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Glacial to paraglacial history and forest recovery in the Oglio glacier system (Italian Alps) between 26 and 15 ka cal BP

Cesare Ravazzi [a,*], Federica Badino [a,b], Diego Marsetti [c], Glauco Patera [a], Paula J. Reimer [d]

[a] C.N.R. – Istituto per la Dinamica dei Processi Ambientali, Laboratory of Palynology and Palaeoecology, Piazza della Scienza 1, 20126 Milano, Italy
[b] Dipartimento di Scienze dell'Ambiente e del Territorio, Università di Milano Bicocca, Piazza della Scienza 1, 20126 Milano, Italy
[c] Ecogeo s.r.l., Via Fratelli Calvi 2, 24100 Bergamo, Italy
[d] Centre for Climate, the Environment & Chronology, School of Geography, Archaeology and Palaeoecology, Queen's University of Belfast, 42 Fitzwilliam Street, Belfast, United Kingdom

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ABSTRACT

The integrated stratigraphic, radiocarbon and palynological record from an end-moraine system of the Oglio valley glacier (Italian Alps), propagating a lobe upstream in a lateral reach, provided evidence for a complete cycle of glacial advance, culmination and withdrawal during the Last Glacial Maximum and early Lateglacial. The glacier culminated in the end moraine shortly after 25.8 ± 0.8 ka cal BP, and cleared the valley floor 18.3–17.2 ± 0.3 ka cal BP. A primary paraglacial phase is then recorded by fast progradation of the valley floor.

As early as 16.7 ± 0.3 ka cal BP, early stabilization of alluvial fans and lake filling promoted expansion of cembran pine. This is an unprecedented evidence of direct tree response to depletion of paraglacial activity during the early Lateglacial, and also documents the cembran pine survival in the mountain belt of the Italian Alps during the last glaciation. Between 16.1 and 14.6 ± 0.5 ka cal BP, debris cones emplacement points to a moisture increase favouring tree *Betula* and *Pinus sylvestris-mugo*. A climate perturbation renewed paraglacial activity. According to cosmogenic ages on glacial deposits and AMS radiocarbon ages from lake records in South-Eastern Alps such phase compares favourably with the Gschnitz stadial and with the oscillations recorded at lakes Ragogna, Längsee and Jeserzersee, most probably forced by the latest freshening phases of the Heinrich Event 1.

A further sharp pine rise marks the subsequent onset of Bølling interstadial.

The chronology of the Oglio glacier compares closely with major piedmont glaciers on the Central and Eastern Alpine forelands. On the other hand, the results of the present study imply a chronostratigraphic re-assessment of the recent geological mapping of the Central Italian Alps.

1. Introduction

During the last glaciation the core belt of the Alps was an ice-field from where radiated a network of valley glaciers (Fig. 1). Comparing timing and extent of glacial advances and deglaciation timelines of individual valley glaciers is a major challenge to unravel the effects of global and hemispheric climate changes on the recent history of the Alpine orogene (Kelly et al., 2004) and on its modern biodiversity and conservation (Stehlik et al., 2002; Schönswetter et al., 2005). The development of valley glaciers on the southern side of the Alps is especially intriguing, given that significant portions of the outer mountain belt remained unglaciated and highly biodiverse. In turn the persistence of herb vegetation and even of woody plants in the surroundings of the glaciated terrain allows for direct radiocarbon dating of glacial settings and of vegetation successions in deglaciated forefields (Monegato et al., 2007).

In the Alps, the last deglaciation from outermost Würmian morainic ridges started around 23 ka cal BP, while the main Alpine trunk valleys were ice free as early as 18 ka cal BP (Ivy-Ochs et al., 2008). The rapid modifications undertaken by formerly glaciated landscapes after deglaciation, i.e. the paraglacial phase (Church and Ryder, 1972; Ballantyne, 2002), is poorly known in the Alpine valleys. Furthermore, while the available palaeoecological record in the glaciated southern side of the Alps extends back to about 17 ka cal BP (Vescovi et al., 2007; one site back to ca 19 ka cal BP, Huber et al., 2010), the response of terrestrial and aquatic habitats to deglaciation and to paraglacial evolution is little documented and only roughly dated.

* Corresponding author. Tel.: +39 035 6224263; fax: +39 035 6224260.
E-mail address: cesare.ravazzi@idpa.cnr.it (C. Ravazzi).

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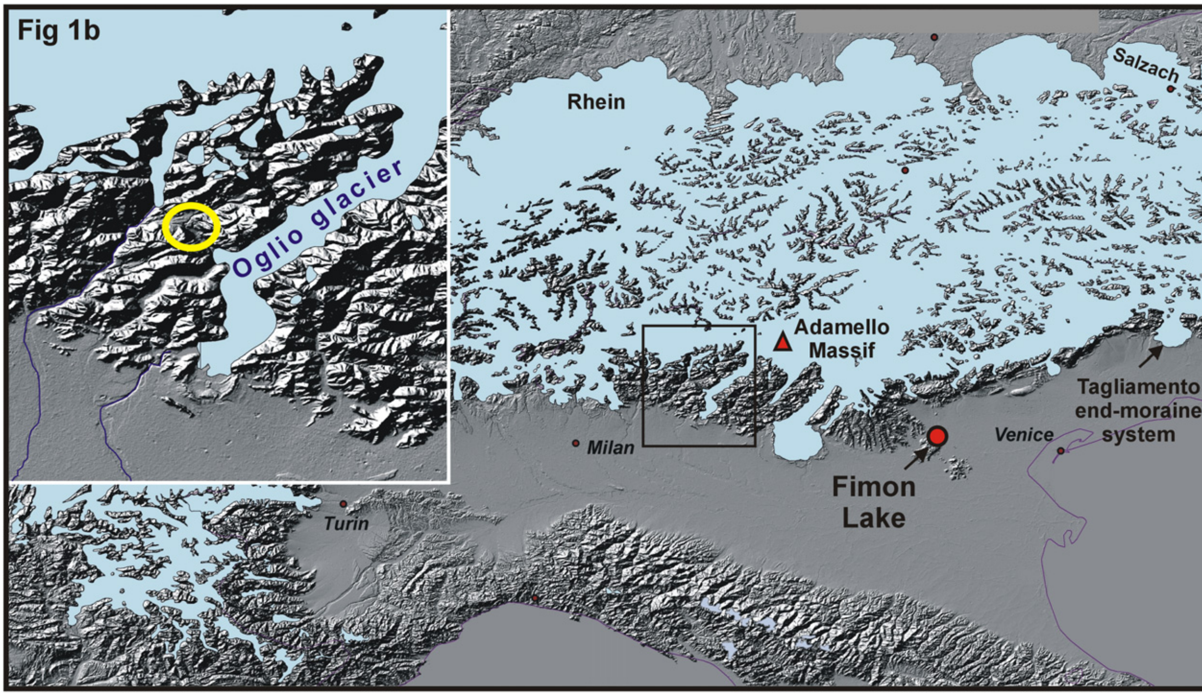

Fig. 1. Maximum extent of the last glaciation in the Alps (from Ehlers and Gibbard, 2004) and detail (b) of the Lombardy unglaciated Alps with location of the study area. Compare Fig. 2 for an updated version of the maximum glacier extents.

In this paper we analyse the glacial and paraglacial evolution of a major glacier system in the southern side of the Alps between the LGM and the Lateglacial. Field geological survey is merged with an integrated stratigraphic analysis of dammed lakes and of the valley floor fills. The recognition of plant debris and pollen in the paraglacial infill allowed chronological and palaeobotanical insight on early environmental history of the Italian Alps at deglaciation time, and also promoted a re-assessment of the glacier limits at LGM culminations for this Italian sector of the Alps. We provide evidence for the persistence of *Pinus cembra* in the unglaciated mountains of the Italian Alps in the Last Glacial Maximum. We show that glacial contraction promoted settlement of pioneer tree stands on the deglaciated terrain on the southern side of the Alps, thousands years before the mass expansion of Lateglacial forest. A subsequent development of the geo-ecological paraglacial landsystem highlights the effects of climate events that affected the early Lateglacial prior to the sharp warming marking the onset of Bølling–Allerød biostratigraphic complex.

2. The most recent glaciations in the study area (Figs. 1 and 2)

One of largest glaciers in the Italian Alps occupied the catchment of the Oglio River in the most recent Quaternary glaciations (Fig. 1). The Oglio glacier originated from the Adamello Massif (top at 3554 m a.s.l.) and settled the Val Camonica, including the cryptodepression nowadays occupied by Lake Iseo, 185 m a.s.l. (Fig. 2). The survey of glacial landforms has been dealt with by Vecchia (1954), Bini et al. (2007), and Cassinis et al. (2012).

The western tributaries of the Val Camonica include a main valley oriented NW–SE, i.e. the catchment of the Borlezza River (Fig. 2). During the last Quaternary glaciations, the uppermost Borlezza Valley hosted only small local cirque glaciers, not coalescent with the main ice flow (Haupt, 1938; Chardon, 1969). On the other hand, a lateral lobe of the Oglio Glacier propagated upstream in the tributary Borlezza Valley forming a second, smaller piedmont lobe, the so-called Clusone amphitheatre, occupying a tectonic basin closed within the mountain belt (45°52′ N, 9°57′ E, Figs. 2 and 3). The age of this amphitheatre is debated. Although early scholars assigned it unanimously to the last (Würmian) glaciation (Desio, 1944; Chardon, 1969; Orombelli and Ravazzi, 1995), a recent survey proposed to set the different morainic ridges to different glaciations, all of them predating the Late Pleistocene (Ferliga, in press). This latter view merged in the maps of the last glaciations by Ehlers and Gibbard (2004, Fig. 1) and by Bini et al. (2009).

The glacial and morainic dam built up by the Clusone amphitheatre diverted drainage westward to the unglaciated reach of the lower Seriana Valley (Chardon, 1969), while the drainage to the Lake Iseo was restored after deglaciation. During deglaciation phases, the lateral lobe retiring back to the lowermost reach dammed the deglaciated valley floor, forming lakes and allowing substantial rearrangement of slope profile, through development of laterally coalescent alluvial fans (Orombelli and Ravazzi, 1995). The lacustrine sediments, as well as their stratigraphic relations to the underlying till and to coeval and subsequent deposits formed during the paraglacial evolution of the valley floor, offer a potential to reconstruct the evolution, maximum extent and chronology of the entire glacier through the analysis of its lateral lobes and of the geomorphic and palaeoecological events at the glacial–interglacial transition.

