

Vegetation responses to climatic variability in the Swiss Southern Alps during the Misox event at the early–mid Holocene transition

VERUSHKA VALSECCHI* and WILLY TINNER

Institute of Plant Sciences and Oeschger Centre for Climate Change Research, University of Bern, Bern, Switzerland

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ABSTRACT: Sedimentary pollen, charcoal and plant macrofossil analyses with high resolution and precision suggest a strong shift in vegetation composition during the early to mid-Holocene transition in the upper mountain belt. At Piano mire (1439 m above sea level (a.s.l.), Ticino, Switzerland) forests were dominated by *Abies alba* during the early Holocene (prior to ca. 8000 cal. a BP). Abrupt collapses of *A. alba* at ca. 7800–7400 cal. a BP enabled the expansion of the light-demanding pioneer *Betula*. Afterwards *A. alba* populations regained their previous abundance in the forests. Within the dating uncertainties of our record we assume that a unique combination of wet and cold years between 8400 and 7500 cal. a BP led to repeated lethal disadvantages for *Abies*. Our record of *Abies* oscillations is in good biostratigraphic agreement with the record that has been used to define the Misox cold event (Pian di Signano, 1540 m a.s.l.), which has been previously correlated with the 8200 cal. a BP event. Given the age estimates of the *Abies* collapses in our well-dated record, our results suggest that additional efforts are needed to understand the linkage between the Misox and the 8200 cal. a BP event. They imply a high sensitivity of mountain vegetation far below the tree line (~800 m) to Holocene climatic changes of about 2°C in annual air temperature. Copyright © 2010 John Wiley & Sons, Ltd.



KEYWORDS: pollen; plant macrofossils; 8200 cal. a BP event; Misox; Alps.

Introduction

The Holocene is a period with relatively stable climatic conditions if compared to glacial–interglacial variability. However, Holocene climatic changes of minor amplitude are documented in several palaeoclimatic records (Mayewski *et al.*, 2004; Wanner *et al.*, 2008). Climatic forcing can operate at low (millennial), intermediate (centennial) and/or high (decadal to annual) scales. Holocene climatic changes may have been triggered by external forcing (e.g. orbital parameters, solar activity) or by internal forcing (e.g. weakening of the North Atlantic thermohaline circulation, volcanic eruptions) or result from changes in internal short-term variability mode changes (e.g. North Atlantic Oscillation, El Niño; Bradley, 2003). Many studies show that climatic forcing at different scales plays an important role in shaping Holocene environmental variability. Short-term climatic changes at annual to decadal scales might result in displacement of ecotones, whereas long-term changes (at centennial to millennial scales) might have profound consequences on the structure of biotic

communities and entire biomes (Webb, 1986; Delcourt and Delcourt, 1991). Wide climate changes may occur very rapidly when the climate system is forced to cross thresholds, and such abrupt changes are particularly harmful for long-lived and relatively immobile ecosystems (i.e. mature forests; Alley *et al.*, 2003). In agreement, high-resolution and accurate studies unequivocally demonstrate fast biotic responses to abrupt Lateglacial and Holocene climatic changes (e.g., Birks and Ammann, 2000; Tinner and Lotter, 2001; Williams *et al.*, 2002; Hu *et al.*, 2003).

During the Holocene, climatic variability in and around the Alps probably reached its maximum at ca. 8400–7400 cal. a BP (Wick and Tinner, 1997; Haas *et al.*, 1998; von Grafenstein *et al.*, 1998; Heiri *et al.*, 2004; Spötl *et al.*, 2010), when climate shifted from the early Holocene dry continental to the mid Holocene moist oceanic mode (Tinner and Lotter, 2001, 2006). This reorganisation of the European climatic system corresponds to the Boreal/Atlantic transition in the early palaeoecological European literature (e.g. Firbas, 1949/1952) and thus has been a topic of discussion for more than 50 years. During this transitional period annual air temperature experienced strong oscillations (reaching about 2°C in magnitude; von Grafenstein *et al.*, 1998) and the precipitation regime was significantly altered (e.g. Magny *et al.*, 2003; Spötl *et al.*, 2010). In the Swiss Southern Alps cold phases were inferred from the evidence of three strong minima of *Abies alba* pollen

*Correspondence to: V. Valsecchi, Centre de Bio-Archéologie et d'Ecologie (UMR 5059), University of Montpellier 2, Institut de Botanique, 163 rue Auguste Broussonet, F-34090, Montpellier, France.
E-mail: verushka.valsecchi@univ-montp2.fr

percentages at Pian di Signano by Zoller (1960) that defined the term 'Misox oscillations'. Misox (or with the Italian place name Mesolcina) is the valley where repeated collapses of *Abies* were observed (Zoller, 1960). The *Abies* oscillations defining the Misox event (sometimes erroneously called 'Miser oscillation'; see, for example, Alley and Ágústsdóttir, 2005) were radiocarbon-dated to ca. 7500–6500 ¹⁴C a BP (Zoller, 1960). Almost 40 years later the event was correlated with the cold event CE-3 comprising tree line and glacier oscillations in Central Europe (Haas *et al.*, 1998) and with the 8200 cal. a BP event in the Greenland ice cores (Wick and Tinner, 1997).

The main aim of this study is to refine the reconstruction of Holocene vegetation dynamics in the upper mountain belt of the Swiss Southern Alps. We analyse a new site to check whether the *Abies* collapses, which define the Misox event, can be reproduced in space and time. The new sediment record was analysed for the Piano mire (1439 m above sea level (a.s.l.), Ticino, Switzerland), about 50 km from Zoller's (1960) site at a similar altitude. We give special attention to the Misox event period, and to gain higher spatial resolution of the vegetation shift we combine pollen and macrofossil analyses (Birks, 2001). In addition, sedimentary microscopic charcoal is used to infer regional fire activity (Whitlock and Larsen, 2001; Conedera *et al.*, 2009).

STUDY SITE

The Piano mire (46° 19' 15" N, 8° 37' 12" E; 1439 m a.s.l.) is located in the Swiss Southern Alps in Val Maggia (Ticino, Switzerland) northwest of Lago Maggiore (Fig. 1). The small mire has a surface area of 0.2 ha, with a minor northern outlet that flows into the Valle del Piano and the River Maggia. It lies in a depression and is surrounded by mountains of crystalline composition (granite gneiss – Antigorio and Levantina type) and by roches moutonnées (Preiswerk *et al.*, 1934). By spatial interpolation from nearby climate stations, present-day annual precipitation is estimated to 2500 mm, with a maximum during spring and autumn (Sevruck and Kirchhofer, 1991). Mean annual and mean temperatures of January and July are 6.0, –1.1 and 15.1°C, respectively (Z'graggen, 2001). Piano is located at the

transition between the mountain belt, which is dominated by *Fagus sylvatica* and *Abies alba*, and the subalpine belt, with *Picea abies* and *Larix decidua* forests. The mire is dominated by *Molinia coerulea*, whereas *Carex* sp. and *Sphagnum* sp. are less abundant. The slopes surrounding the mire are covered by mixed forests of *A. alba* (with *Fagus sylvatica* and *Picea abies*). At higher altitudes (ca. 1600 m a.s.l.), *F. sylvatica* disappears and the forest is composed of *A. alba*, *P. abies*, *L. decidua* and *Betula pendula*. The timber line in the area is situated at ~1800 m a.s.l., while the tree line reaches about 2200 m a.s.l. (Landolt, 1992).

