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Soil evolution over the Quaternary period in a Mediterranean climate (SE Spain)

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Abstract

Palaeosols in the Granada Basin (SE Spain) have been studied in two different situations: surface soils on geomorphically stable surfaces since the Early Pleistocene with younger pedogenic overprinting and buried soils on unstable surfaces from the Middle–Late Pleistocene on which successive erosional–depositional episodes have alternated with pedogenic episodes. For each soil clay and iron accumulation indices, the $Fe_t + Al_t/Si_t$ ratio, clay mineralogy and micromorphological features were used to estimate the degree of soil development. From the Early to the early Late Pleistocene, the main pedogenic processes were the leaching of carbonates, weathering, illuviation and rubification, which resulted in Bt horizons with red colours, clay texture, clay coatings and kaolinite neoformation. The degree of weathering and the development of these Bt horizons varied over time, and the soils that formed on the surfaces from the Early Pleistocene show strongest weathering and development. However, after their formation, there were periods in which they were partially truncated and recalcified, resulting in polygenetic soils. The different degrees of development of the buried soils during the last 474,000 years indicate that the wettest warm period was stage 7 and the driest, stage 5. Stages 9 and 11 must have had climates with intermediate wetness. Since the clay accumulation and iron oxide accumulation indices, the differences in $Fe_t + Al_t/Si_t$ ratio between Bt and C horizons, the extent of kaolinite neoformation and the micromorphological features of the soils formed during stage 7 are all similar to the surface soils that formed on Early Pleistocene deposits, these features cannot be used to date surfaces older than 242,000 BP. By contrast, the soils that formed during stage 7 and later periods show different extents of development and thus can be used for the approximate dating of landforms.

Keywords: Pleistocene; Soil development; Relict soils; Buried soils; Climatic changes

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1. Introduction

According to Jenny (1941), soils and their properties are the product of the different soil-forming factors (climate, organisms, relief, parent material and time) that control the degree of soil development, as indicated by comparisons with the parent material (Harden, 1990). Because the soil-forming factors also govern geomorphic processes, landscape evolution is intimately related to soil development (McFadden and Kneupfer, 1990).

Over time, soil-forming factors, especially climate and vegetation, may change in such a way that many old soils, palaeosols, are not related to the present climate and vegetation. Palaeosols, defined as soils formed in a landscape of the past (Ruhe, 1956; Yaalon, 1971), include both relict and buried soils (Bronger and Catt, 1989). Relict soils are surface soils which show inactive characteristics inherited from past periods when soil-forming conditions were sufficiently different from those of the present to produce features unlike any of those developing currently in the same area. They are likely to have properties similar to those of buried soils that formed during the same past periods (Catt, 1989). It is possible to reconstruct relief and/or palaeoclimate as soil-forming factors on the basis of the processes inferred from palaeosol properties (Bronger and Catt, 1989).

Properties of the different soil horizons have also been used to determine the age of soils (Harden, 1982; Levine and Ciolkosz, 1983; Harrison et al., 1990) and thus the approximate age of the landforms (Semmel, 1989). For this reasoning to be valid, climatic conditions must remain relatively stable over the entire soil-forming period, for only then do the soil properties increase constantly with time (Bockheim, 1980; Birkeland, 1990). However, dating becomes complex on surfaces subject to long-term climatic fluctuations. This can be solved if relationships between specific soil properties and climatic fluctuations are known.

Among soils up to 85,000 years old (estimated from the clay accumulation index of Levine and Ciolkosz, 1983) in Sierra Nevada (SE Spain), Simón et al. (2000) distinguished two well-differentiated groups: (a) soils approximately 85,000 years old (early Late Pleistocene), with strongly developed Bt horizons, which are red in colour, clayey in texture and contain abundant clay coatings and much kaolinite; and (b) soils younger than 15,000 years (Late Pleistocene–Holocene), with less developed Bw horizons, which are brown in colour, without evidence of clay illuviation and with small kaolinite contents. The degree of development of the latter group was less in the younger surface. No soils of intermediate ages (between 85,000 and 15,000 years old) were found, apparently because of unstable surfaces and a cold climate during this time interval, which would have discouraged chemical weathering and soil development (Catt, 1989). In Sierra Nevada, SE Spain, some surfaces older than 85,000 years are preserved, probably formed during the Riss glacial period (Hempel, 1960; Messerli, 1965; Lhenaff, 1977), but their soils are strongly eroded and cannot be used for determining soil–time relationships. However, lower elevation alluvial fan deposits around Sierra Nevada that resulted from the tectonic activity in the Late Pliocene exhibit stable surfaces with soils of Early Pleistocene age (Estévez and Sanz de Galdeano, 1983). Also, the reorganization of the relief during the Middle–Late Pleistocene formed unstable surfaces on which successive depositional episodes alternated with pedogenic episodes (Sanz de Galdeano and López-Garrido, 1999).

In this paper, we compare soil development on two different types of surfaces in SE Spain: (1) geomorphically stable surfaces with old soils also showing younger pedogenic overprinting (surface soils) and (2) unstable surfaces with successive erosion–deposition episodes, forming sequences of buried soils in which the successive pedogenic stages are spatially distinct. The aim is to reconstruct soil development over the Quaternary period in a Mediterranean climate.