3. Sediment sources signatures

The entire Borlezza Valley catchment cuts through the sedimentary cover of Southern Alps. Predominant detritus is made of Triassic carbonate rocks with a minor contribution of Triassic arenites (Jadoul et al., 2000). Detritus originating from local rock erosion is characterized by low to very low magnetic susceptibility (typically 0–10 $\times$ 10^{-5} unit SI), while higher values are recorded in weathered profiles, up to 50 $\times$ 10^{-5} unit SI, especially in cover of fersiallitic soils resting on unglaciated carbonate bedrock, because of their content of Fe-hydroxides and sesquioxides. The soil contribution to sediment composition is expected to be higher for

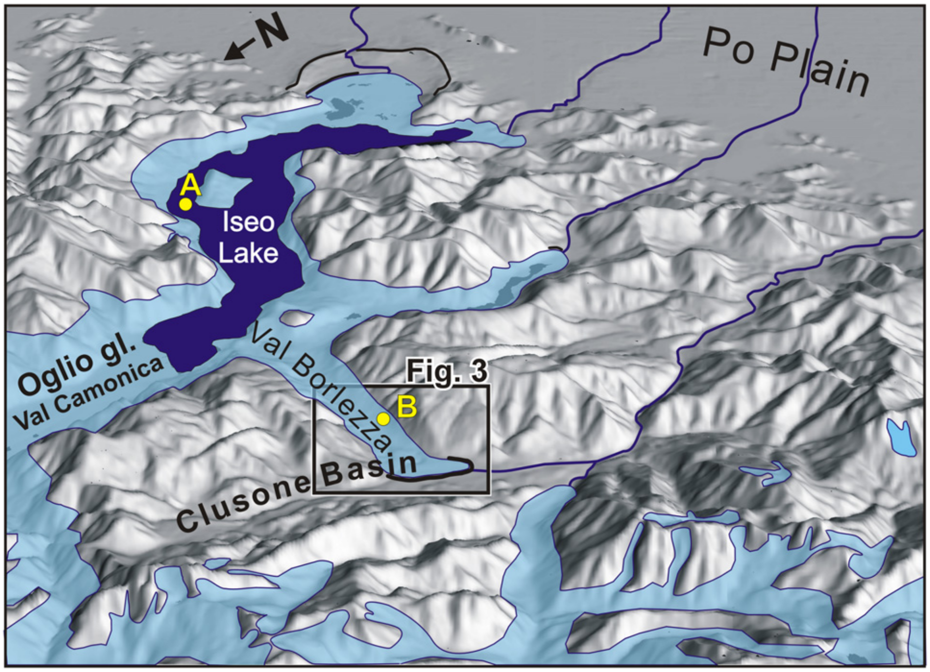

Fig. 2. Digital Terrain Model of the south-eastern sector of Lombardy Alps facing the Po Plain in a view from NW. The maximum extent of the Oglio glacier and of local glaciers during the last glaciation is drawn taking into account the results of the present work. The outer limit of the main end-moraine systems of the Iseo Lake and of Clusone plain are outlined by a thicker line. A – Drilling site SEB (Lauterbach et al., 2012). B – Drilling site Cerete Basso S3 (this work).

mass-movements affecting unglaciated slopes, while is limited for sediments deposited by the Borlezza River.

The major catchment of the Oglio River cuts through an extensive belt of Permian silicoclastic and volcanic rocks, but upstream it also reaches the axial belt of the Alpine chain, formed by metamorphic basement and by the Oligocene batholithe of the Adamello Massif (Bianchi et al., 1971). All these petrographic types, easily recognized on coarse detritus, display a susceptibility signal over 20×10^{-5} unit SI. Macroscopic individuals clearly assigned to crystalline-derived detritus can be treated as allochthonous to the Borlezza River basin. Micas and magnetite supplied from crystalline rocks mark the fine sediment yield released by glacial abrasion to lodgement till and ice-contact lacustrine deposits, with susceptibility peaks up to over 100×10^{-5} unit SI.

No reworked palynological assemblages derived from the sedimentary succession forming the Borlezza – inner Oglio River catchment have been detected.

4. Methods

A preliminary field survey of Quaternary deposits in the middle Borlezza Valley allowed recognition and characterization of the outcropping lithostratigraphic succession, its petrofacies and palynological content. Rotation drillings implemented with hydraulic extrusion, effective in fine sediments, were carried out at master borehole sites (S1–S4) selected at two basins apparently dammed by moraines ("Lama di Clusone", cores S1 and S2 see Figs. 3 and 6) or documenting the glacial–paraglacial sequence at the valley floor (Cerete Basso cores S3 and S4, Fig. 7 and 8). Geographic coordinates obtained from Google Earth (browsed 2012), altimetry from the Technical Regional Map 1:10.000. The following guidelines were referred for lithostratigraphy and geomorphology: Sanesi (1977), Eyles (1983), Miall (1983), FitzPatrick (1984), and Ballantyne (2002). Magnetic susceptibility was measured with a Bartington MS2 device equipped with a MS2E core logging sensor. Measures were taken on wet sediment at 10–20 cm intervals, and repeated to check environmental conditions changes.

Samples for LOI (Loss on Ignition) determinations were weighed wet and progressively heated at 105 °C for 12 h, 550 °C for 4 h and 950 °C for 2 h to evaluate the water, organic matter, total inorganic C contents and the residual fraction (Heiri et al., 2001). TOC, $CaCO_3$, and the non-carbonate residue were obtained stechiometrically (Dean, 1974, 1999). Lithostratigraphic units hence identified were numbered from bottom upward.

Six AMS ages were obtained from wood and debris of terrestrial plants on core Cerete Basso S3 (Table 1). Nine conventional radiometric ages derive from a cropping section previously studied at the same site (Orombelli and Ravazzi, 1995). Other ^{14}C ages were obtained from an exposed section of ice-contact lacustrine clay underlying till (Cuca section, Table 1), and from the base of an alluvial fan (Fonteno section). Overall, we avoided bulk sediment, aquatic plants and shells, due to well-known pitfalls in dating these materials. Calibration has been carried out using CALIB (v 6.1, Queen's University Belfast) with the IntCal09 calibration curve (Reimer et al., 2009). Throughout the text calibrated ages are used, which are indicated as yr cal BP or ka cal BP (AD for written sources).

The palynological record of core Cerete Basso S3 was built on 49 samples. Samples were treated according to standard methods (including HF and acetolysis), after adding *Lycopodium* tablets for pollen and charcoal concentration estimations (Stockmarr, 1971). Identification was performed at ×400, ×630 and ×1000 magnification under light microscopes. Pollen identification followed Moore et al. (1991), Punt et al. (1976–2009), Reille (1992–1998), Beug (2004) and the palynological collection of C.N.R – IDPA. Distinction between *P. cembra* type and *Pinus sylvestris/mugo* type was performed on complete pollen grains, according to Pini et al. (2009, Appendix 1). Pollen diagrams were drawn using Tilia 1.11, TGView 2.0.2 (Grimm, 1992, 2004). The pollen sum used for % calculations includes trees, shrubs, chamaephytes and all upland herbs except aquatic and wetland plants, with a pollen count of 585 ± 207 pollen grains (6 samples below 400). Pollen zonation

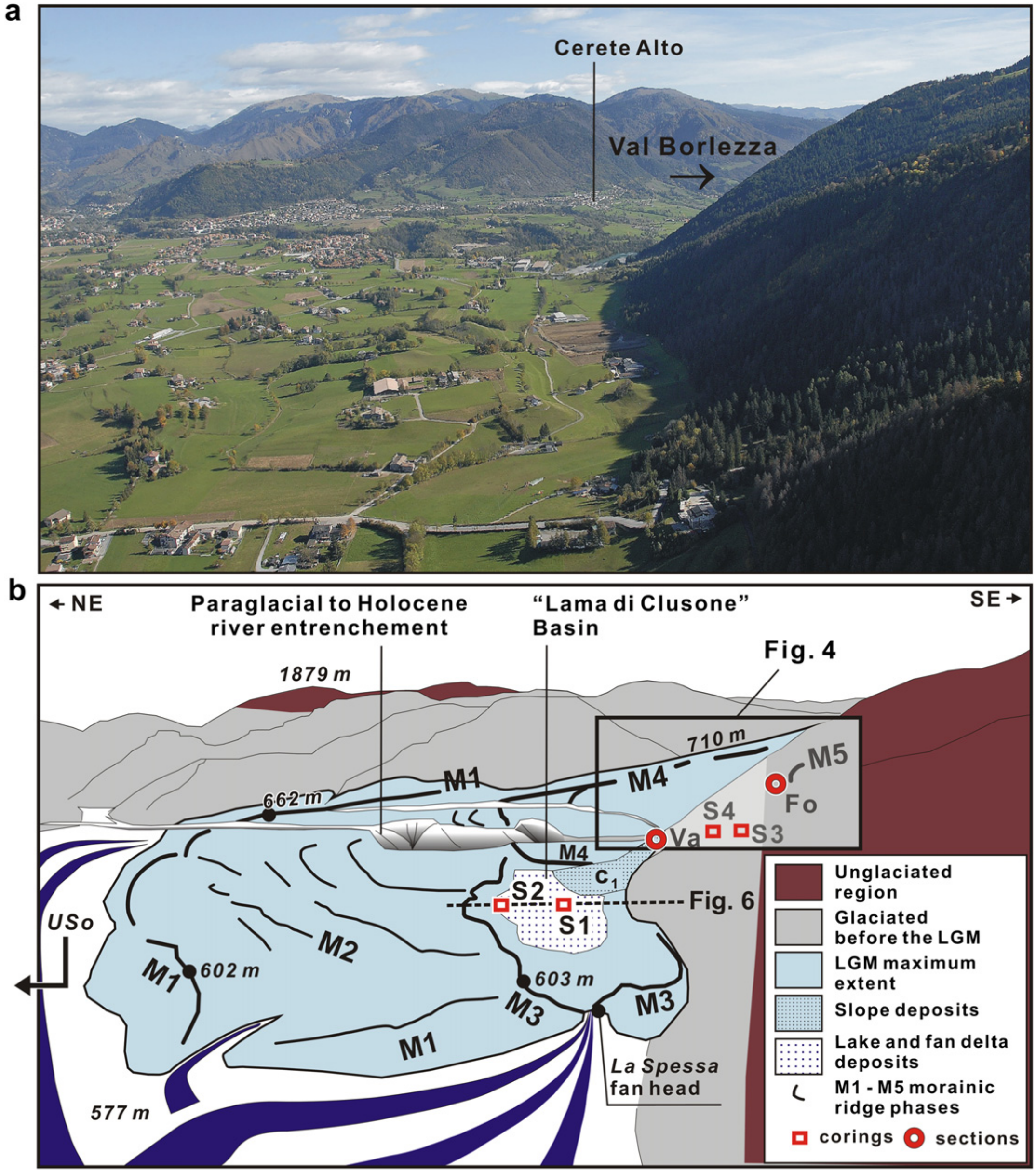

Fig. 3. (a) View from West to East of the plain of Clusone within the mountain belt of Lombardy Alps. On the right the Borlezza Valley. (b) Sketch of a showing the end-moraine system, morainic ridge phases, and related outwash plains in the glacier foreland. Also shown the position of the stratigraphic logs from cores (S1–S4) and from outcrops discussed in the present paper (Va = Valeggia section; Fo = Fonteno sections; Cu = Cuca section). USo, Unit of Stalle d'Onito (100 m northward, out of image).

into biostratigraphic zones of abundance was realized by a constrained incremental sums of squares cluster analysis using the Edwards and Cavalli-Sforza's chord distance as dissimilarity coefficient (CONISS, Grimm, 1987), restricted to taxa whose pollen percentage reached over 2%. Sporadic taxa were excluded from elaboration.