MATERIAL AND METHODS

Two parallel cores (PIA 1 and PIA 2) 360 and 375 cm long were retrieved from the centre of the mire with a Streif-modified Livingstone piston corer of 4.8 cm diameter (Merkt and Streif, 1970). After having compared the lithostratigraphy of the two cores, we used the core section from 160 to 352 cm of PIA 1 and from 352 to 375 cm of PIA 2 for this study. The upper part of the core (0–160 cm) is shown elsewhere (Valsecchi *et al.*, 2010).

Radiocarbon dating

The chronology is based on nine accelerator mass spectrometry (AMS) radiocarbon dates carried out at Poznan Radiocarbon Laboratory. Sediment slices 1 cm thick were sieved at 200 µm mesh and only terrestrial plant macrofossils were used for dating (Table 1). The resulting radiocarbon dates were converted to calendar years BP with the CALIB rev. 5 program (Stuiver and Reimer, 1993; Stuiver *et al.*, 1998). The age–depth model was based on non-parametric weighted regression within the framework of generalised additive models. The resulting age–depth model and associated age estimates and 95% confidence intervals are based on the combined uncertainty of the calibrated dates and the regression line (Birks and Heegaard, 2003; Heegaard *et al.*, 2005).

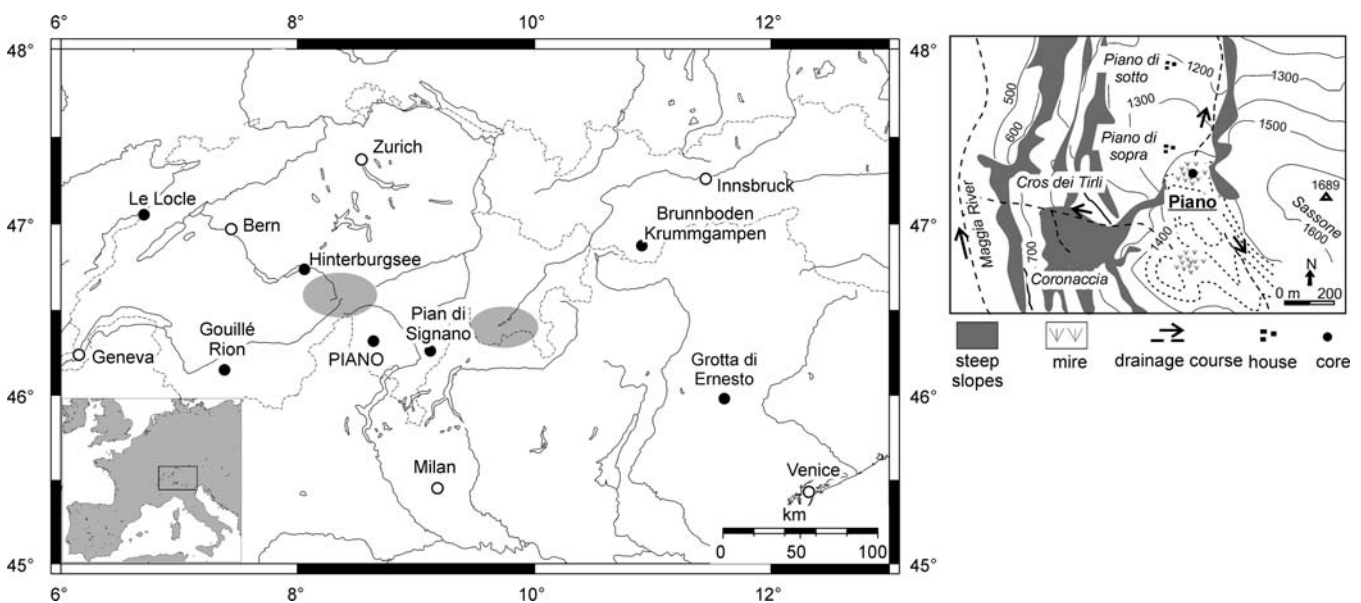


Figure 1 Left: map showing the locations discussed in this paper: Piano mire (this study), Pian di Signano (Zoller, 1960), Gouillé Rion (Tinner *et al.*, 1996; Heiri *et al.*, 2004), Brunnboden and Krummgampen (Kofler *et al.*, 2005); Hinterburgsee (Heiri *et al.*, 2004), Grotta di Ernesto (Spötl *et al.*, 2010) and Le Locle (Magny and Schoellammer, 1999); grey areas indicate the location of the investigated glaciers (Joerin *et al.*, 2006, 2008). Right: map of Piano mire and surroundings

Table 1 AMS radiocarbon dates on Piano core section

Laboratory reference	Depth (cm)	Material dated	¹⁴ C a BP	Calendar age (cal. a BP, 2 SD)
Poz-2034	160	<i>Abies alba</i> seed	6170 ± 40	6 950–7 150
Poz-10668	174	<i>Sphagnum</i> sp. stems	6400 ± 40	7 250–7 400
Poz-23352	200	<i>Abies alba</i> seed	6900 ± 40	7 650–7 800
Poz-10670	210	<i>Abies alba</i> needles	7120 ± 40	7 850–8 000
Poz-14769	232	Conifer cone scale	7950 ± 50	8 650–9 000
Poz-2035	252	<i>Larix decidua</i> needles	8490 ± 50	9 450–9 550
Poz-2036	328	Periderm fragments	9290 ± 50	10 300–10 650
Poz-2037	352	<i>Larix decidua</i> cone scale	9680 ± 50	10 800–11 200
Poz-2038	364	<i>Larix decidua</i> needles	9850 ± 60	11 150–11 400

Palaeoecological analysis

Samples of 1 cm³ were used for pollen analysis, with *Lycopodium* tablets added for estimation of pollen concentration (Stockmarr, 1971). The peat samples were prepared with the following procedure: KOH, 500 µm sieving, decanting, acetolysis and KOH. Pollen grains and spores were counted at a magnification of 400× and 1000× and identified using keys, pollen atlases (Punt, 1976; Punt and Clarke, 1980, 1981, 1984; Punt *et al.*, 1988, 1996; Moore *et al.*, 1991; Punt and Blackmore, 1991; Reille, 1992–1998) and the reference collection of the Institute of Plant Sciences of Bern. At least 350 pollen grains per sample were counted, excluding aquatic plants, wetland herbs, ferns and extrafossils. Charcoal particles longer than 10 µm were identified with a light microscope at ×200 magnification on the pollen slides (Conedera *et al.*, 2009). Charcoal was identified as black, completely opaque, angular fragments (Clark, 1988; Tinner *et al.*, 1999). Following Tinner and Hu (2003) and Finsinger and Tinner (2005), at least 200 objects (charcoal fragments and *Lycopodium* spores) were counted for each sample to estimated microscopic charcoal concentrations (particles cm⁻³). Non-pollen palynomorphs were identified in the pollen slides using keys and reference plates (Beyens and Meisterfeld, 2001; Komarek and Jankovska, 2001; van Geel, 2001). The sediment was sampled continuously in 4 cm sediment slices between 160 and 240 cm depth for plant macrofossil analysis, resulting in a time resolution of 52–160 cal. a per sample (see 'Results and interpretation'). Samples were sieved with tap water through sieves at 200 µm mesh. Macrofossils > 200 µm (including macroscopic charcoal) were analysed with a stereomicroscope at 10–40× magnification and identified with the seed reference collection of the Institute of Plant Sciences in Bern and plant macrofossil atlases (Katz *et al.*, 1965; Schoch *et al.*, 1988). Concentrations in the plant macrofossils diagram are adjusted to a volume of 25 cm³. The program TILIA was used for plotting the pollen and macrofossil diagrams (Grimm, 1993). The pollen record was zoned with the program ZONE 1.2 and optimal partition with the sum-of-squares criterion was selected (Birks and Gordon, 1985). The broken-stick model (Bennett, 1996) defined eight significant pollen zones (LPAZ), whereas the plant macrofossil data were zoned visually into three macrofossil assemblage zones (LMAZ).

