New dates of a Northern Italian loess deposit (Monte Orfano, Southern pre-Alps, Brescia)

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Abstract

Purpose

Loess in Northern Italy has been usually considered deposited during the MIS 4-2 period, which corresponds to the last Pleistocene glacial cycle. In particular, no absolute dating evidenced loess depositions older than ca. 89 ka. We investigated two strongly rubified soil profiles in the southern margin of the Alpine range in Lombardy to prove their aeolian origin and age of formation.

Methods

We analysed the granulometry of all genetic horizons of these strongly rubified soils and a total of 8 samples were collected for luminescence dating purpose.

Results

Most of the analysed soil horizons were dominated by silt and were characterized by the s-shaped granulometric curve, typical of loess materials. A particularly high clay content evidenced a strong weathering degree. A deep horizon was particularly clay-rich and it was interpreted as a typical Terra-Rossa horizon. Luminescence dates increased with depth, reaching 122 ka for the deepest loess layer and 453 ka (minimum age) for the Terra-Rossa horizon.

Conclusions

The deepest observed loess layer represents the oldest quantitatively dated aeolian deposition in Northern Italy up to now.

Keywords

Loess; MIS1-MIS6; OSL-IRSL dating; Terra-Rossa soil

Declarations

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1 Introduction

Loess is a prevalently silty sediment transported by wind, usually in glacial periods, during which the grinding action of glaciers on the enclosing rocks was active and fluvio-glacial sedimentation occurred on large surfaces in proglacial braided stream beds; these barren areas acted as deflation sources for great amounts of silty materials, which could be deposited in dust traps, which were more vegetated, stable surfaces (Pye 1995; Li et al. 2020). After deposition, loess was subjected to erosion, solifluction, cryoturbation and pedogenesis (Muhs and Bettis 2003).

A large loess belt covers much of mid-latitude Eurasia (Haase et al. 2007). The presence of loess in northern Italy has long been historically neglected or underestimated (e.g., Haase et al. 2007; Muhs 2013) in the international scientific literature, but the presence of a loess basin between the Alps, the Apennines and the Dalmatian coast is well known (Cremaschi 1988). Usually, this loess cover is considered to be deposited between the Würm alpine ice stage and the Late Glacial, between MIS 4 and MIS 2 (e.g., Costantini et al. 2018; Cremaschi et al. 1990; Ferraro 2009; Zhang et al. 2018). Most dated loess deposits in the Po plain (fig. 1) show that aeolian depositions have been active since 60 ka, at the onset of full glacial conditions in MIS 4 (Cremaschi et al. 2015). A more ancient loess layer on an isolated hill in the central Po Plain in Lombardy had an OSL date of 89 ± 9 ka (MIS 5b), while nearby alluvial sands and gravels were slightly more ancient, dated back to 107 ± 13 ka - MIS 5d (Panzeri et al. 2011). Much older, dated loess covers are widespread in other European areas, such as Germany (Kreutzer et al. 2012), Austria (Preusser and Fiebig 2009), and Serbia (Marković et al. 2011).

Some northern Italian loess sections, however, have been attributed to the Middle Pleistocene or even earlier periods, but no absolute dating is available in the literature. For example, Busacca and Cremaschi (1998), based on pedogenic and magnetostratigraphic evidences, attributed ca. 400.000 years of age to some loess layers in the southern Po Plain margin. In the Lanzo alluvial fan (Torino), Billard and Orombelli (1986) attributed some loess sections to the 5th glacial stage, corresponding to 1.8-1.0 Ma BP (MIS 63-23). Recently a thick and strongly rubefied silty deposit was locally found on some slopes of Monte Orfano, an isolated hill on the northern margin of the Po Plain, a few km south of Lake Iseo (Brescia province, Lombardy). Our aim was, thus, to check if this silty deposit was actually loess (using granulometric analysis) and to date its deposition using luminescence methods both on quartz (OSL) and feldspars (IRSL, Infrared Stimulated Luminescence).