Palynomorphs preservation is good, apart from some disarticulated pine pollen grains. Black and opaque microcharcoal particles longer than 10 μm were counted in pollen slides.

4.1. Reference nomenclature and chronology

We rely on the regional (Alpine) chronostratigraphic subdivision of the last glaciation (Würm) (Chaline and Jerz, 1984; Preusser, 2004). However, the Last Glacial Maximum (LGM) is taken as a global climate-stratigraphic unit according to Lambeck et al. (2002), Orombelli et al. (2005), and Clark et al. (2009). The Lateglacial period is considered the time span between the end of the global LGM (e.g., 18 or 19 cal ka BP) and the beginning of the Holocene (e.g., 11.7 cal ka BP; Walker et al., 2009). The Lateglacial interstadial, broadly encompassing the Bølling–Allerød biostratigraphic complex, is taken from the Greenland ice stratigraphy GI-1 (Johnsen et al., 1997). The period constrained between global LGM and the onset Bølling–Allerød complex is informally given as "early Lateglacial". Chronology of hemispheric events refers to GICC05 (Svensson et al., 2008).

5. Presentation of sedimentary, radiocarbon and palaeobotanical evidence

5.1. Radiocarbon chronology

The radiocarbon chronology presented in Table 1 includes both original AMS determinations and conventional ages previously

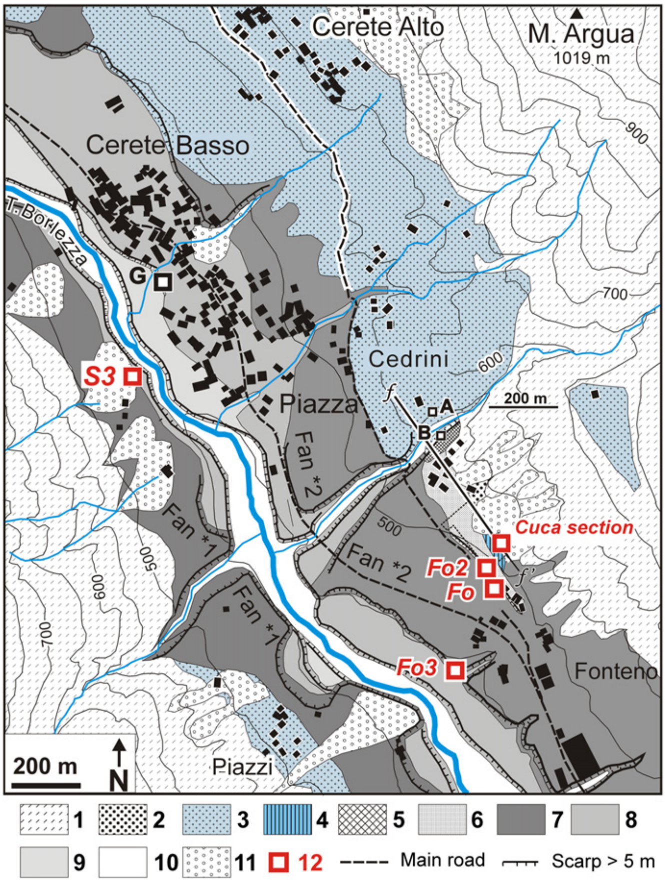

Fig. 4. Geological map of the middle Borlezza Valley valley floor, showing paraglacial evolution during deglaciation and later downcutting and terracing sequence. 1 – Triassic carbonate bedrock; 2 – Slope deposits covered by fersiallitic soils (Middle Pleistocene), mostly buried by subsequent glacial deposits; 3 – Glacial deposits related to M1–M4 phases of advance (Last Glacial Maximum); 4 – Ice-contact deposits, dated 21,550 ± 265 ^{14}C BP, i.e. 26–25.5 ka cal BP (Last Glacial Maximum); 5 – Glacial deposits related to M5 phase of advance; 6 to 8 – Terraced sequence of slope deposits (6) and of coalescent alluvial fans (7–9): 6 – Debris flows and landslides (early Lateglacial); 7 – Remnants of dissected, early Lateglacial fans, 17 to 15 ka cal BP; 8 – Bronze Age; 9 – Historical Age; 10 – Modern river bed including catastrophic floods occurred in the XIX century. 11 – Postglacial slope deposits (Lateglacial and Holocene); 12 – Position of drillings and studied sections. f–f′ – Trace of section shown in Fig. 5.

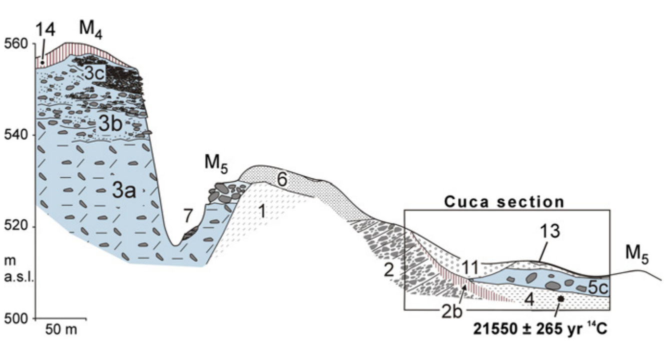

Fig. 5. The Quaternary sequence exposed at the slope foot of Mount Cuca along the f–f′ trace section (see position in Fig. 4). Bedrock (1) is irregularly mantled by glacigenic deposits (units 3 and 5c in light blue). The numbering of stratigraphic units and of morainic ridges M1–M5 are given in Fig. 4. Glacial deposits related to M4 phase of advance (number 3 in Fig. 4) are here detailed as follows: 3a – Lodgement till; 3b – Ice contact stratified drift and flow till; 3c – Melt-out till; 3d – Proximal glacifluvial deposits, partly collapsed after deglaciation. Glacial deposits related to M5 phase of advance (number 5 in Fig. 4) are here detailed as: 5c – Melt-out till. Additional slope units: 13 – Soil colluvial wedges and pockets. 14 – *In situ* alfisols. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

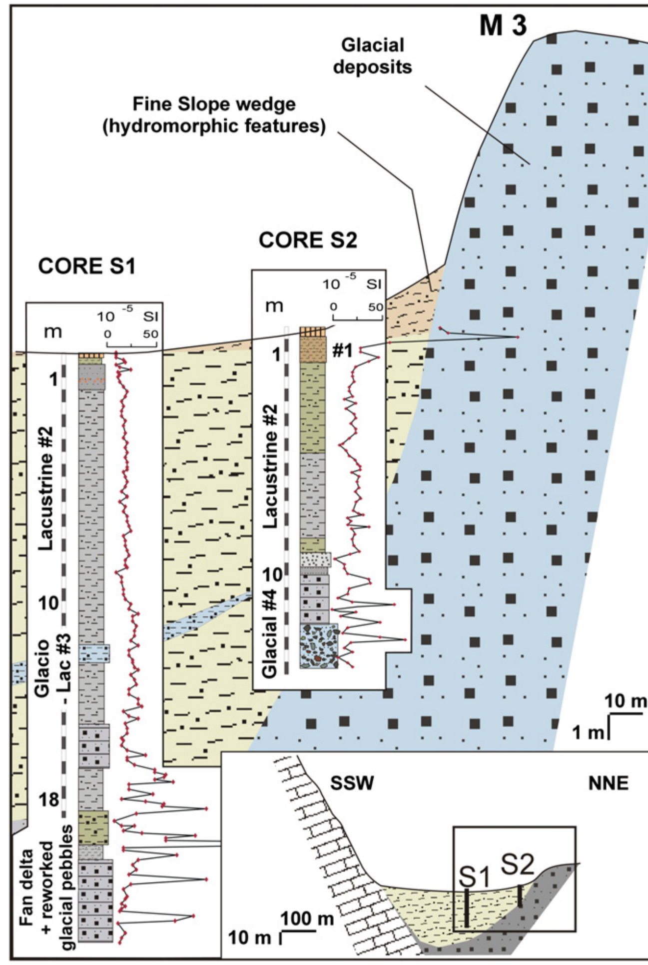

Fig. 6. Cores S1 and S2 obtained from the basin of the "Lama di Clusone" and magnetic susceptibility profiles. Drilling S2, which was cored close to the northern border of the basin, reached glacial deposits at 12 m depth. a (nested) – Geological sketch between the Triassic limestone bedrock (SSW) and the summit of the morainic arc M3 on the NNE side.

obtained on the uppermost segment of the Cerete Basso succession (CB) and on related alluvial fans (Fonteno 95). All of them were obtained from terrestrial plant material (Cyperaceae terrestrial peat in case of bulk samples). The precision obtained at UBA on early Lateglacial wood and terrestrial plant debris from the Cerete Basso S3 core allowed for a precise age–depth model spanning the paraglacial phase (Fig. 8). The AMS ages also allow chronostratigraphic inferences on age of glacial advances and of deglaciation.

5.2. The morainic amphitheatre of Clusone

A complex system of discontinuous, gentle-slope morainic arcs emerges from the intermountain plain of Clusone, despite burial of glacial landforms by meltwater and alluvial deposits (Fig. 3). Based on field observation of surface weathering profiles and of the terracing sequence, two main units could be distinguished:

- An outer glacial body (USo, *Unit of Stalle d'Onito*), highly terraced on the main alluvial plain and strongly modified by post-depositional processes, extends on the northernmost limit of the plain. Morainic ridges cannot be recognized. The weathering profile, exceeding 5 m, is strongly affected by clay illuviation and rubefaction (Munsell chroma 5 YR). The thickness of the uppermost Bt horizon is deeply influenced by