Data analysis

Cross-correlations help to identify relations between two different time series at any time step of interest, though this procedure is limited to lags $k \leq n/4$ (Bahrenberg *et al.*, 1992). Correlations for negative lags indicate the influence of the dependent variable (Y) on the independent variable (X) in regard to the temporal lag (k), while positive lags are a measure for the

influence of the independent on the dependent variable (Dodson, 1990; Bahrenberg *et al.*, 1992; Tinner *et al.*, 1999). Here we apply cross-correlations between microscopic charcoal influx (X) and pollen percentages (Y) to analyse the influence of fires on vegetation succession. The significance levels (0.05) of the correlation coefficients are given in the correlograms. Continuous sampling was applied in order to have time intervals as close to equal as possible between adjacent samples (19 ± 4 a, mean ± standard deviation) and to obtain complete charcoal records – a prerequisite for this kind of analysis (Tinner *et al.*, 1999). Cross-correlations were calculated for the period of particular interest (8050–7350 cal. a BP; 175–212 cm).

RESULTS AND INTERPRETATION

Lithostratigraphy and age model

The sediment consists of clay and silty gyttja from 370 to 360 cm (colour dark grey, Munsell, 1954), followed by dark-brown gyttja up to 302 cm. Then the peat content increases towards 253 cm. Finally, between 253 and 170 cm, the material is composed of sedge–*Sphagnum* peat and the upper sediment consists of herbaceous–sedge peat (170–160 cm; Fig. 2). Nine radiocarbon dates (Table 1) were used to build the depth–age model, which provides a calibrated age of ca. 11 280 ± 65 cal. a BP at the base of the sequence studied and of ca. 7080 ± 75 cal. a BP at the top (Fig. 2). At the bottom the sedimentation rate is ca. 0.48 mm a⁻¹ (21 cal. a cm⁻¹), increasing to a maximum of 1.3 mm a⁻¹ at 290 cm (transition gyttja to peaty gyttja). In the middle (from ca. 260 to 200 cm – transition to peat) the sedimentation rate decreases, reaching the lowest values of 0.32 mm a⁻¹

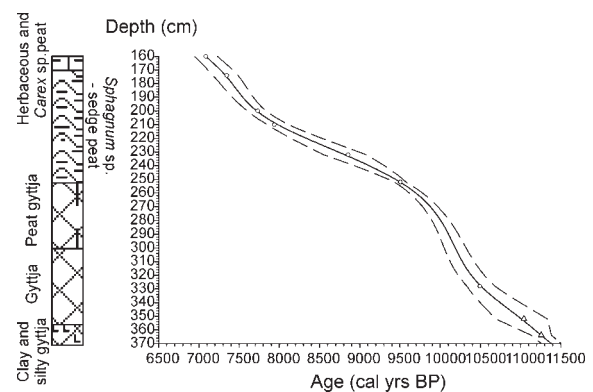


Figure 2 Lithostratigraphic column of cores PIA 1 and PIA 2. Left: depth–age model (line) and 95.4% confidence intervals (dashed lines). Calibrated radiocarbon dates are represented by dots (core PIA 1) and triangles (core PIA 2)

(33 ± 9 cal. a cm^{-1}). The highest sediment accumulation rate (average 0.62 mm a^{-1}) occurs in the upper part (with herbaceous-sedge peat).

Vegetation and fire dynamics inferred from pollen, plant macrofossils and microcharcoal

Pinus–Betula forests, 11 300–9500 cal. a BP

The basal pollen zone is characterised by *Pinus sylvestris* type (35–45%), *Betula* (20–40%) and *Pinus cembra* (around 15%). Scattered *Larix* and *Juniperus* pollen were also found. Non-arboreal pollen types have a low percentage from ~20% to ~5% and are represented by *Artemisia*, *Ranunculus acris* type, *Rumex acetosa/acetosella* type and Apiaceae (Fig. 3). The local presence of *Pinus sylvestris* (*P. mugo* is not present in the study area), *Larix decidua* and *Juniperus* are documented by the presence of stomata and by *Pinus sylvestris* and *Larix decidua* needles (found in the radiocarbon samples, data not shown). Our palaeobotanical record thus suggests the presence of *Pinus sylvestris*–*P. cembra*–*Betula* forests around the former lake, with a scattered presence of *Larix decidua* and *Juniperus*. The low pollen values of *Quercus*, *Tilia*, *Ulmus* and *Corylus* are best explained by long-distance transport from lower altitudes, but their local presence cannot be excluded. Similar early Holocene vegetation compositions had been found at a site located in the mountain belt in the Ticino Valley (Zoller, 1960; Zoller and Kleiber, 1971; Müller, 1972). Our site was then a shallow lake (gyttja, peaty gyttja, with *Pediastrum boryanum* var. *boryanum* and var. *longicornis*, *P. angulosum*, which are algae that live in small peaty lakes; Fig. 3). Pollen concentration is quite high and microscopic charcoal concentration and influx follow the same pattern, suggesting that regional fires were intense (Fig. 3). Pollen concentrations decrease around 9300 cal. a BP and charcoal influx markedly declines around 9500 cal. a BP, almost in correspondence with a change in the sediment from peat gyttja to peat.

Abies alba–Betula forests, 9500–7800 cal. a BP

Around 9500 cal. a BP, *Abies* pollen increased markedly. Pollen percentages of *Pinus sylvestris* type and *Betula* show a gentle decrease, whereas the values of upland herbs are stable (Fig. 3). *Abies alba* and *Larix decidua* plant macrofossils are present from 9100 cal. a BP (PIA-M 1), attesting to the local presence of these taxa (Fig. 4). Our data point to the presence of a new type of forest, in which rather drought-sensitive and shade-tolerant *Abies alba* was dominant, whereas *Betula*, a pioneer light-demanding species, became less important. Nevertheless, forests were still rather open, as shown by the local presence of *Larix* and *Juniperus*. Given the lack of macrofossils of *Pinus sylvestris* and *P. cembra* (Fig. 4) we assume that most pollen of these species was coming from extra-local sources. The change in the vegetation goes along with a change in the aquatic ecosystem: the beginning of the pollen zone (LPAZ PIA-2) delimits the beginning of the conversion of the lake into mire. This is shown by the sediment change towards sedge–*Sphagnum* peat, marked by a pronounced increase of the testate amoebae species *Amphitrema flavum* (a sphagnum-bog species; Beyens and Meisterfeld, 2001) and *Assulina*, both indicating oligotrophic conditions (van Geel, 1978). Charcoal concentration and influx remain rather stable and lower than in the previous zone (Fig. 3), suggesting low regional fire activity when *Abies alba* dominated the vegetation.