2 Material and methods

2.1 Study area characterization

The Monte Orfano is an isolated relief, located on the northern edge of the Po Plain, south of Lake Iseo, west of Brescia and east of Bergamo, Lombardy (fig. 1). Its ridge has an elongated shape in the prevailing WNW-ESE direction and has a maximum elevation at 452 m a.s.l. The maximum cross width is 1,200 m. The northernmost point of the mountain has a latitude 45°35'40.5" N and a longitude 9°56'12" E; the southernmost one is 45°33'49.9" N and 9°59'08.6" E. The occurrences, although discontinuous, of loess cover, with a thickness up to a few meters, and Terra Rossa soils make the site interesting for the study of the Quaternary paleoenvironments of Northern Italy, the Po Plain and the Alpine and Apennine fringes.

The hill is composed of a single geological formation called "Conglomerato di Monte Orfano" (MOC), an orthoconglomerate with massive to poorly-bedded arrangement of pebbles and cobbles of limestones, marly limestones, chert, cherty limestones, radiolarites, dolostones, sandstones and few volcanic fragments, with carbonatic cement. (Sciunnach et al. 2010). It was recently dated to the Late Oligocene (Sciunnach et al. 2010), while in the past its age was believed to be between Early and Middle Miocene (Vecchia and Cita 1954). The clasts, mainly derived from sedimentary Norian and Aptian formations, were deposited in a shallow-marine fan delta during the uplifting front of the Southern Alps, without significant lithological variations in the different sedimentary strata with the exception of rare intercalations of decimetric layers of sandstones and marls (Sciunnach et al. 2010).

The climate (1960-1990 data) in nearby Chiari weather station (located an elevation of 148 m a.s.l.) is characterized by an average yearly temperature of 13.5°C, a total mean precipitation of 946 mm, with equinoctial maxima and a primary winter minimum and a secondary summer one. The moisture regime for the described soils, calculated with the Newhall method (Newhall 1972), is Udic according to Soil Taxonomy rules (Soil Survey Staff 1998). The sites of the two profiles are covered by *Castanea sativa* Mill. mixed with *Robinia pseudoacacia* L. woodlands, presently unmanaged but coppiced in the past. While other sectors of the hill are terraced, in our sites there are no terrace remnants, it is thus unlikely that the sites were ever used for agriculture.

2.2 Field and laboratory methods

Two soil sections were investigated: the loess section (LS) located at 410 m a.s.l., with latitude 45°35'24.9" N and longitude 9°56'57.6" E; and a Terra Rossa soil profile (TR) located at 310 m a.s.l., 45°34'39.70'' N, 9°58’31.83” E. The LS section was opened with an excavator, down to a depth of ca. 4.8 m, in the upper part of a west-facing slope. The TR section was located in a middle steep slope facing north-east. Different soil horizons were recognized and described (table 1, fig. 2), according to the FAO (2006) guidelines. Soil samples were taken from the main pedogenic horizons, treated with 20% H_2O_2 solution for 3 days (until complete disappearance of bubbles) and, after adding a 5% Na-hexametaphosphate solution, particle size was measured by sieving and sedimentation using a hydrometer according to ASTM standards (ASTM D 422). The analysis was carried out in the Pedology lab in the DISAT, Milano Bicocca University. The results were shown as cumulate curve using a base 2 cologarithmic scale for equivalent diameters (Krumbeinϕ scale).

2.3 Luminescence measurements

Optically stimulated luminescence dating methods can be used to estimate the time elapsed since buried sediment grains were last exposed to daylight. Luminescence has been successfully applied, in the last decade, on loess and Terra Rossa-like sequences in Italy (Andreucci et al. 2012; Zucca et al 2014) and Europe as well (Guerin et al. 2017; Zhang et al. 2018; Stevens et al. 2020). It is based on the measurement of the electric charges trapped in mineral grains since the time of the sediment deposition, as a consequence of the irradiation due to the natural radioactivity field. The upper age limit is normally controlled by saturation of the luminescence signal. Because the natural OSL signal from quartz extracted from most of the Monte Orfano samples was close to the limit of saturation, the K-feldspars were chosen as dosimeters in luminescence dating. K-feldspar IRSL signals, in fact, normally saturate at higher doses than quartz (Wintle and Murray 2006).