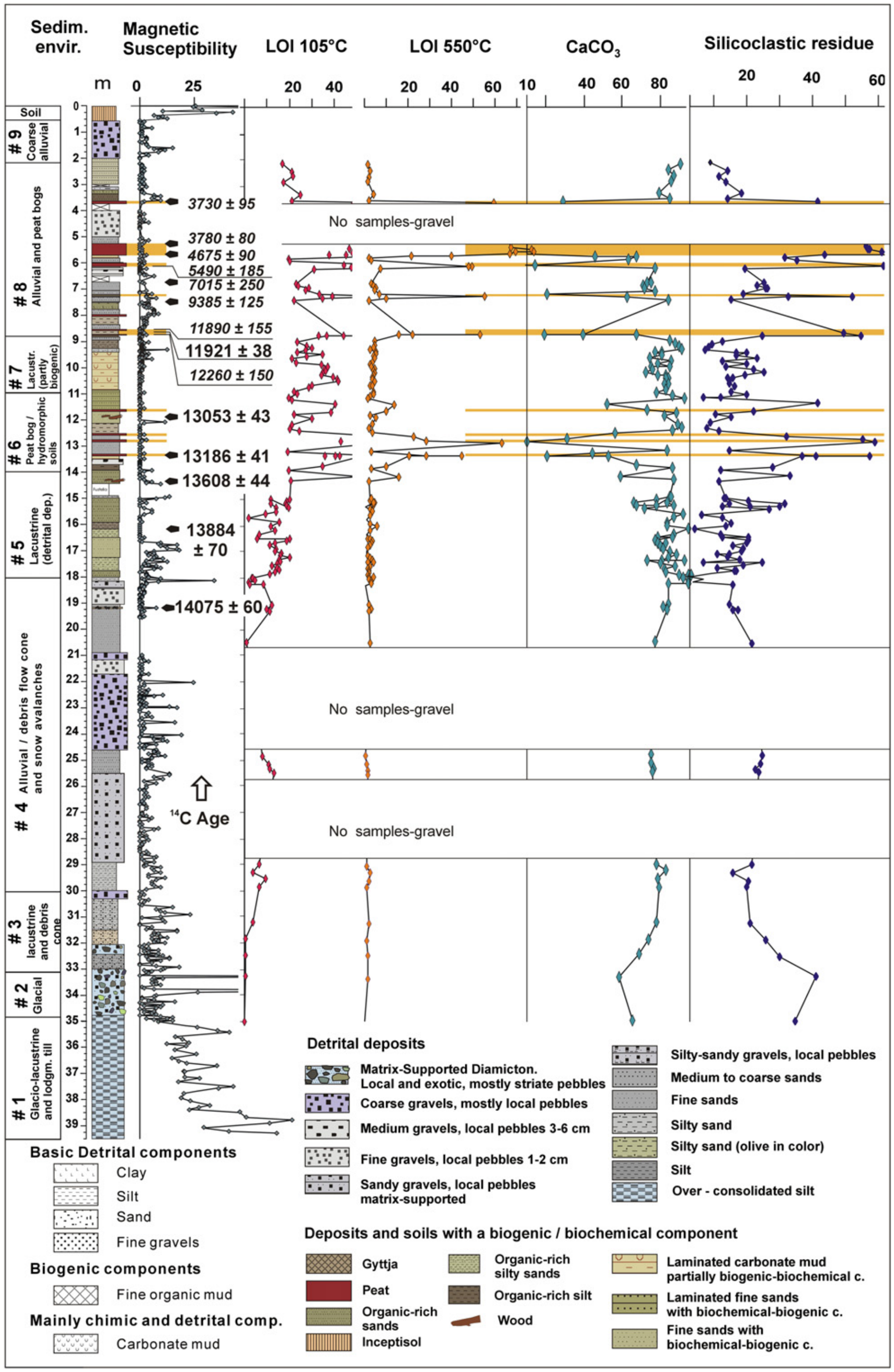

Fig. 7. Integrated stratigraphy and chronology of core Cerete Basso S3.

colluvial processes. Nevertheless, faceted blocks of allochthonous provenance can be still recognized imbedded in the rubified clay matrix, thus suggesting both glacial origin and provenance from the Oglio glacier.

- A complex system of morainic arcs emerging from the plain, to which outwash heads are connected by 3–4% gradient fans, covers an area of about 4.5 km^2 (Fig. 3). Four morainic systems (M1–M4) have been be distinguished. The external one (M1) forms a continuous ridge on the eastern side, up to 15 m high. This terminal moraine is dissected by an outwash on its western side (Fig. 3b). On the other hand M2 is composed by several discontinuous and smooth ridges, separated by small alluvial plains. M3 is marked by a prominent ridge, up to 25 m thick, constraining a lacustrine plain, called "Lama di Clusone" (575–585 m a.s.l., Fig. 3b), which is dammed by M4. The M4 ridge is well preserved in its eastern lateral side; the village of Cerete Alto rests on its top (Figs. 3 and 4). More glacial ridges occur downstream along the lower reach of the Borlezza River,

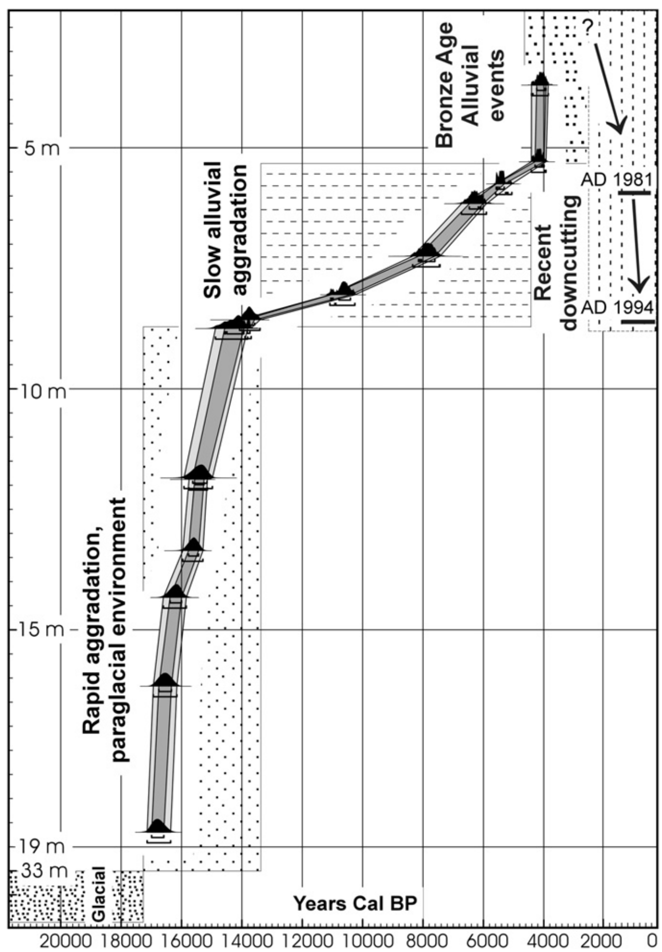

Fig. 8. Age–depth model of drilling Cerete Basso S3 showing 1 and 2 sigma probability envelopes and valley floor depositional history.

at lower altitude. They are mostly buried by alluvial fans and severely eroded after the postglacial downcutting of the valley floor. Lodgement till underlying melt-out till, belonging to a subsequent ridge (M5), outcrops close to the locality of Fonteno di Cerete (485 m a.s.l., ***Fo*** in Fig. 4; see Fig. 9.7).

Stable surfaces of M1–M5 moraines and of related glaciofluvial deposits display similar weathering profiles, supporting alfisols (see Appendix 1 for further information).

In order to obtain information about the age and evolution of deglaciation from the end-moraine amphitheatre, we cored the stratigraphic succession filling the Lama di Clusone basin down to the presumed underlying glacial body forming the M3 ridge (see Section 5.4). Geochronometric data could be obtained from a drift-mantled slope along the right side of the glacial lobe (Section 5.3) and from the sedimentary succession filling the valley floor (Section 5.5).

5.3. The LGM advance. Evidence from a drift mantled slope

The eastern side of the morainic amphitheatre of Clusone is well exposed after deglacial downcutting in the middle reach of the Borlezza Valley. Both the advancing phase, the culmination phase M4, the withdrawal phase M5 and the paraglacial evolution subsequent to deglaciation are documented by deposits preserved at the slope base and on the valley floor. The Quaternary sequence exposed at the foot of the drift-mantled slope of Monte Cuca (Fig. 4) allowed dating the time of the advancing glacier during the last glaciation.

A great mass of Quaternary deposits was quarried here in years 2003–2005, thus exposing the section sketched on the right of Fig. 5 (45°51′10.46″ N, 10°00′15.37″ E, 500–520 m a.s.l.). A clinostratified breccia with local petrofacies (unit #2) supports a cover of fersiallitic soils (unit #2b), chroma 5 YR, up to 3 m thick in pockets, whose irregular decarbonation front depends on intensity of carbonate dissolution, and translocation – accumulation of residual clay and hydroxides ("geologische Orgeln", see Egger and Van Husen, 2009). This palaeosol is buried by a pile of massive blue–grey silt (3 m visible), nesting allochthonous pebbles, a few of them faceted and striated (unit #4). Silt contains diatoms (Centrales) suggesting open lacustrine conditions, and pollen of *P. sylvestris/mugo* and *P. cembra*, pointing to their occurrence in the region, although a dominance of Cichorioideae (Table 2) depends on the bias by differential pollen deterioration (Havinga, 1984; Campbell, 1999). Within this unit, we recovered a fragment of bark, very compressed and deformed. Xylotomic identification is not feasible, but the periderm contains resins ducts and is assigned to a tree conifer, either *Pinus* or *Picea/Larix*. The fragment is dated to 21,550 ± 265 yr ^{14}C BP (25.8 ± 0.8 ka cal BP). The upper contact of unit 4 is marked by shear planes, oscurated by ground-water weathering, and in turn is covered by a blocky, matrix-supported massive diamicton, composed both by faceted and striated allochtonous blocks and by small angular, local pebbles (unit #5c in Fig. 5, thickness 4 m). The top profile is truncated.

The Cuca section testifies to significant events occurring before and during the last glaciation. In N-Italy the evolution of fersiallitic soils is restricted to warm-temperate phases of the Quaternary (Cremaschi, 1987) and to Eemian/MIS 5 in the Late Pleistocene (Cremaschi and Busacca, 1994; Ferraro, 2009). Given the minimum age of 26 ka cal BP, we argue that the parent material supporting soil evolution, i.e. a slope scree, accumulated before the last glaciation. Preservation of such a pre-glacial soil also suggests that the site remained stable until burial by ice-contact deposits related to an advancing glacier (unit #4). The latter advance occurred after 26 ka cal BP. Most probably the dated pine tree was living on the sunny slope, eradicated by the advancing glacier and incorporated after some floating in lake sediments, shortly before the glacier overriding.

We conclude that the local advance can be constrained soon after 25.8 ± 0.8 ka cal BP, i.e. in the early LGM. The palaeobotanical evidence also speaks for the occurrence of *P. sylvestris/mugo* and *P. cembra* at middle altitude on sunny slopes, within the inner belt of Italian Alps, at time of the Last Glacial advance.

5.4. The sedimentary succession filling an intermorainic basin

Two cores were drilled in the "Lama di Clusone" (Figs. 2 and 6), respectively at the basin centre (core S1, 45°52′24″ N, 9°57′40.32″ E, 575 m a.s.l.) and at the slope base of the delimiting moraine (core S2, 45°52′28″ N., 9°57′40.93″, 582 m a.s.l.). Both cores recorded, below the ploughing horizon, weathered silty clay whose thickness increase towards the moraine slope base, reaching 1.20 m in core S2 (#1). This is rich in manganese dioxide coatings and siderite concretions, likely hydromorphic features related to oscillating ground-water table into a fine-textured colluvial wedge at the moraine slope base. Peaks of magnetic susceptibility mark diagenetic ferromagnetic formations. The soil rests over a thick belt of unweathered, laminated fine carbonate sands (#2 Fig. 9.1). This lithofacies accumulated in a lake close to a carbonate bedrock detritus source. Given local sediment source and absence of any silt component, the lake was probably maintained by the aggradating fan delta (see c_1 in Fig. 3b). The sediment is poor in pollen, a circumstance indicating high sedimentation rate and low pollen

Table 1
Radiocarbon chronology of the studied sedimentary records.