Rapid and repeated reciprocal replacements of *Abies alba* and *Betula* forests, ca. 7800–7400 cal. a BP

Pollen and macrofossil data suggest fast vegetational changes with mutual replacements of *Abies alba* (and to lesser degree *Larix*) and *Betula* during this period. At 7800 cal. a BP (± 120), percentage values of *A. alba* and *Pinus sylvestris* type start to decline, while *Betula* strongly increases, reaching values of up to 80% (Fig. 5). The plant macrofossil diagram shows a similar

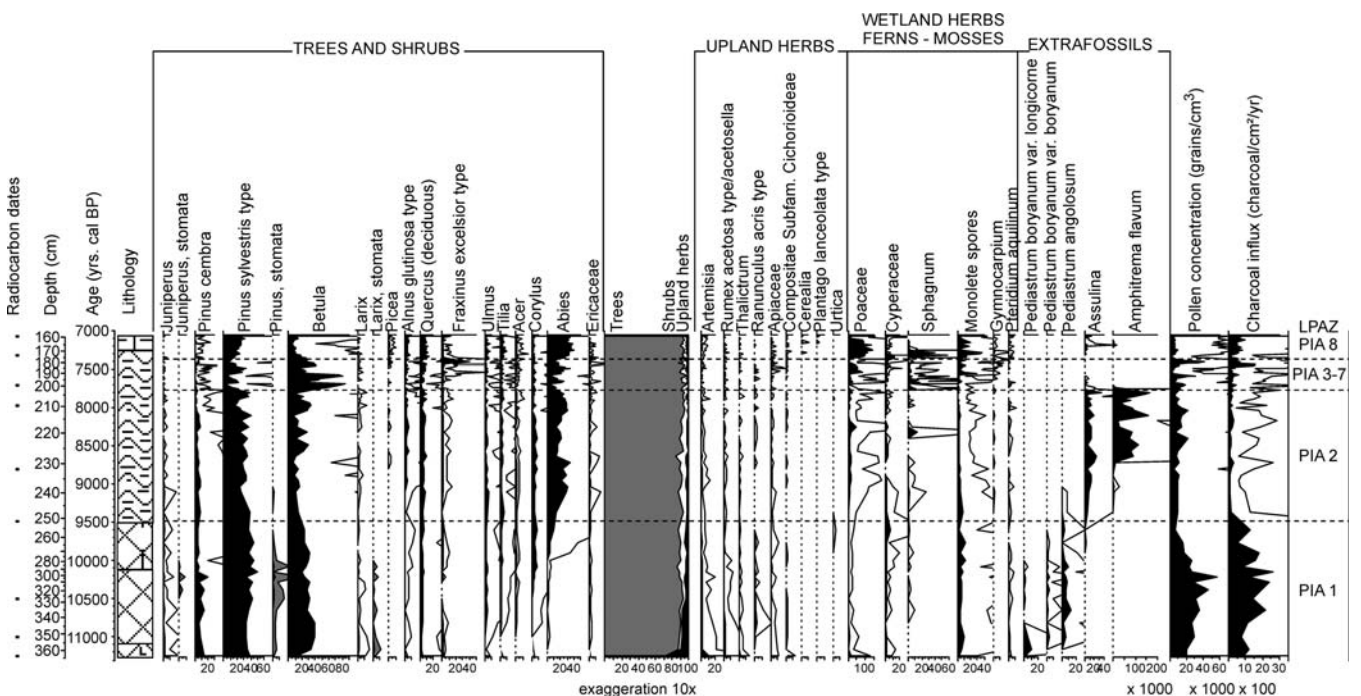


Figure 3 Pollen percentages and charcoal concentration and influx for Piano mire showing the most important taxa. Wetland herbs, aquatics, fern, mosses and extrafossils are not included in the pollen sum. On the left the lithostratigraphy is shown (see Fig. 2 for legend)

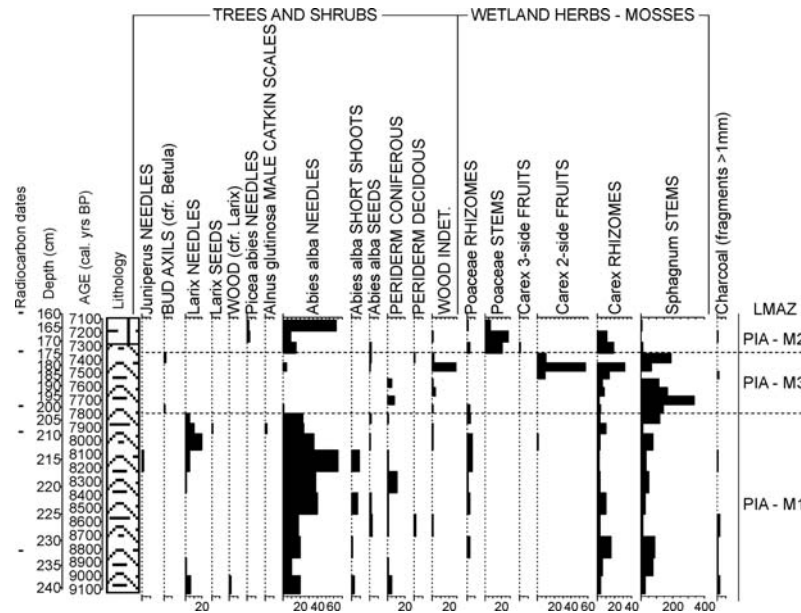


Figure 4 Plant macrofossil concentration for Piano mire. Values refer to number of items in a sediment volume of 25 cm³. On the left the lithostratigraphy is shown (see Fig. 2 for legend)

pattern, with *Abies alba* needles first decreasing (7900–7700 cal. a BP) and then disappearing (7700–7400 cal. a BP; Fig. 4). These palaeobotanical data suggest a sudden collapse of the *Abies alba* stands that had previously dominated for almost 2 ka (Fig. 3). *Pinus sylvestris* also declined in the area and we hypothesise that the species was growing on dry/shallow soils (e.g. rock ridges) or at higher altitudes, where it is also abundant today. The pollen data suggest that *Betula* stands replaced the former *Abies* forests around the mire. It was not possible to identify whether the species was *Betula pendula* or *B. pubescens*. A local expansion on the mire of

B. pubescens seems unlikely since no macroremains were found despite the high pollen percentages of *Betula* (Fig. 4). Instead we assume that *B. pendula* expanded on the slopes around the bog. Indeed, today *Betula pendula*, a light-demanding species, is present in the Piano area. In correspondence with the decline of *Abies* (and *P. sylvestris* type) pollen percentages ca. 7450 a huge (and unusual) peak of *Fraxinus excelsior* pollen (reaching 40%) suggests the expansion of the species on the Piano mire area. This pattern is mirrored to a lesser degree by pollen representatives of the vegetation at lower altitude (e.g. *Quercus* deciduous, *Tilia* and *Ulmus*).