Samples for OSL analysis were collected using specific core samplers able to get undisturbed soil materials at least 30 cm from the vertical surface of the soil pit, at different depths. In particular, we collected undisturbed soil samples at six depths in LS, and two in TR profile (table 2). In order to separate quartz from K-feldspars (grains size 180-250 µm), samples were prepared following the conventional procedure (Lang et al. 1996).

To measure the annual radiation dose provided to the sample from the radioactive elements surrounding it, Th and U concentrations of each sample were measured with total alpha counting using ZnS scintillator discs (Aitken 1985), assuming a concentration ratio Th/U equal to 3. Content of ^{40}K was hypothesized from the total concentration of K measured with flame photometry. Attenuation of the beta dose (Bell 1979) and a probable water content of the loess were taken into account (table 2) while alpha contribution was eliminated by an HF etching (10%; 30 minutes). The cosmic ray contribution to the final dose rate was based on Prescott and Hutton (1994). The ^{40}K internal radioactivity on K-feldspar grains contributing to the final dose rate was calculated assuming a K content of 12.0 ± 0.5% (Huntley and Baril 1997).

The measurements were performed with an automated luminescence system (Risø TL/OSLDA-20) equipped with a 90Sr/90Y beta source delivering 0.11 Gy/s (± 3%) to the sample position. Feldspars IRSL was stimulated by an array of IR LEDs (830 ± 10 nm; 360 mW/cm2) and detected through a blue filter (Schott BG39/Corning 7-59 filter combination). The Single-Aliquot Regeneration (SAR) dating protocol (Murray and Wintle 2000) was applied using different protocols to analyse samples along the studied profiles. In particular, from the top of the profiles downward, the postInfrared-IRLS (pIRIR) at 150°C protocol was used to analyse TR25 and LS 40 samples (Reimann and Tsukamoto 2012), while the pIRIR at high temperature (290°) was selected for TR100 and LS 120 samples (Buylaert et al. 2012). For all the other samples (LS 170, LS 270, LS 350 and LS 440) the Multi-Elevated-temperature MET-postIRIR procedure was applied (Li and Li 2011) using multi-steps of IRSL measurements with increasing stimulation temperature from 50 to 250°C. At high stimulation temperatures (200 and 250°C), the MET-pIRIR Equivalent Dose (De) reached a plateau and these values were used for age determination. For all samples, the measured residual doses were subtracted from the calculated D_e and negligible anomalous fading was achieved. OSL-IRSL measurements were performed at the Department of Materials Science of the University of Milano Bicocca and at the Luminescence Dating Laboratory of the University of Sassari, Italy.

3 Results and discussions

The main morphological properties of the investigated soil profiles are shown in table 1. The LS soil profile was very thick (more than 4 m), and it included at least 4 main pedogenetic discontinuities separating different stratigraphic units, in which different soil forming processes created different types of horizons (Bw, Bt and Btx horizons). The limit between the different stratigraphic units was usually clear and linear, it was abrupt only between the surface Bw horizon and the underlying Bt one. The deep 4Bt horizon had a small quantity of stones (chert fragments), evidencing a partial mixing with slope materials. Nearby, close to rock outcrops, Terra Rossa horizons (strongly rubified horizons with Munsell colour of 2.5YR 3/6 or 4/6, particularly rich in clays) and weakly developed plinthites were observed as well. The abrupt lateral limit between thick loess covers and shallow Terra Rossa soils on rock outcrops was likely associated with tectonic activities, even if no data nor precise map is available at the moment. According to the WRB taxonomic system (IUSS Working Group WRB 2014), the LS profile can be classified as Rhodic Alisol (Siltic) over Rhodic Fragic Luvisol (Siltic, Profondic) over Rhodic Luvisol (Loamic).

The TR profile was shallower, limited by hard rock at ca. 170 cm. Two discontinuities were immediately visible, between the light-coloured, silt and sand-rich EB horizon and the underlying red, silt- and clay-rich 2Bt1 horizon, and between this latter and the redder, clayey and stone-rich 3Bt below. The limit between the two upper stratigraphic units was irregular, with glossae, possibly derived by root channels. The stone fragments, observed mainly in the EB and 3Bt horizons, are composed of chert, which is resistant to weathering. According to the WRB taxonomic system (IUSS Working Group WRB 2014), the TR profile can be classified as Chromic Cambisol (Siltic) over Rhodic Luvisol (Clayic).