Core/site acronym	Lab code	Strat. position/ depth (cm)	Analysed fraction	Technique	Dry height	$\delta^{13}C$ VPDB	^{14}C age BP	95% calibration range (cal BP)	Median probability	Reference
Succession of Cerete Basso peat bog										
CB section 93	C2-704	370	Peat	Conventional			3730 ± 95	4406–3850	4091	Orombelli and Ravazzi, 1995
CB section 93	GX-18984	529	Peat	Conventional		−27.9	3780 ± 80	4414–3930	4164	Orombelli and Ravazzi, 1995
CB section 93	GX-18985	575	Peat	Conventional		−28.5	4675 ± 90	5596–5058	5407	Orombelli and Ravazzi, 1995
CB section 93	GX-18987	616	Peat	Conventional		−28.7	5490 ± 185	6718–5895	6275	Orombelli and Ravazzi, 1995
CB section 93	GX-20587	675	Drifted peat	Conventional		−28.2	7015 ± 250	8330–7436	7857	Orombelli and Ravazzi, 1995
CB section 93	GX-19707	725	Peat	Conventional		−26.9	9385 ± 125	11,083–10,255	10,627	Orombelli and Ravazzi, 1995
CB section 93	GX-18988	857	Compressed peat	Conventional		−27.8	11,890 ± 155	14,065–13,387	13,733	Orombelli and Ravazzi, 1995
CB core S3	UBA-17221	874	Compressed peat	AMS	300	−29.1	11,921 ± 38	13,913–13,638	13,782	This paper
CB section 93	GX-18989	875	Compressed peat	Conventional		−28.2	12,260 ± 160	14,972–13,803	14,282	Orombelli and Ravazzi, 1995
CB section 93	GX-19424	1185	Fine organic	Conventional		−28.1	13,025 ± 150	16,511–15,096	15,752	Orombelli and Ravazzi, 1995
CB core S3	UBA-17222	1188	Compressed wood	AMS	58	−26.6	13,053 ± 43	16,375–15,192	15,756	This paper
CB core S3	UBA-17223	1336	Terrestrial herbaceous plant debris	AMS	10.2	−27.4	13,186 ± 41	16,611–15,405	16,085	This paper
CB core S3	UBA-17224	1433	*Salix* wood	AMS	310	−30.7	13,608 ± 44	16,948–16,571	16,770	This paper
CB core S3	UBA-18823	1617	Compressed wood	AMS	108	−27.6	13,884 ± 70	17,177–16,771	16,957	This paper
CB core S3	UBA-18822	1920	Compressed wood	AMS	64	−27.7	14,075 ± 60	17,469–16,874	17,115	This paper
Succession of Monte Cuca										
CUCA section	Ua-20964	Imbedded in diamicton underlying till	*Pinus sylvestris* bark	AMS	90	−24.3	21,550 ± 265	26,637–25,036	25,775	This paper (see Fig. 5)
Succession of Fonteno di Cerete										
FONTENO 95	C2-694	Uppermost peat layer underlying the breccia body	Peat	Conventional			12,580 ± 150	15,444–14,082	14,740	Orombelli and Ravazzi, 1995
FONTENO 95	C2-700	Lowermost peat layer underlying the breccia body	Peat	Conventional			13,590 ± 110	16,998–16,378	16,735	Orombelli and Ravazzi, 1995

influx in poorly vegetated conditions. The deeper part of core S1 (15–24 m depth) recorded massive gravelly sands, interpreted as fan delta flows. Clasts are composed by subangular limestone from local bedrock, while an allochthonous component (Permian siltite, micas) only occurs in the sandy matrix. This suggests denudation of a recently glaciated slope. A substantially different lithofacies was recorded in core S1 between 11.8–12.5 m and 15–16.8 m depth (#3). Here, bluish silts and silty-sands irregularly laminated with striate and faceted pebbles, both of local and allochthonous provenance (Fig. 9.2), suggest that the advancing glacier became in contact with the lake. This episode may have occurred at deposition time of moraine M4.

Drilling S2 reached a matrix supported, overconsolidated diamicton rich in striate pebbles, which was cored between 12.5 and 13.8 m depth (#4), interpreted as the buried slope of moraine M3. There is no evidence of buried soils, colluvium, weathering, or unconformity at the contact. This suggests no major hiatuses between glacial deposition and subsequent fan aggradation. Both depositional sequence and surface soil development are consistent with basin origin at an early stage of the last deglaciation.

5.5. The paraglacial phase. Evidence from the valley floor filling

A terraced sequence of alluvial fans developed on the floor of the middle Borlezza Valley (Fig. 4). After catastrophic floods in years 1977 and 1981 AD, the scarp of the lower terraced unit (number 8 in Fig. 4), highlighted a palustrine and alluvial alternation deposed between the Bølling–Allerød complex and the Bronze Age (Orombelli and Ravazzi, 1995, see S3 #8–9 in Fig. 7). Lacustrine deposits appearing at its base were dated 13,025 ± 150 ^{14}C BP (15.9–14.9 ka cal BP). This Lateglacial lake was related to a downstream dam built up by coalescing alluvial/debris flow cones radiating from opposite tributary gullies (see Fig. 4 – fan *1/fan *2). Support to this figure was provided by coeval ^{14}C ages from the body of fan *2 (section Fonteno 95; 45°50′57.67″ N, 10°00′06.59″ E; see Table 1 and Fo3 in Fig. 4), testifying that the main phase of cone aggradation occurred between about 16.6 and 15 ka cal BP. However, relationships between this phase of slope adjustment, deglaciation and vegetation recovery remained unravelled. Recently, two cores (site Cerete Basso, S3 and S4 in Fig. 4) were drilled looking for the presumed underlying till and to provide sedimentary, radiocarbon and vegetation data of the valley floor development in the early phases of deglaciation.

The 40 m-long Cerete Basso S3 drilling (45°51′26.60″ N, 9°59′27.00″ E, 464 m a.s.l., Fig. 7) cored the basal glacigenic sequence between 40 and 33 m depth. Unit S3 #1 is formed by overconsolidated and deformed lacustrine silt with alternations of fine overconsolidated matrix-supported diamicton bearing faceted-striate pebbles (lodgement till). It is covered by a matrix-supported diamicton with faceted-striate, allochthonous blocks

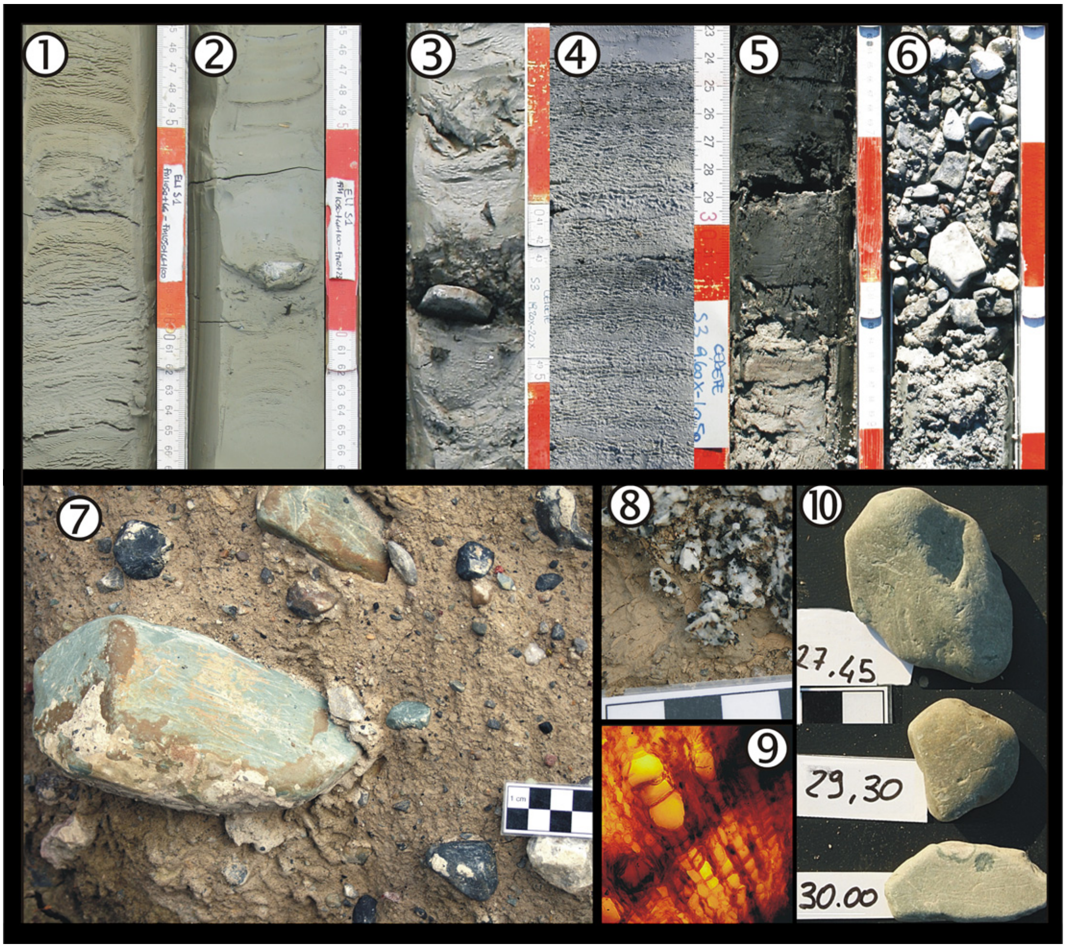

Fig. 9. Selected lithofacies, weathering features and ^{14}C dated samples encountered in cores and outcrops. 1 – Laminated Carbonatic Fine Sands, S2, 11.45–11.66 m depth (lacustrine); 2 – Massive silts imbedding striate pebbles, S2 11.80–12.00 m depth (ice-contact lacustrine); 3 – Coarse angular clast-supported gravel, rich in organic matter, imbedded in fine laminated silty sands, S3 19.2 m depth (massive slope deposit, likely an avalanche event, plunged into lacustrine sequence). The organic fraction consists of partly humified debris and twigs of terrestrial plants, rich in pollen of tundra herbs (*Parnassia*) and was dated 14,075 ± 60 ^{14}C BP; 4 – Partly biogenic and biochemical laminated carbonate mud rich in charophyte stems and gyrogonites (calcareous gyttja), S3, 10.24–10.40 m depth (biochemical lacustrine); 5 – Alluvial fine sands passing upward to compressed peat, S3, 5.60–5.92 m depth (alluvial and palustrine); 6 – Coarse gravel, local petrofacies, S3, 6.10–6.42 m depth (alluvial channelled deposits); 7 – Overconsolidated diamicton at the base of the section of Fonteno di Cerete, morainic ridge M5 (lodgement till); 8 – Detail of (7) showing destructured tonalite in silty matrix; 9 – *Salix* wood, dated 13,608 ± 44 ^{14}C BP; 10 – Striate pebbles of Permian siltite and core S3 depth.