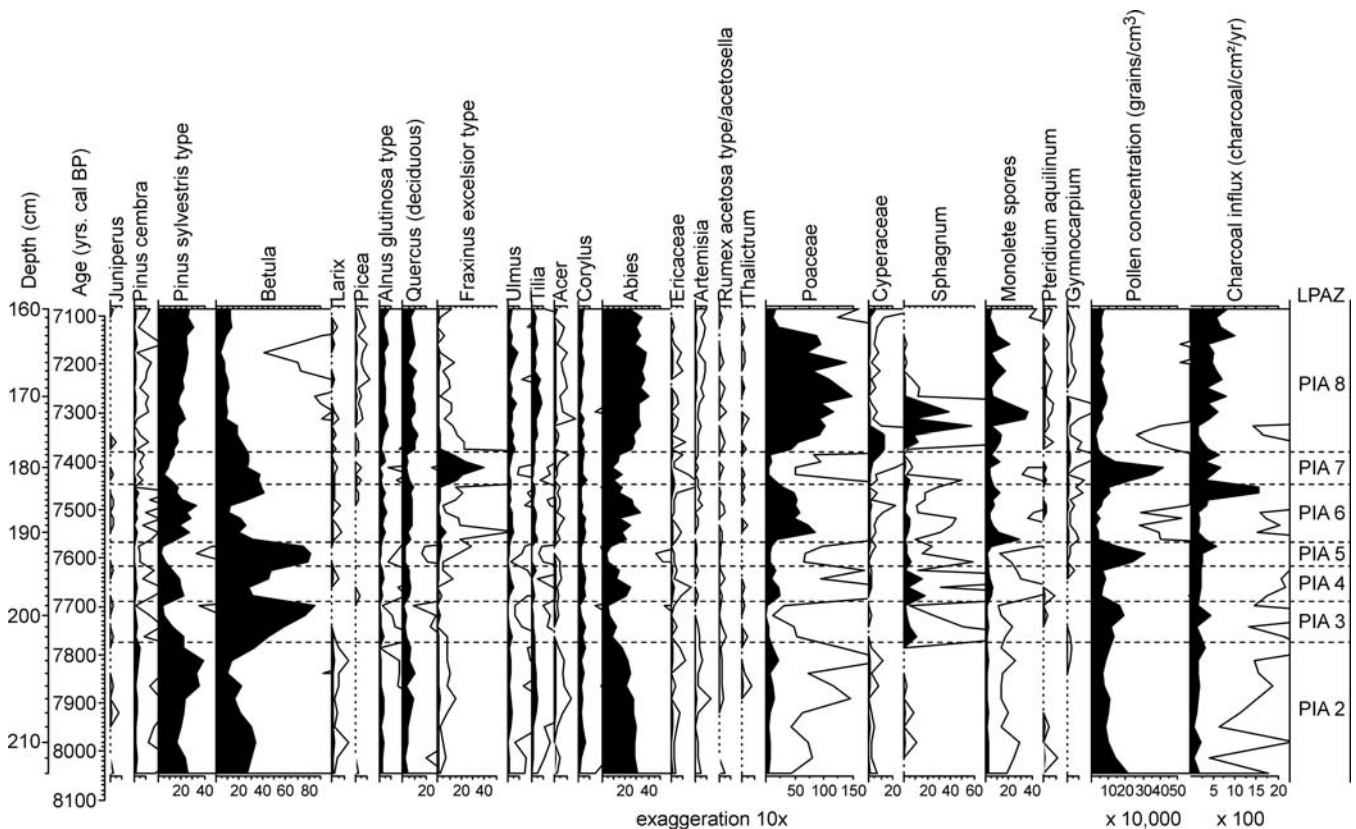


Figure 5 Pollen percentages and charcoal influx for Piano mire for the high-resolution section (ca. 7100–8100 cal. a BP)

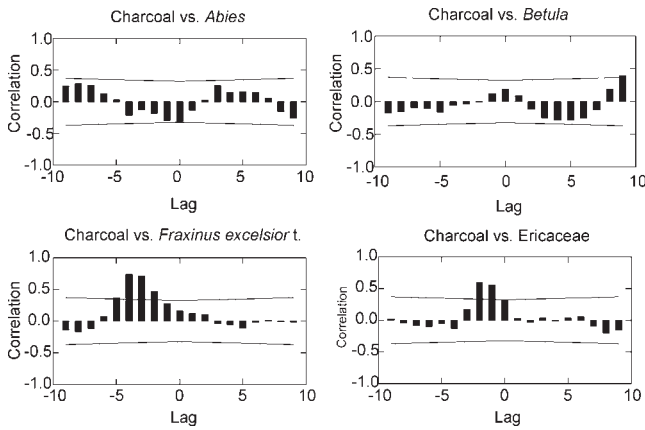


Figure 6 Cross-correlograms of charcoal influx vs. pollen percentages for the time frame ca. 7350–8050 ± 140 cal. a BP. The x-axis indicates the lag (19 ± 4 calibrated years) and the y-axis is the coefficient of correlation. Significance levels are indicated by horizontal lines

Microscopic charcoal peaks at ca. 7720, 7580 and 7450 cal. a BP (during the minima of *Abies*) suggest that regional fire activity increased during *Abies* (and *Pinus sylvestris* type) collapses (Fig. 5). However, cross-correlation between charcoal and *Abies* (and *Betula*) for the period 7800–7400 cal. a BP show no statistically significant correlation, though negative correlations at lag 0 almost reach the significance level for *Abies* (Fig. 6), indicating that these taxa were not influenced by the fire activities. The major changes in the pollen percentage spectra are roughly mirrored in pollen concentration and influx (Fig. 7). Nevertheless, the occurrence of three distinct collapses of pollen percentage of *Abies* is not easily visible in the influx values. Indeed, *Abies alba* influx decreased around 8000 cal. a BP, reaching a minimum that lasted until 7500 cal. a BP. The last *Abies* pollen percentage decline (as well as those of *Pinus sylvestris* type) is accompanied by a strong increase in total influx. Indeed, *Sphagnum* reaches very low values for both spores and macrofossils, possibly due to a lowering of the water table. This may have affected peat accumulation rates around

7400 cal. a BP. On the other hand, pollen percentages and influx agree well for replacing taxa (e.g. *Betula*).

Mixed *Abies alba* forest from 7400 cal. a BP onwards

Around 7400 cal. a BP *Abies* and *Pinus sylvestris* type pollen regain their previous importance, whereas *Betula* pollen decreases. At the same time *Picea* pollen reaches its empirical limit (Fig. 5). Our pollen results are corroborated by the reappearance of plant macrofossils of *Abies alba* at ca. 7400 cal. a BP (Fig. 4). These data suggest the local re-establishment of *Abies alba*-dominated forests after 7400 cal. a BP. The local presence of *Picea abies* in these forests after 7400 cal. a BP is unambiguously documented by needles (after 7200 cal. a BP). As in the previous zone, a marked increase of Poaceae accompanies the recovery of the *Abies*-dominated forests. We assume that this mirrors a local expansion of Poaceae on the mire, perhaps at the cost of *Sphagnum* mosses (conversion from a bog to a fen). This marked transition in local wetland vegetation is also recorded in plant macrofossils (Fig. 4). The peat younger than 7200 cal. a BP is richer in herbaceous components and *Sphagnum* stems completely disappear from the macrofossil record (PIA-M 3). After a slight decrease of microscopic charcoal influx, it increases slightly around ca. 7300 cal. a BP, suggesting a slight increase in regional fire activity.