The granulometric analysis in the LS profile showed that all soil horizons down to 410 cm of depth were dominated by silt, but with an increasing clay fraction (fig. 3, table 3) with depth. All samples in the LS profile are also slightly richer in clays, thus their curve falls into the range of weathered loess, in which pedogenesis (clay lessivage and illuviation) and mineral weathering caused an important increase in the clay fraction. The curve is also typical for reworked loess deposits in Northern Italy (Cremaschi et al. 1987), slightly enriched in sand. The EB horizon in the TR profile has a curve compatible with a colluvial loess mixed with slope materials (particularly rich in sand), in agreement with its stone content (Costantini et al. 2018); the granulometric composition could resemble the upper layer of the Central European cover beds (Semmel and Terhorst 2010). Below, the 2Bt1 horizon was mainly silty and its curve clearly resembles the one characterizing most LS soil horizons, while the 3Bt3 one was mainly clayey (clay 59.9%, table 3), evidencing a mainly non-aeolian origin. Some Terra Rossa soils in Italy have higher clay contents (e.g. Priori et al. 2008; D'Amico et al. 2015), as it often happens when soils are mainly derived from the residuals of dissolution of limestones. However, the MOC is rich in non-calcareous materials, such as chert and sandstone fragments, which are likely related with the not-so-high clay content in the 3Bt3 horizon.

Luminescence dating results showed that surface soil horizons are recent (table 2). In particular, TR-EB horizon has an age of ca. 2.7±0.8 ka; LS-Bw horizon is a bit older (7.7 ± 1.6 ka). IRSL shows that this horizon has been isolated from sunlight since the Early-Middle Holocene. Both horizons are, however, derived from reworked materials, and they likely include Late Glacial loess mixed by slope processes and tree uprooting. In both horizons, the presence of loess is verified by texture and granulometric curves; however, TR EB has a quite large stone content. The red, clay-rich 3Bt3 horizon in the TR profile was much older. In fact, both quartz and K-feldspar are saturated or close to saturation. The minimum age is 453 ka, thus this profile started its formation at least in Marine Isotopic Stage MIS 12 (Middle Pleistocene), or even in older periods.

In LS soil, the 2Bt3 horizon, at ca. 120 cm depth, has an age of ca. 40 ka (39 ± 4 ka). This loess layer was thus deposited during MIS 3, corresponding to a glacial period preceding the Last Glacial Maximum. The lower part of the same horizon (2Bt3), at a depth of ca. 170 cm below the surface, with age of ca. 48 ± 3 ka, is formed in a loess layer still apparently deposited during MIS 3. The 3Btx3 horizon at 270 cm depth, particularly enriched in Fe-Mn coatings and with a different glossae orientation compared to the 2Bt3 horizon above, had an older deposition age, dating back to ca. 83 ± 6 ka (MIS 5a or early MIS 4). The same horizon, but at 350 cm depth, had a slightly older age, dating back to ca. 105 ± 8 ka (MIS 5c or MIS 5d). The underlying

4Bt horizon, which did not have fragic properties, was deposited 122 ± 10 ka BP (MIS 5e or MIS 6), perhaps reaching back to a previous glacial period.

As it frequently happens in Italian loess covers, no loess-paleosol sequence is recognizable (with the notable exception of Monte Netto, Zerboni et al. 2015). Loess covers deposited in different periods are all pedogenized and are part of complex polygenetic soils (Costantini et al. 2018), and only differences in pedogenic features are recognizable. This could be explained by a possible truncation of profiles during erosive periods, or because each loess deposition was not thick enough to allow isolation of deeper soils from the surface pedogenesis during following biostasy periods.