(S3 #2, melt-out till). Average magnetic susceptibility is higher in S3 #1, in agreement with a significant content of micas and/or magnetite released by glacial abrasion. The S3 #1 unit outcrops 1 km upstream along the valley floor, where a long varved sequence can be observed (Va in Fig. 3b, see Appendix 1 for details). S3 #3 and S3 #4 are dominated by massive/irregularly laminated (never varved) fine carbonate sand lithofacies, free of any organic or biogenic component. Alternations of poorly sorted gravels still include either clast-supported levels with subangular clast of local petrofacies and occasional horizons with still allochthonous striate pebbles (Fig. 9.10). This lithofacies association may relate to an ice-dammed valley floor mostly supplied by local fans and occasionally fed by flow till (upward in core succession until 27.45 m depth). Irregularly laminated fine sands represent a thick and fast infill of shallow, ephemeral lakes, subjected to irregular water-level oscillations, as marked by immediate replacement of gravels, suggesting fan progradation. The uppermost S3 #4 is polliniferous and includes a thin layer of coarse angular clast-supported gravel, rich in organic matter, i.e. a massive slope deposit – likely a single avalanche event – plunged in detrital lake sediments (Fig. 9.3). The organic fraction consists of partly humified debris of terrestrial plants, rich in herb pollen of wet-limestone tundra (*Parnassia*). A woody twig was dated 14,075 ± 60 ^{14}C BP (17.2 ± 0.3 ka cal BP). S3 #5 is composed by a laminated silty sand carbonate lithofacies, poor of allochthonous detritus, likely the infill of a larger lake whose threshold is definitively independent from the glacier. It spans about five centuries (17.2 ± 0.3 ka–16.7 ± 0.2 ka cal BP). S3 #6 records a terrestrialization of the site, as shown by development of hydromorphic soils, expansion of peat bogs as well as of cembran pine (see Section 5.6). S3 #6 spans about one millennium between 16.7 ± 0.2 ka and 15.7 ± 0.4 ka BP. S3 #7 is formed by a partly biogenic and biochemical laminated carbonate mud, rich in charophyte stems and gyrogonites (calcareous gyttja), suggesting a lake with reduced bed-load discharge. As already mentioned (Orombelli and Ravazzi, 1995), this lake originated after dam by coalescing alluvial/debris flow fans.

A summary chronostratigraphical interpretation of the complex succession so far examined derives support from an age–depth model (Fig. 8). The oldest available age at 19.2 m depth allows setting a minimum age for the local deglaciation of the middle Borlezza Valley at 17.2 ± 0.3 ka cal BP. A persistently high sedimentation rate of the valley floor through fan aggradation and fast infill of ephemeral lakes is suggested by the subsequent, radiocarbon-dated history (e.g., 11.8 mm/year in average for S3 #4–

Table 2
Pollen spectrum from the lacustrine silt embedding the tree bark dated 21,550 ± 265 ^{14}C BP. Sample for pollen was collected 10 cm apart from the dated plant remain (analysis Roberta Pini).

	Grains number	Pollen %	Concentr. (grains/cm^3)
Pinus sylvestris/mugo	26	25	7
Pinus cembra	2	2	1
Ephedra fragilis type	1	+	+
Alnus glutinosa type	1	+	+
TOTAL Trees and shrubs	30	29	8
Gramineae	1	+	+
Aster type	3	3	1
Cichorioideae	56	54	16
Caryophyllaceae	3	3	1
Umbelliferae	6	6	2
Polygonum aviculare type	1	+	+
Cruciferae	1	+	+
Ranunculus	2	2	1
TOTAL Herbs, chamephytes	73	71	21
Gelasinospora	1	+	+
Selaginella selaginoides	5	5	1
Pollen sum	103		
Charcoal particles >10 μm	2		
Pollen concentration (grains/cm^3)	29		
Charcoal concentr. (particles/cm^3)	<1		

5). This figure is consistent with the general model of paraglacial evolution during the course of deglaciation (Church and Ryder, 1972; Ballantyne, 2002), but bears some peculiarities given that the end-moraine system is “reversed” (i.e. projected upstream). The studied succession misses a glacifluvial phase, while the valley dam is maintained until ultimate deglaciation of the complete valley reach. This dynamics allows excluding major erosional phases. Assuming a constant sedimentation rate down to the glacial/proglacial boundary at 33 m, the time allowed is 1186 yrs, or 694 yrs down to the last evidence of flow till/ice proximity. This translates in a deglaciation age of 18.3 ± 0.2 or 17.8 ± 0.2 ka cal BP respectively.

5.6. Early Lateglacial vegetation history

The Lateglacial succession filling the valley floor so far examined is polliniferous since the deepest dated layer shortly post-dating deglaciation, although pollen concentration remains low until 16.5 ka cal BP. Furthermore, intervals of coarse sediment could not be analysed. Nevertheless, it provided the chance to date the vegetation history in the Italian Alps between 17 and 15 ka cal BP (Fig. 10).

The deeper pollen zone (Cer1) records a basal phase of open vegetation, marked either by steppic taxa of dry places (*Artemisia*,

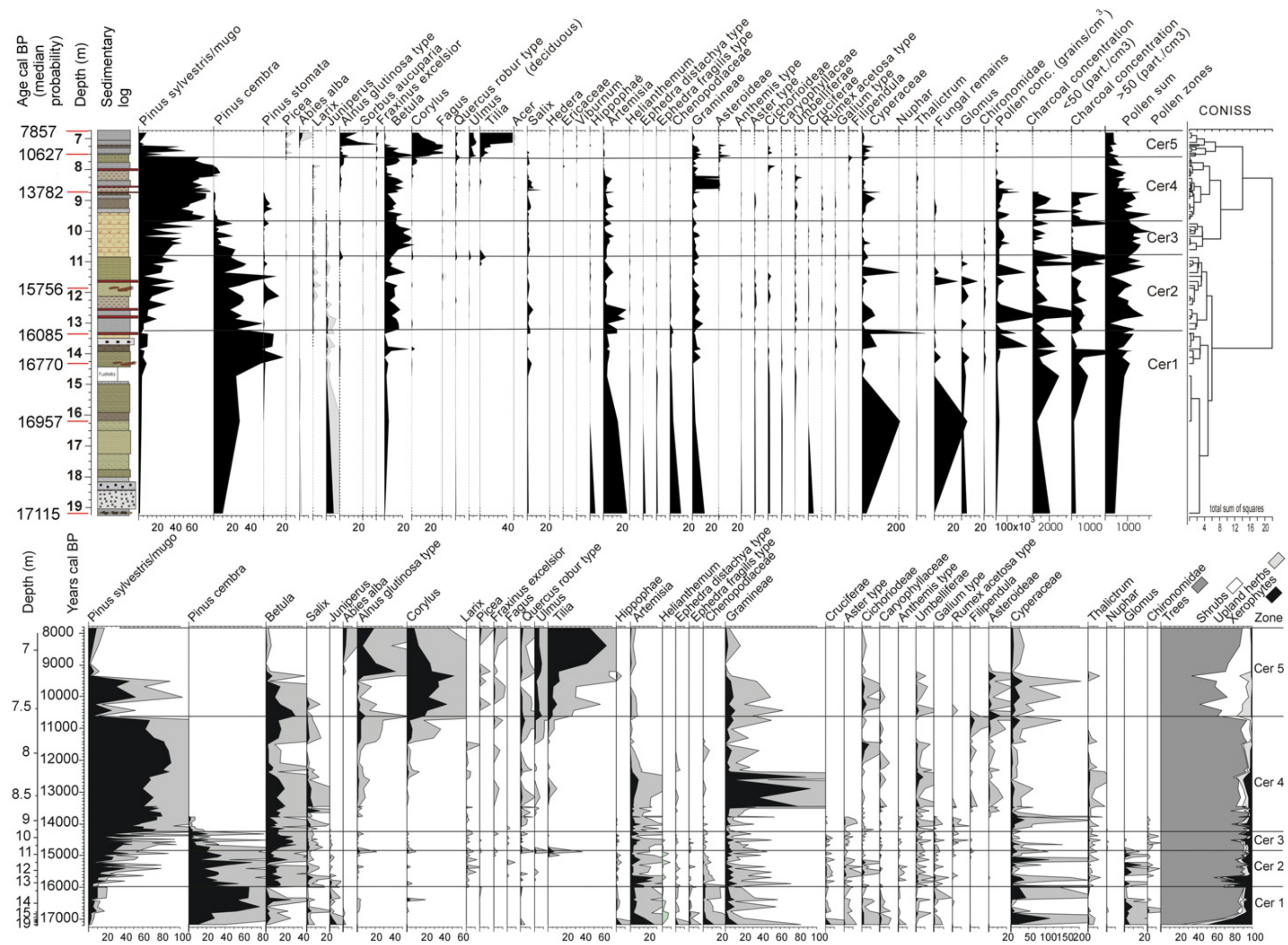

Fig. 10. Upper panel: Pollen record of the Cerete Basso S3 core (% selected curves) plotted on a stratigraphic scale. *Picea*, *Larix*, *Abies*, *Juniperus* are also shown by five times exaggerated curve (in grey). See Fig. 7 for the key of the stratigraphic log. Lower panel: Pollen record of the Cerete Basso S3 core (% selected curves) plotted on the age scale. Five times exaggerated curves in grey.

Helianthemum, *Ephedra fragilis* type, *Ephedra distachya* type, and chenopods) and by some plants of aquatic–wetland habitats. Remarkably, since the onset of the record, *P. cembra*-type maintains values 5/7 times higher than *P. sylvestris-mugo*. Part of the latter pine pollen type may have been transported over long distances, which is not the case for *P. cembra*, whose representation at the end of Last Glacial Maximum (18 ka cal BP) is negligible at the Italian Alpine foothills (Schneider, 1978; Wick, 1996; Monegato et al., 2007; Pini et al., 2009, 2010). Pine stomata provide further support to the occurrence of cembran pine in the area. Climate conditions and the paraglacial activity are likely to have prevented tree expansion.

Peat and hydromorphic soils in S3 #6 point to terrestrialization and stabilization of the valley floor. In turn this is coeval to the marked expansion of cembran pine observed at 16.7 ± 0.3 ka cal BP, about one millennium after the estimated age of local deglaciation. Cembran pine may have occupied habitats close to the valley floor, which were previously inaccessible due to persistent fan aggradation and water-saturated parent material subjected to seasonal freezing. Although dwarf pine (*Pinus mugo*), Scottish pine (*P. sylvestris*) and Larch (*Larix decidua*) withstood the last glaciation in the mountains of Lombardy or at their foothills (Ravazzi et al., 2004), they did not replicate the observed, early expansion of cembran pine. Today, cembran pine may withstand extreme continental and waterlogged habitats (Schulze et al., 2002) thanks to its higher drought resistance at low temperature compared to other boreal trees (Anfodillo et al., 1998). This feature places its bioclimatic envelop at the extremes of the forest landscape (Casalegno et al., 2010). We suggest that the success of cembran pine was triggered by stabilization of the paraglacial valley floor landsystem, under persistent pressure of a dry-continental climate.

At about 16.0 ± 0.2 ka cal BP, a cembran pine withdrawal is mirrored by *Artemisia* and tree-*Betula* expansion (zone Cer2). This event is not coeval to a significant sedimentary change, neither may be predicted by a model of progressive forest succession on a deglaciated terrain, given that cembran pine is a late-successional species (Ellemberg, 1988; Matthews, 1992). Instead, the observed pine replacement by open vegetation and by pioneer trees points to a recessive dynamics. A hypothesis of climate forcing may be explored (see Section 6.4).