Discussion

Early to mid Holocene vegetational responses to millennial-scale climatic trends in the Alps

Variation of orbital forcing resulted in a maximum of summer insolation and a minimum in winter insolation in the Northern Hemisphere during the early Holocene (Fig. 8). Seasonality was

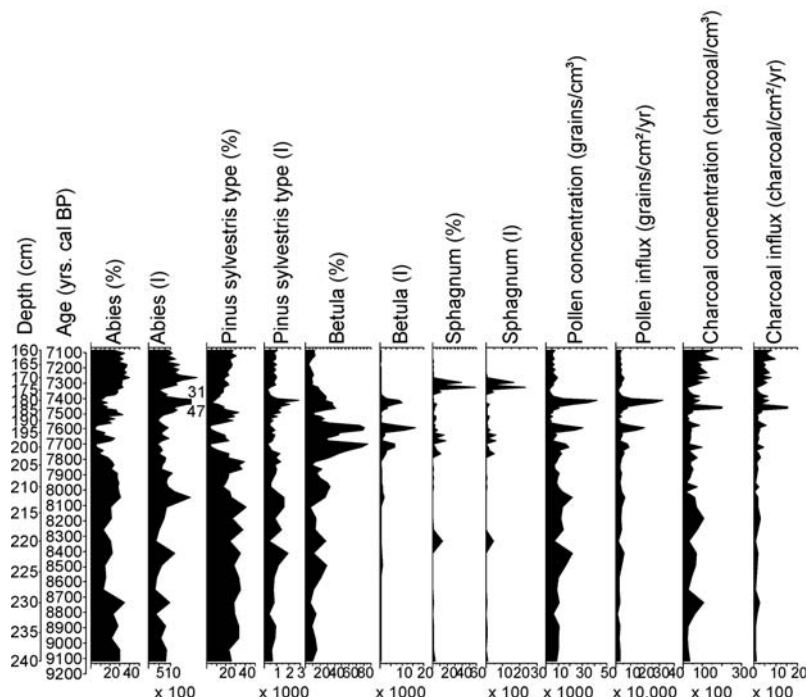


Figure 7 Pollen percentage (%), pollen influx (l), charcoal concentration and charcoal influx for Piano (7100–9200 cal. a BP)

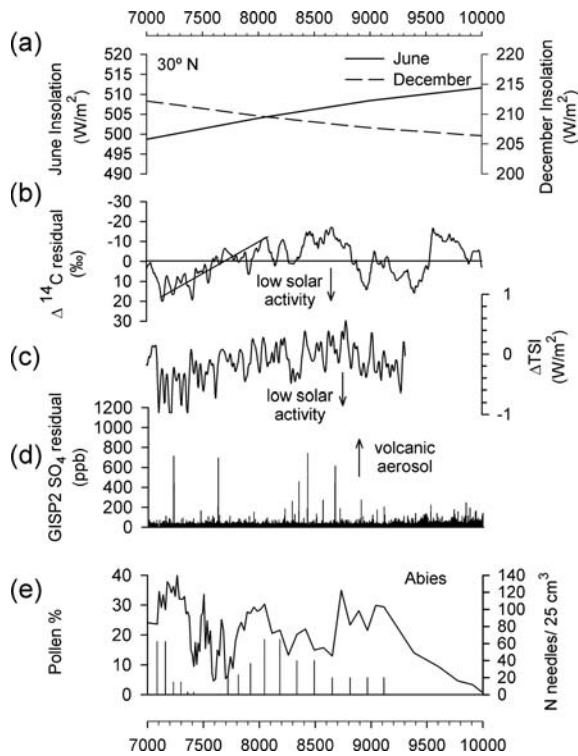


Figure 8 Comparison between external factors that may cause climate to change and *Abies* pollen percentages in Piano sediments. (a) Summer and winter insolation values at 30° N (Berger and Loutre, 1991). (b) Solar activity variations (higher values indicating lower solar activity), as recorded by radiocarbon variations in tree rings of known calendar age (Reimer *et al.*, 2004). (c) Reconstructed total solar irradiance based ^{10}Be measurements in polar ice (Steinhilber *et al.*, 2009). (d) Volcanic sulphate residuals from the GISP2 ice core (Zielinski *et al.*, 1994). (e) *Abies* pollen percentages and plant macrofossil concentrations (black bars) in the Piano sediments (this study)

greater than today during the interval 12 000–6000, with a maximum around 10 000 cal. a BP in the Northern Hemisphere (Kutzbach and Webb, 1993). Insolation variation affected regional climate dynamics, resulting in warmer summers and colder winters and less moisture availability (Kutzbach

and Webb, 1993). These extreme climatic conditions most probably inhibited the expansion of *Abies alba* at Piano during the early Holocene. The expansion of *Abies alba* occurred around 9000 cal. a BP, probably when climate became more favourable, i.e. soil and air moisture increased (Tinner *et al.*, 1999) and regional fire activity decreased (Figs 3 and 8). Similarly, unsuitable (orbitally forced) climatic conditions have been used to explain the almost complete lack of mesophilous forests in central Europe during the early Holocene (e.g. Rudolph, 1928; Welten, 1944; Firbas, 1949/1952; Tinner and Lotter, 2001).

Vegetation response to centennial-scale climatic change in the Alps

Superimposed on orbitally induced climatic changes, centennial to decadal climate variability might have influenced ecosystems and vegetation. The climatic variability related to changes in solar irradiance, ocean forcing and volcanic eruptions might have pushed ecosystems over critical thresholds due to abrupt changes in the climate system (Alley *et al.*, 2003). Such sudden changes in the climate are likely to be harmful for long-lived ecosystems such as mature forests (Williams *et al.*, 2002). At our site abrupt shifts in the vegetation started ca. 7800 cal. a BP, with collapse of the main tree *Abies alba*. At that time anthropogenic influences on vegetation was still weak (Mesolithic or Mesolithic/Neolithic transition). There is no evidence of pollen-inferred cultural indicators (e.g. *Plantago lanceolata* type, *Cerealia* type or *Urtica*; Behre, 1981) in our record (Fig. 3). Hence we assume that vegetation changes were not primarily caused by human activities. Similarly, cross-correlation analyses at Piano suggest that fire disturbance (natural or anthropogenic) was not the primary trigger of the *Abies alba* collapses (Fig. 6).