The at least Middle Pleistocene age of the 3Bt horizons in the TR profile is in agreement with the age of red soils in Central European loess areas; for example, Buggle et al. (2014) found that Early and Middle Pleistocene interglacials had climatic conditions favouring the formation of hematite, and red paleosols in loess-paleosols sequences were formed in MIS 11 and older. The red colour of the more recent 2Bt2, 3Btx and 4Bt horizons in LS soil (with IRSL ages younger than ca. 125 ka), however, are not explainable in the same way. This is in contrast with Busacca and Cremaschi (1998), who found 2.5YR colours only in the deep alluvial substrate, deposited between 400 and 780 ka (MIS 11-17). MIS 3 paleosols in the southern Po Plain Apennine margin, formed during temperate interstadial conditions, did not become redder than 7.5YR (Zuffetti et al. 2018).

Quite a large number of samples appear as deposited during temperate interstadial periods (i.e. LS 270 and LS 350 deposited during MIS 5a and 5c respectively) or even during the warm Eemian interglacial (LS 440, dated from MIS 5e). In particular, it is well known that the climatic conditions during the Eemian were warm and humid in the Po Plain, normally leading to strongly weathered and rubified soils (e.g. Ferraro 2009; Zerboni et al. 2015). The plant cover was presumably thick forest (Klotz et al. 2003), and the small glaciers in the Alps associated with the slightly higher temperatures compared to the Holocene (Pons et al. 1992) were likely producing little amounts of sediments, in a similar way to what is happening during the Holocene. Thick loess deposits were thus unlikely forming during that period. Strong erosive processes, able to deeply rejuvenate the soil layer were unlikely as well under the thick forest cover. Loess deposition needs colder and drier climates with lower vegetation cover, which permit the existence of large deflation surfaces. Thus, an underestimation of the oldest loess deposition periods cannot be excluded due to mixing caused by tree uprooting or other slope morphodynamic processes. Likewise, loess deposition of the deep LS 4Bt horizon during full glacial conditions in MIS 6 or older is thus much more likely than during the warm interglacial MIS 5e. In the same way, LS 3Btx3 (350 cm in depth) could be better attributed to MIS 5d, characterized by slightly colder and drier conditions than MIS 5c (Wohlfahrt 2013). Considering a hypothetical age underestimation in our deep samples would make our results comparable to other dated loess-paleosols sequences in Europe (e.g. Novotny et al. 2011).

4 Conclusions

Our results show that the deepest loess layer in LS profile (122 ± 10 ka) on Monte Orfano appears to be the oldest loess deposit among those quantitatively dated in Northern Italy (Cremaschi et al. 2011, 2015; Livio et al. 2014; Peresani et al. 2008; Zerboni et al. 2015; Frigerio et al. 2017; Costantini et al. 2018), between the Alpine margins and the Apennine fringe. In fact, the oldest published numerical ages until now are those of Ghiardo terrace (Reggio Emilia, Italy), which is 81.6 ± 10.9 ka BP (Cremaschi et al. 2015), and the San Colombano one, which is 89 ± 8.8 ka (Panzeri et al. 2011). Northern Italian loess cover seems to have been deposited between MIS 4 and MIS 2 (Costantini et al. 2018), even if most European loess-paleosols sequences started their formation in Early or Middle Pleistocene. Thus, based on our results, we can assume that loess deposition was actually active in the Po Plain also before MIS4, as assumed only by soil properties by many older studies (e.g. Coudé-Gaussen 1990; Billard and Orombelli 1986) and by more recent ones (Negri et al. 2020), but never verified by numerical dates. The pedogenic and paleoclimatic implications of our results will be analysed in a following paper.

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Figure captions

Fig. 1 location of the Monte Orfano, and the location of the other OSL-dated loess layers, available in the literature, in the Po Plain (Northern Italy). 1: Frigerio et al. (2017); 2: Cremaschi et al. (2011; 2015); 3: D'Amico et al. (present paper); 4: Zerboni et al. (2015); 5: Ferraro et al. (2009); 6: Peresani et al. (2008); 7: Accorsi et al. (1990); 8: Cremaschi et al. (2015); 9: Panzeri et al. (2011)

Fig. 2 the LS (left) and TR (right) profiles

Fig. 3 granulometric curves for the analysed soil horizons. Typical curves for loess (reworked and weathered) are observed for LS samples, and TR60, while mixing is visible in TR30 from the high sand content; the curve of TR160 has a different shape, evidencing a non-aeolian origin