Between 15.5 and 15.0 ± 0.3 cal BP, a *P. sylvestris mugo*/*Betula* expansion and a further decline of cembran pine (zone Cer3) is recorded within biochemical lacustrine sediments (S3 #7). As already mentioned, this lake was dammed by alluvial/debris flow cones radiating from opposite tributary gullies. Both age and the palaeobotanical context indicate that fan development was not a primary paraglacial feature, i.e. shortly post-dating deglaciation (Ballantyne and Benn, 1994; Harrison and Winchester, 1997; Curry et al., 2006; Mercier et al., 2009), nor occurred before forest vegetation becoming established in the valley. Two concurrent events may explain this late fan activity and vegetation development: (i) a delayed downcutting of the valley floor until ultimate deglaciation of the complete valley reach, i.e. a phase of secondary paraglacial activity promoted by fluvial processes (Ballantyne, 2002); (ii) a moisture increase, enhancing gelifluction, snow avalanches, and favouring trees withstanding wet climates, i.e. a climate perturbation after the termination of the initial period of paraglacial response. Increasing rainfall between 16 and 15 ka in SE-Alps is shown by oxygen isotope values from speleothems (Frisia et al., 2005) and is coeval to an expansion of open conifer forests at the southern alpine foothills (Vescovi et al., 2007; Monegato et al., 2011).

Finally, a mass expansion of *P. sylvestris-mugo* at 14.3 ± 0.3 ka cal BP (zone Cer 4) is consistent with the regional vegetation history, a sharp pine rise being recorded throughout the lower altitude Alps at Bølling onset (Finsinger et al., 2006; Vescovi et al., 2007; Ravazzi and Vescovi, 2009). The onset of Bølling–Allerød biostratigraphic complex in the Italian Alps was marked by a sharp warming and by a further rainfall increase (Frisia et al., 2005; Vescovi et al., 2007).

6. Discussion and synopsis

6.1. Identification of the LGM maximum extent in the end-moraine system of Clusone

The stratigraphic and palynological record studied in the end-moraine system of Clusone/Borlezza Valley in the Italian Alps provided evidence for a complete cycle of glacial advance, culmination and withdrawal, which is constrained by radiocarbon ages within the early LGM and early phases of the Lateglacial. In addition, weathering profiles preserved on stable surfaces of relevant deposits display similar time-markers, e.g., clay illuviation and rubefaction development, while the morphological setting of morainic arcs M1 to M5 can be assigned to a complex of maximum advance and recessional moraines within one major glacial episode. Our results add precision to the chronostratigraphic interpretation provided by early scholars, but contradict the setting proposed by a recent survey of the Italian Geological Map (Ferliga, in press). Contrary to Bini et al. (2009) and Ferliga (in press), the LGM maximum extent of the Oglio Glacier was not limited to the trunk valley (Val Camonica); instead the glacier flux propagated upstream to the end-moraine of Clusone (Figs. 2 and 11).

The study by Ferliga (in press) was based on landform preservation, weathering degree, and stratigraphic position, but was not supported by geochronometric or biochronological determinations on the Oglio Glacier system (see Figs. 1 and 2). The present study highlighted severe limitations to chronostratigraphic inferences based solely on pedofeatures, weathering, unconformities and geometric setting of Quaternary deposits. We observed that even gentle slopes of morainic arcs may be affected by colluvial accumulation of fine residual clay to intermorainic depressions; besides, geometry of rock units and rank of unconformities may not be apparent from natural exposures in a valley landsystem affected by glacial–paraglacial processes. In the present paper, these limitations were overcome by an integrated stratigraphic analysis from coring intermorainic basins and valley floor fills.

6.2. Age and palaeoenvironment at the time of the Last Glacial advance

The advancing glacier dammed a deep, ice-contact lake, infilled by a regular varved sequence, today overconsolidated and deformed after subsequent glacier overriding (site Va in Fig. 3b, and unit S3 #1 in Fig. 11A). While we could not obtain organic material from varves, the bark fragment dated at Cuca section establishes that the Last Glacial advance took place soon after 25.8 ± 0.8 ka cal BP. Note that the Cuca site is 2.5 km from the LGM culmination position. This age compares very closely to the early glacial culmination within the two-fold LGM advance in the Tagliamento amphitheatre, SE Alps (Monegato et al., 2007; Fig. 12) and to the onset of glacifluvial damming of the Lake Fimon, Italian foothills (Monegato et al., 2011). An earlier onset of glacifluvial activity in Central and Eastern Alps is shown by ^{14}C and luminescence ages of around 30 ka cal BP for direct peat burials by outwash deposits and loess (Schlüchter, 2004; Preusser et al., 2007; Ivy-Ochs et al., 2008; Starnberger et al., 2011; Ravazzi et al., 2012; Fig. 12) or by direct burial from advancing ice (Fliri, 1988; de Graaff, 1992). The time lag from 30/29 to 26 ka cal BP may actually represent the interval spanned by the advancing phase from the outer reach of Alpine valleys to the maximum culmination in the foreland amphitheatres.

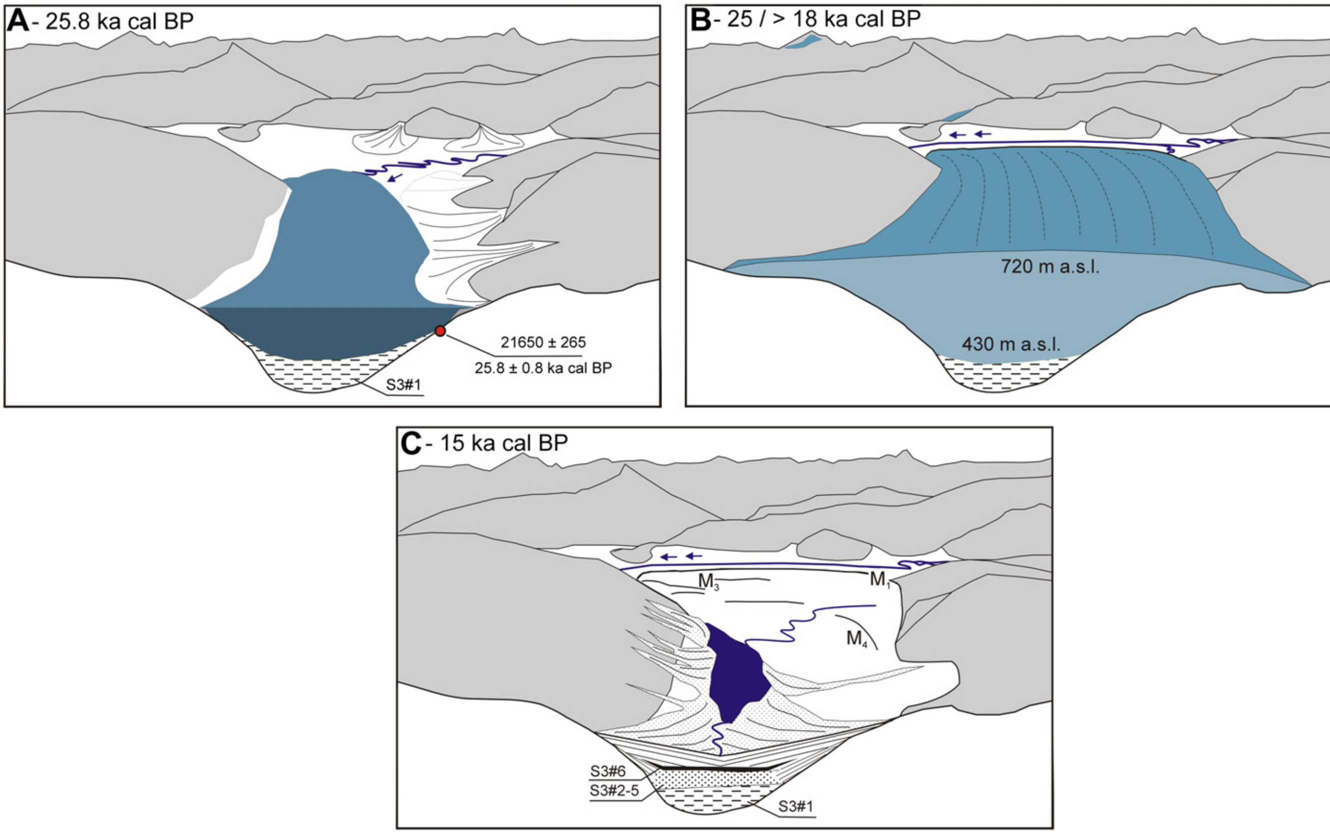

Fig. 11. Three main steps of the LGM-deglaciation development in the studied area. A – During the glacier advance, dated shortly after 25.8 ± 0.8 ka cal BP, a deep, ice-dammed lake developed in the Borlezza Valley; B – At the maximum culmination, the glacier overrode the anaglacial lacustrine deposits and formed the Clusone end-moraine system, diverting drainage westward (25–>18 ka BP); C – After ultimate deglaciation of the entire valley reach, coalescing debris cone dammed a last Lateglacial lake with biochemical sedimentation (15.5–15 ka cal BP).

Pollen data and the dated bark itself support the occurrence of *P. sylvestris/mugo* and *P. cembra* at middle altitude on sunny slopes, within the inner belt of Italian Alps. This is consistent with earlier macrofossil finds testifying to conifer trees survival in the lower altitudes of the Italian and Austrian Alps at the time of the Last Glacial advance (Fliri et al., 1970; Bortenschlager and Bortenschlager, 1978; Ravazzi et al., 2004).

6.3. Age of deglaciation

The age of deglaciation at the Cerete Basso S3 drilling site, 2.5 km within the end-moraine limit, has been constrained between a maximum age of 18.3 ± 0.3 and a minimum age of 17.2 ± 0.3 ka ago. There, the glacier was about 300 m thick at maximum culmination (see Fig. 11B). Site Cerete Basso S3 is located within the main morainic arcs M1–M4 (see site B in Fig. 2), thus its deglaciation age relates to an intermediate step, the Oglio glacier having lost 18% of its total linear length. A comparable age of 18.2 ± 0.3 ka cal BP has been obtained (Lauterbach et al., 2012) for the withdrawal of the main branch of the Oglio glacier from the middle Iseo lake (see site A in Fig. 2), corresponding to a 19% loss of the glacier total linear length. Consistent ages of 18.5–17 ka cal BP for the deglaciation of the valley/intermorainic Lakes Como, Lugano, Zürich and Ragogna (Lister, 1988; Niessen and Kelts, 1989; Comerci et al., 2007; Monegato et al., 2007) correspond to 9%, 10%, 21% and 22% loss of total glacier length respectively.

The deglaciation of major Alpine lakes stepped about two thousand years after the initial withdrawal from the outermost Würmian morainic ridges, currently dated between 23.1 ± 0.4 ka cal BP and 20.8 ± 1.3 ka cal BP (Monegato et al., 2007; Ivy-Ochs et al., 2008).

6.4. Paraglacial development, early forest pioneering and climate change

A synopsis of the glacial, paraglacial history and of summary Lateglacial vegetation record plotted on age axis (Fig. 12) allows for synchronization with the broader climatic scenario.