A similar series of rapid population collapses and expansion of *Abies* was recorded at Pian di Signano (Zoller, 1960), a site located in the upper mountain vegetation (1540 m a.s.l.) belt of the Southern Alps (Fig. 1). There *Abies alba* was repeatedly replaced by *Pinus sylvestris*, whereas at our site *Betula* mainly replaced *A. alba* (Figs 5 and 9). Because the two sites are at

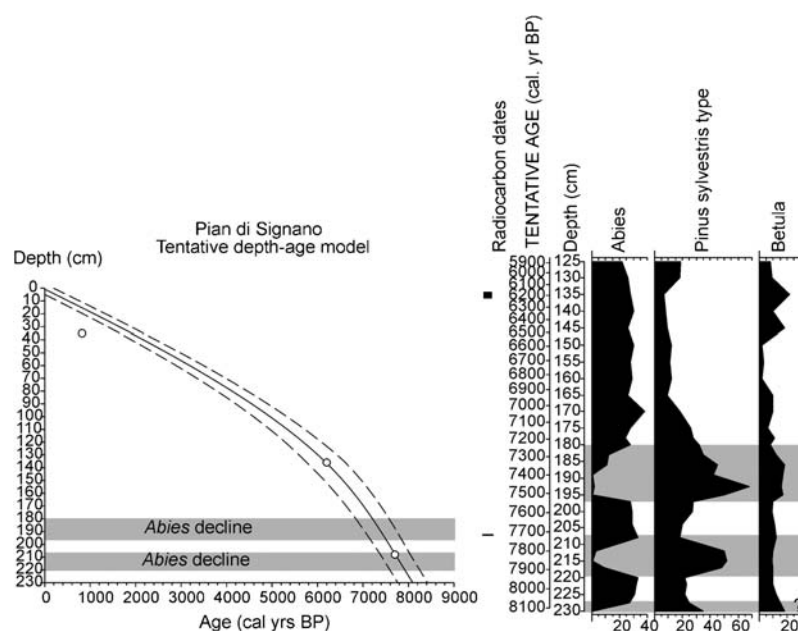


Figure 9 Left: tentative depth–age model (line) and 95.4% confidence intervals (dashed lines) of Pian di Signano defining the Misox event. Dots represent the calibrated radiocarbon dates (see Table 2). Right: pollen percentage diagram digitised from Zoller (1960) showing the most important taxa. Poaceae, Cyperaceae and ferns are included in the pollen sum. Grey bands indicate the *Abies* minima that define the Misox event (see Discussion)

similar altitudes, it is unlikely that the differences in vegetation composition were caused by temperature. Pian di Signano is located in a more continental and drier valley, where today *Pinus sylvestris* is still an important element of the forest. We assume that in the more oceanic Piano area *Pinus sylvestris* (which prefers continental environments; Ellenberg, 1996) was more affected by precipitation increases than at the continental site, so that pioneer *Betula* took advantage of the forest collapses. Our results and non-pollen evidence of climatic change (Fig. 10; e.g. Heiri *et al.*, 2003; Wick *et al.*, 2003) confirm the original interpretation of Zoller (1960) that widespread *Abies* population collapses in the mountain belt of the Southern Alps were probably caused by cold phases (MisoX event). Indeed, *Abies alba* is highly sensitive to late spring frost and keen winter frost (Ellenberg, 1996). However, both sites recording the *Abies* declines are situated ~800 m below the tree line and ~400–500 m below the upper limit of *Abies alba* in the region. Thus *Abies* communities should have been well buffered against a temperature decrease of Holocene (~2°C). We hypothesise that a unique combination of wet (e.g. Magny and Schoellammer, 1999; Spötl *et al.*, 2010) and cold years led to repeated lethal disadvantages for *Abies* (Fig. 10). Afterwards, *Abies* populations recovered when temperature in the Alps increased at ca. 7500 cal. a BP, as suggested by chironomid-inferred temperature (Heiri *et al.*, 2003) and Swiss glacier retreated (Fig. 10; Joerin *et al.*, 2006, 2008).

Within the mutual dating uncertainties, the repeated collapses of *Abies* at Piano are in good chronological agreement with a long-lasting decline of *Pinus cembra* forests at the tree line in the nearby Swiss Central Alps (Gouillé Rion, Fig. 10; Tinner *et al.*, 1996; Wick and Tinner, 1997; Reasoner and Tinner, 2008), which has also been successfully simulated by using independent climatic evidence from the Northern Alps (Heiri *et al.*, 2006). These modelling experiments showed that tree line pollen and macrofossils primarily record population dynamics and not productivity variations as an alternative explanation (Ammann *et al.*, 2000; van der Knaap and van Leeuwen, 2003). Farther east in Austria, high-altitudinal *Pinus cembra* forests also declined at ca. 8400–7400 cal. a BP as a result of low temperature during the growing season (Kofler *et al.*, 2005). Climatic evidence from the Eastern Alps (Grotta di Ernesto; Frisia *et al.*, 2005; Spötl *et al.*, 2010) suggests an anomalously high rainfall between ca. 8200 and 7300 cal. a BP, which may have additionally affected *Pinus cembra* stands in that region (Fig. 10).

In the Alps a series of cold and wet climatic excursions had started ca. 8400 cal. a BP (Fig. 10; e.g. Haas *et al.*, 1998; Heiri *et al.*, 2003, 2004; Tinner and Kaltenrieder, 2005; Spötl *et al.*, 2010). Linkages between solar output fluctuations and Holocene climatic and environmental centennial-scale variability have been suggested by several authors (e.g. Denton and Karlén, 1973; Hu *et al.*, 2003; Rohling and Palike, 2005). The *Abies* collapse at Piano shows a rather good temporal agreement with high values of residual $\Delta^{14}\text{C}$ (indicating low solar activities) and low reconstructed total solar irradiance (Fig. 8), which may have influenced climatic and environmental systems (Reimer *et al.*, 2004; Steinhilber *et al.*, 2009). Nonetheless, chronologically the onset of the *A. alba* collapse at Piano is at ca. 7800 cal. yr BP \pm 120 a, in agreement with final freshwater pulses into the Atlantic ocean (e.g. Carlson *et al.*, 2009), perhaps suggesting that this marine event may have influenced climatic and environmental conditions in the Alps. However, this does not explain why previous (e.g. 8400–8200 cal. a BP) changes in the Atlantic ocean circulation did not affect the climate and vegetation in the Southern Alps. In addition to this, multiple volcanic

