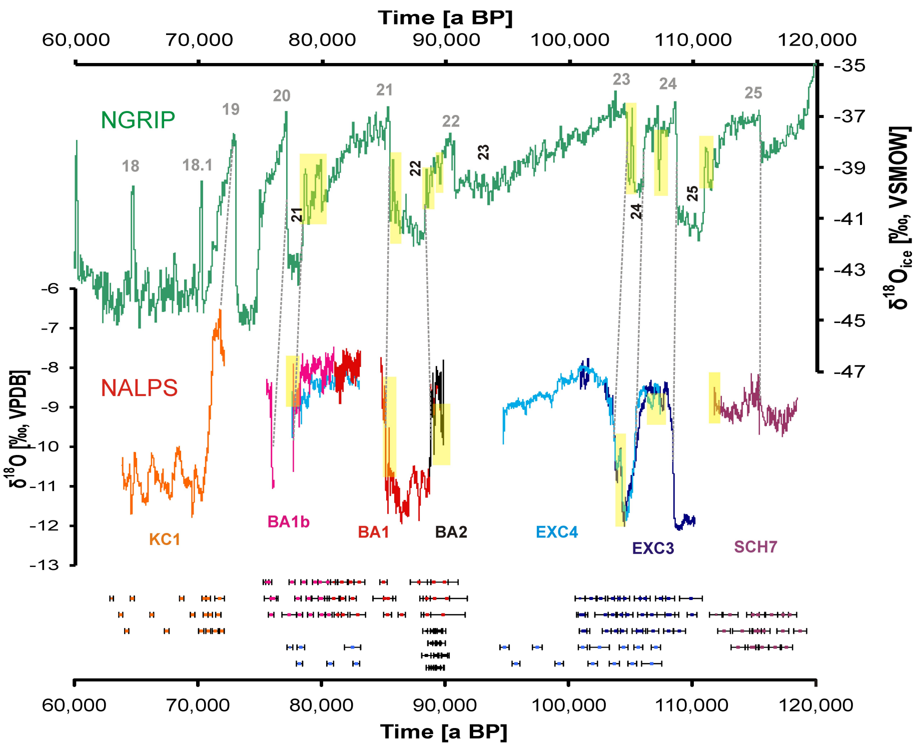

Fig. 2. The NALPS record (consisting of seven stalagmites) compared to NGRIP plotted on the ss09sea timescale in the interval from 120 to 60 ka. Individual U-Th ages and associated 2σ-uncertainties are plotted at the bottom. The grey, dashed lines connect the mid-points of major D-O transitions or isotopic maxima. The recurrent warm Greenland Interstadials of the D-O cycles are indicated by horizontal grey numbers, while cold Greenland Stadials are indicated by vertical black numbers. Short-lived warming and cooling events (sub-D-O events) are highlighted by the yellow rectangles. NALPS suggests overall younger ages of rapid stadial and interstadial transitions and a longer duration of stadial 22.

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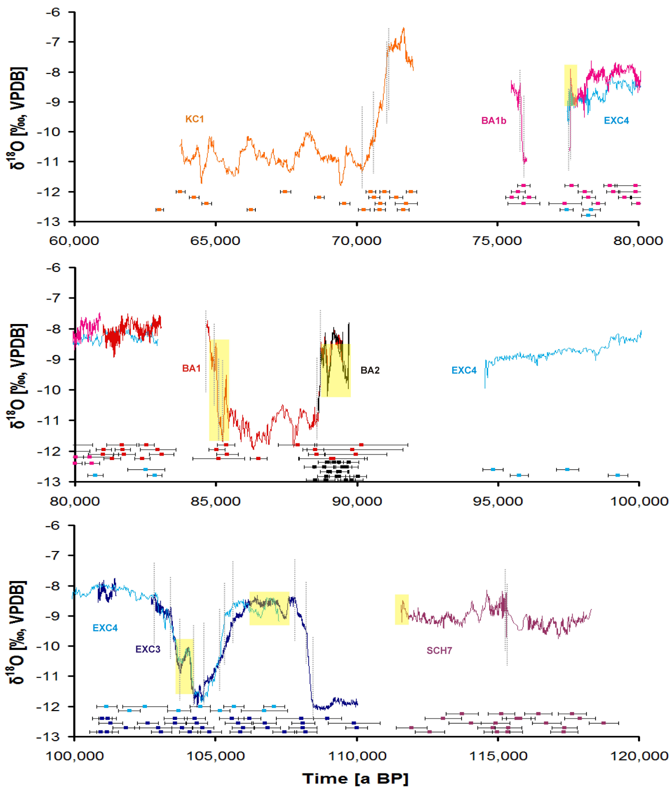

Fig. 3. Detailed structure of the NALPS record. The timing of central and whole D-O transitions (see text) is indicated by the grey, dotted lines. U-Th age data with 2σ-error bars are plotted at the bottom of the diagrams and the yellow rectangles highlight sub-D-O warming and cooling events.

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