Since the Clusone end-moraine system is projected upvalley, downcutting was delayed until complete deglaciation of the valley reach, allowing preservation of a complete sedimentary record produced by primary paraglacial activity in the first millennium (ca 18–16.7 ± 0.3 cal BP) after deglaciation. Even afterwards, the surface discharge remained small, thus the evolution of the valley floor was controlled by slope-damming processes. Hence, the valley fill misses the glacifluvial phase which normally ends the proglacial sequence. This favoured an early stabilization of the floor and promoted the pioneering of cembran pine as early as 16.7 ± 0.3 ka cal BP. This is the first evidence of early tree pioneering deglaciated terrain as a response to deglacial/paraglacial activity in an inner valley floor of the European Alps. So far, the documentation on Lateglacial cembran pine development at low altitude in the Italian Alps was either undated, or restricted to the interval <16 ka BP (Schneider and Tobolski, 1985; Tinner et al., 1999; Hofstetter et al., 2006; Filippi et al., 2007; Vescovi et al., 2007). Only at lake site of Ragogna in SE-Alps, an expansion of cembran pine at about 17.5 ka cal BP shortly postdated the end of fluvioglacial activity (Monegato et al., 2007), thus suggesting a correlation to a decline in paraglacial activity. At Lake Jeserzersee (Austrian Alps) a cembran pine increase has a modelled age of 16,600 ± 150 yr cal BP and is related to a diatom-inferred warming phase (JPZ 5, Schmidt et al., 2012). Surprisingly, whenever the

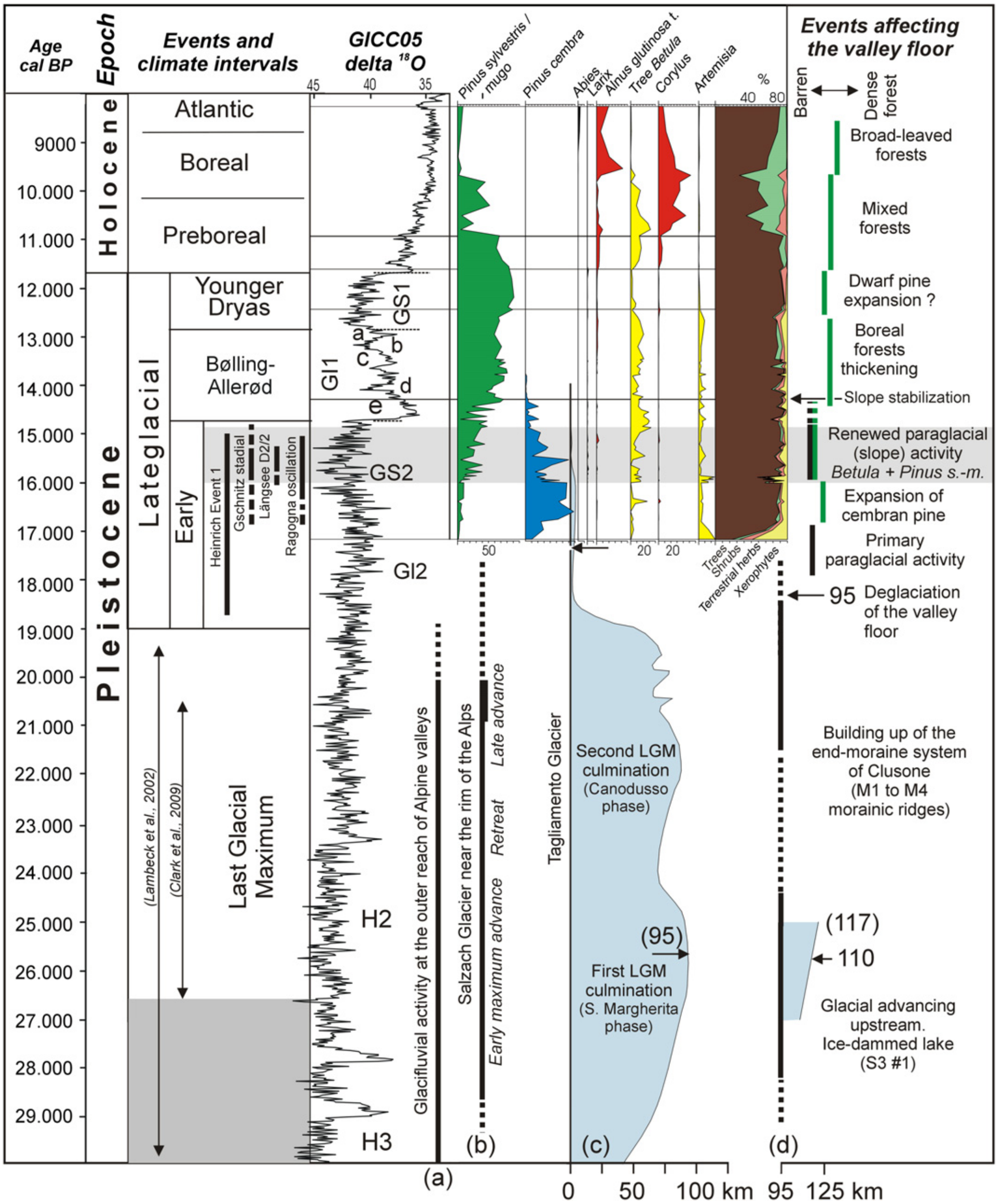

Fig. 12. Chronological synopsis of the LGM and the Lateglacial events affecting evolution of the studied valley floor in the Oglio glacier system and synchronization with the Alpine climatic scenario. *In the lower panel* (30–17 ka cal BP): glacier oscillations are given as changes in linear distance (km) from the origin of the glacier. Thick black bars indicate a glacier *in situ* at a given time-distance. A few recently dated case studies of end-moraine systems in the Alps (Fig. 1a for location) are shown for comparison: (a) aggradation of glacifluvial outwash rivers (Tagliamento, Fimon, Rhône, Rhein, Salzburg); (b) Salzach Glacier, Austrian Alps (Starnberger et al., 2011); (c) Tagliamento Glacier, Italian Alps (Monegato et al., 2007); (d) Oglio Glacier (this work). *In the upper panel* (17–8 ka cal BP): the cumulative pollen record and selected % pollen curves, obtained from the valley floor infill, are plotted on age axis. A grey belt spanning from 16.05 ± 0.6 to 14.6 ± 0.5 ka cal BP underlines a perturbation recorded by renewed paraglacial activity and by a cembran pine decline (pollen zone Cer 2), compared with the contemporary climatic events occurred in the Alps. Chronology and duration of global LGM refer to Lambeck et al. (2002) and Clark et al., (2009); Heinrich Event to Stanford et al. (2011); Ragogna oscillation to Monegato et al. (2007); Längsee cold phase to Huber et al. (2010); Gschnitz stadial to Ivy-Ochs et al. (2006).

cembran pine played a major role after deglaciation, it did not leave a trace in the subsequent forest expansion at the onset of the Bølling–Allerød biostratigraphic complex. Widespread steppe plants also declined; some of them became extinct later on (*E. fragilis* type).

Between 16.0 ± 0.6 and 14.6 ± 0.5 ka cal BP, enhanced slope activity, a cembran pine withdrawal coupled with *Artemisia* and tree-*Betula* expansion (pollen zone Cer 2) suggest that an external perturbation renewed paraglacial response, frost action and snow avalanches. The initial perturbation occurred at least one millennium after deglaciation of the formerly glaciated landscape. Similar delayed vegetation recessions, occurred millennia after the local deglaciation, were recorded at three lakes in south-eastern Alps (Ragogna, Längsee, Jeserzersee), and interpreted as the effect of secular phases of cooling and/or increasing moisture, compared to the semiarid continental conditions prevailing previously in the early Lateglacial. Modelled ages of 16.5 ± 0.8 for the onset of the “Ragogna oscillation” (Monegato et al., 2007), 15.75 ± 0.8 for the onset of Längsee D2/2 phase (Huber et al., 2010) and 16.25 ± 0.25 for the onset of a late cold phase at Lake Jeserzersee (Schmidt et al., 2012) would place these cold/wet events at the end of the Heinrich Event 1 (Stanford et al., 2011). Furthermore, the onset of a phase of glacial advance in the Alps, identified as the Gschnitz stadial, has been dated 15.9 ± 1.4 ka cal BP (Ivy-Ochs et al., 2006). Actually, placing these Alpine events in a broader climate scenario would require high-precision ages, currently unavailable. Despite a significant synchronism (see grey belt in Fig. 12), their precise timing has not been determined yet, thus time-correlations and teleconnections remain speculative. The speleothem isotope oxygen record suggests that the influence of Atlantic air masses on the climate of the southern side of the Alps had started increasing as early as 16–15.5 ka cal BP (Frisia et al., 2005). This may have favoured replacement of early colonizers, loving continental climates, which survived the last glaciation in the unglaciated mountains, with vegetation more adapted to oceanic moisture.

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Appendix 1. Field descriptions of weathering profiles selected either on stable and eroded surfaces of M1–M5 morainic arcs; end moraine system of Clusone.

§ Profile developed on a terraced, stable surface on glacifluvial deposits related to M5 morainic arc (45°51′05.25″ N, 10°00′15.12″; see ***Fo2*** in Fig. 4).
A continuous argillic horizon 5 YR, 20–40 cm thick, rests over crudely stratified, weathered Gm (matrix-supported gravels) with subordered layers of diamicton. Pebbles 10–30 (80) cm are mostly faceted but surfaces are not striated. This is interpreted as a glacifluvial sequence with debris flows intercalations. The decarbonation front is recorded at 1.5–2 m depth. Carbonate leaching is commonly accompanied by immediate, and complete, texture destructuration of granitoid blocks (Fig. 9.8), and development of a cortex on carbonate pebbles, while Permian siltites and tuffs do show manganese dioxides or clay coatings.

§ Profile observed on the ridge of M3 morainic arc, eroded at the outwash head (45°51′57.00″ N, 9°58′47.20″E see La Spessa fan head in Fig. 3b), shows an almost unweathered Dmm (matrix-supported diamicton), with mostly allochthonous striate clasts, including metric blocks of tonalite from the Adamello Massif, interpreted as melt-out till. This section is truncated, therefore its low weathering degree is not considered significant for stratigraphic purposes.

§ Section cropping along the left bank of the Borlezza River, upstream to the village of Cerete Basso, locally called “Valleggia section” (45°52′25.83″ N, 9°57′15.73″E; see ***Va*** in Fig. 3). This section exposes a 22 m-thick succession of lodgement till and glaciolacustrine deposits. From base to top: unit 1 – Laminated silt, formed by millimetric rythmites, folded, containing dropstones (glaciolacustrine, 1.5 m visible); unit 2 – Shear-stressed, over-consolidated Dmm, containing faceted, striate pebbles (lodgement till, 2 m); unit 3 – Laminated silt, formed by sets of millimetric rythmic couplets, strongly contorted by folds and thrusts-folds, fold noses trending upstream, containing dropstones (partially varved glaciolacustrine, 15 m visible); unit 4 – Laminated silt containing lenses of open-work, monogenic breccias (lacustrine with slope pulses; 3.5 m). The top section is truncated.

According to lithological properties, lithofacies, magnetic susceptibility, stratigraphic position and geometric relationships, these deposits belong to an ice-contact lake formed during the Last Glacial advance (see Section 6.2 and Figs. 7 and 11A).

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