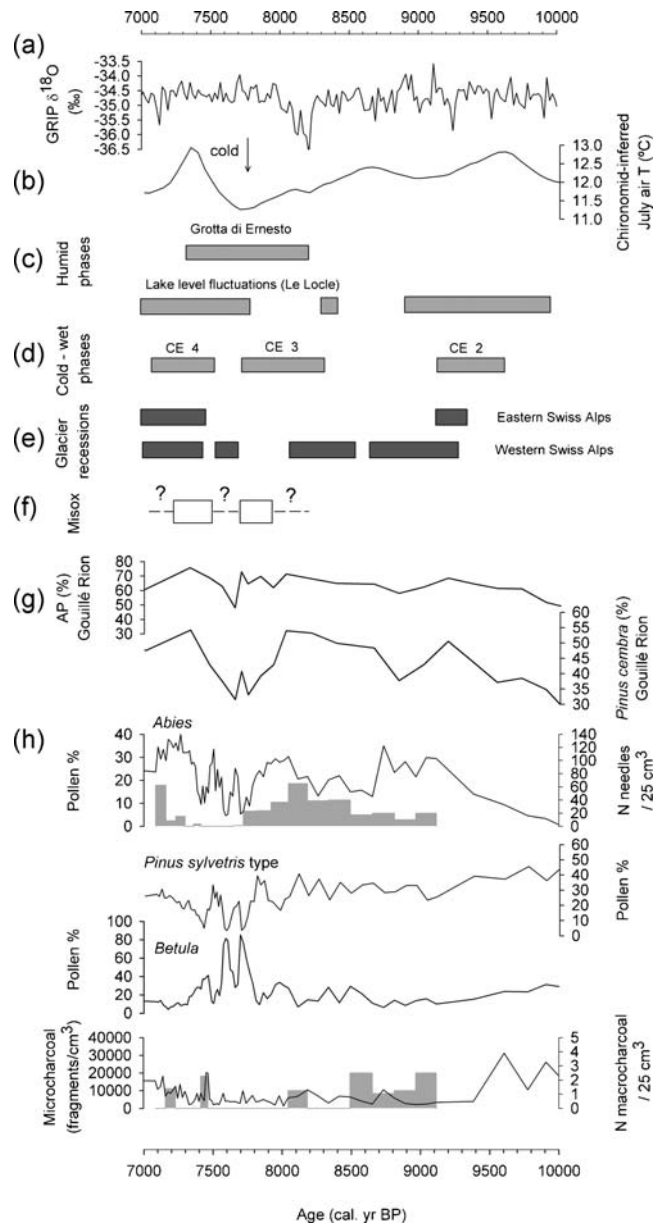


Figure 10 Comparison between well-dated climate proxy records and vegetation changes at our site. All records, including those from the ice cores, are presented vs. calibrated age (a BP), where ‘before present’ refers to before 1950. (a) GRIP $\delta^{18}\text{O}$ (‰) proxy for temperature (chronology GICC05; Rasmussen *et al.*, 2006; Vinther *et al.*, 2006). (b) Chironomid-inferred July air temperatures from the Hinterburgsee (Switzerland; Heiri *et al.*, 2003, 2004). (c) Episode of distinct humid phases. Above: block identifies a major humid phase reconstructed by the $\delta^{18}\text{O}$ (‰) from a stalagmite in a small cave called Grotta di Ernesto (Northern Italy; Spötl *et al.*, 2010) and blocks below identify higher lake-level events reconstructed at Le Locle (Switzerland; Magny and Schoellammer, 1999). (d) Cold and wet phases defined in the Swiss Plateau and at the timber line in the Alps (Haas *et al.*, 1998) indicated by grey blocks. (e) Episodes in which Swiss alpine glaciers were shorter than the 1985 reference level (Joerin *et al.*, 2006, 2008). (f) MisoX event as defined by Zoller (1960) with calibration of the former dates (see discussion). (g) Arboreal pollen (AP) and *P. cembra* pollen percentage in Gouillé Rion as proxy for tree line fluctuations (Switzerland; Tinner *et al.*, 1996; Heiri *et al.*, 2004). (h) Major pollen types and microcharcoal concentration in Piano sediments (this study); grey blocks indicate plant macrofossils (or macrocharcoal) concentrations

eruptions could have reduced hemisphere temperature on decadal and multi-decadal timescales (Briffa *et al.*, 1998), reinforcing the effects of meltwater pulses and declining solar activity at 8000–7500 cal. a BP (Fig. 8).

Table 2 Calibration of radiocarbon dates on Piano di Signano core (Zoller, 1960)

Depth (cm)	Material dated	¹⁴ C a BP	Calendar age (cal. a BP, 2 SD)
35	Peat	890 ± 60	700–900
135–137	Peat	5440 ± 100	6000–6400
208	<i>Abies</i> wood	6850 ± 120	7500–7900

The Misox event and its correlation with the 8200 cal. a BP event

The pioneering study of Zoller (1960) relied in total on three radiocarbon dates (two on bulk peat and one on *Abies* wood), two of which fall into the period of interest (Table 2). We recalibrated the radiocarbon dates using the CALIB rev. 5 program (Stuiver and Reimer, 1993; Stuiver *et al.*, 1998) and developed a depth–age model using the same technique as for the Piano record (Fig. 9). According to this depth–age model the two marked *Abies* declines start at ca. 7900 ± 320 cal. a BP (lasting ca. 200 a) and 7500 ± 320 cal. a BP (lasting ca. 300 a; Fig. 9). Nonetheless, the sequence starts with low *Abies* pollen percentages (one pollen sample at the base of the core) at ca. 8100 ± 320 cal. a BP, pointing to an earlier *Abies* decline that is chronologically in agreement with the 8200 cal. a BP event, as previously suggested (Wick and Tinner, 1997). As shown by the tentative depth–age model, the dating of the onset of the Misox event is based on extrapolation, while the termination age of ca. 7200 cal. a BP relies on interpolation to the next radiocarbon date, together underscoring that the chronology of the Misox event is not well defined (Fig. 9).

Considering that the Piano record has good chronological control across the *Abies* collapses (Fig. 2), it seems evident that the vegetation reorganisation at Piano cannot be linked with the 8200 cal. a BP event (Fig. 10). On the other hand, in the Alps a series of cold and wet climatic excursions had started around 8400 cal. a BP (i.e. chironomid-inferred temperature, Grotta di Ernesto and CE's records; Fig. 10) and lasted significantly longer than the 8200 cal. a BP event as recorded in the oxygen isotope records (e.g. Dansgaard *et al.*, 1993; von Grafenstein *et al.*, 1998; Thomas *et al.*, 2007). Taken together, two alternative explanations may summarise the situation according to our comparative study: (a) given the lack of chronological control the Misox event (as defined by Zoller, 1960) is not correlated with the 8200 cal. a BP event (see, as an example, Bennett, 2002); (b) climatic deteriorations in and around the Alps had complex and small-scale spatial patterns, which are not easily linked with the 8200 cal. a BP event. We conclude that further investigations in the Southern Alps are needed to better establish the link between the 8200 cal. a BP event and the Misox event.

Conclusions

High-resolution pollen, charcoal and plant macrofossil analyses at Piano (1439 m a.s.l., Ticino, Swiss Southern Alps) allow reconstruction of vegetation and fire dynamics during the early to mid Holocene transition in the upper mountain belt. *Abies alba* forests in the region were established during the early Holocene, when climatic conditions gradually became more suitable for the needs of this species. A marked vegetational change occurred ca. 7800–7400 cal. a BP, when

Abies forests were replaced by pioneer *Betula* stands. Independent palaeoclimatic proxies suggest that *Abies* declined in response to cooler and wetter conditions. The decline of *Abies alba* at Piano di Signano (a site located 50 km from Piano), was used to define the Misox cold event by Zoller (1960). Biostratigraphically Piano di Signano and our new site Piano are in good agreement. Although considerable uncertainty exists about the chronology of Zoller's site, the Misox event has been correlated with the 8200 cal. a BP cooling in Greenland (e.g. Wick and Tinner, 1997). Our new study points to substantial spatial and temporal variability; the initial decline of *Abies alba* at around 8200 cal. a BP is missing in our new record. This implies that care must be taken when correlating climatic anomalies or climatic impacts at around 8000 cal. a BP with the 8200 cal. a BP event. Our study confirms the point of view of Zoller (1960) that Holocene climatic changes were strong enough to profoundly influence vegetational trajectories far below the tree line.

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