2. Site details

The Granada Basin is located in the central sector of the Betic Cordillera (SE Spain) in the contact area between the External and the Internal Zones (Fig. 1). The External Zones, located to the north of the Granada Basin, are made up of Mesozoic and Tertiary carbonate

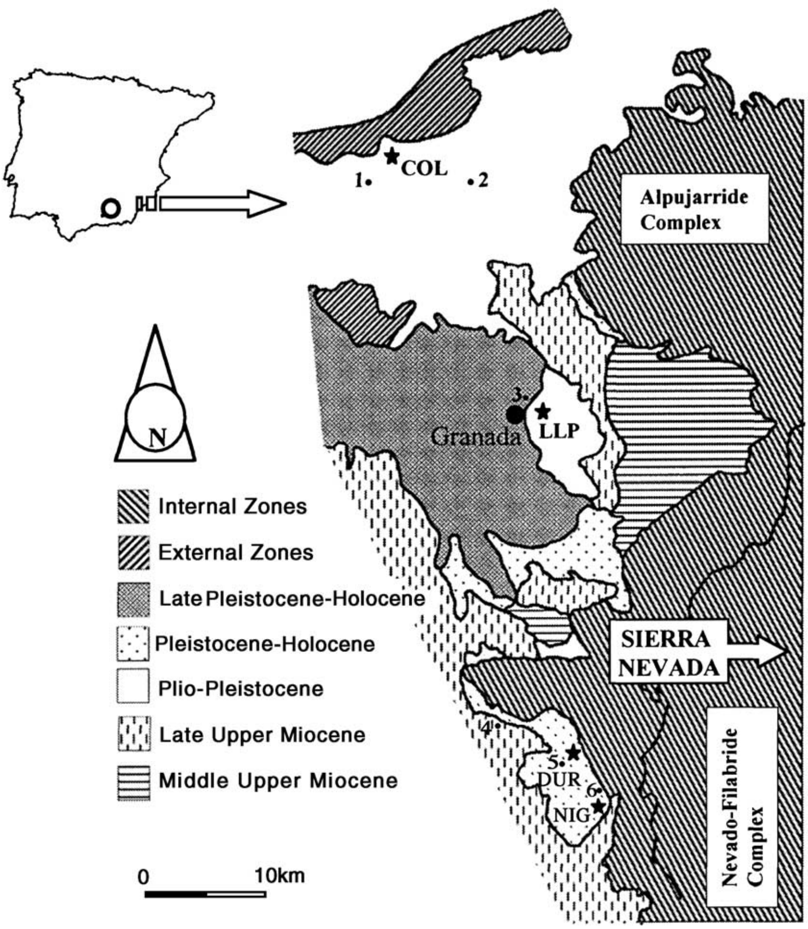

Fig. 1. Simplified geological map of the study area showing the location of the soils studied (COL = Colomera, DUR = Dúrcal, LLP = Llano de la Perdiz and NIG = Nigüelas) and weather stations (1 = Colomera El León, 2 = Pantano Cubillas, 3 = Granada Cartuja, 4 = Padul, 5 = Dúrcal Central and 6 = Nigüelas).

rocks (limestones and dolomites). The Internal Zones, located to the east of the Granada Basin, are made up of two complexes: the Nevado–Filabride Complex, occupying the central sector of Sierra Nevada and composed mainly of mica schists and quartzites; and the Alpujarride Complex, forming a ring around Sierra Nevada and composed of phyllites, quartzites, limestones and dolomites. During the Late Miocene, Sierra Nevada was strongly uplifted whereupon massive erosion gave rise to major alluvial fans containing large blocks reworked from the Nevado–Filabride Complex on the borders of the basin, and lacustrine formations were deposited in subsiding areas within the basin (Fernández and Soria, 1986–1987). In the Late Pliocene, there was renewed uplift (Estévez and Sanz de Galdeano, 1983), and significant new coarse detrital inputs from the External and the Internal Zones were deposited in the basin during the Early Pleistocene. Consequently, the northern half of the basement of the Granada Basin is made up of Mesozoic and Tertiary carbonate materials from the External Zones, and the eastern half consists of Paleozoic and Triassic materials from the metamorphic complexes of the Internal Zones (Fernández et al., 1996). Further reorganization of the relief occurred during the Middle?–Late Pleistocene. Low areas on the borders of Sierra Nevada rose and subsequent erosion episodes, probably activated by cold episodes, have left behind abundant coarse-grained deposits. In summary, the uplift of Sierra Nevada and the different sedimentation episodes in the Granada Basin were not continuous processes but resulted from pulses of tectonic activity separated by periods of relative quiescence (Sanz de Galdeano and López-Garrido, 1999).

The present climate of the area (Table 1) is typically Mediterranean (hot, dry summers; cold, wet winters; temperate autumns and springs with variable rainfall). The natural vegetation (Valle, 1985; Ruiz de la Torre, 1990) is oak forest (*Quercus rotundifolia*) with shrubs (*Juniperus oxycedrus, Ruscus aculeatus, Daphne gnidium, Clematis flammula, Lonicera etrusca* and *Hedera helix*) and herbaceous plants (*Paeonia coriacea, P. broteroi, Primula vulgaris* and *Viola* sp.). However, in many sectors, this has been replaced by crops such as olive and almond trees.

Table 1
Climatic data of the weather stations closest to the different sectors studied

Weather station	Altitude (m)	MAP (mm)	MAT (°C)	Mean seasonal precipitation (mm)				Mean seasonal temperature (°C)			
				Summer	Autumn	Winter	Spring	Summer	Autumn	Winter	Spring
Granada (Cartuja)	720	477.6	15.3	7.5	43.5	56.7	51.5	23.9	16.0	7.7	13.3
Pantano Cubillas	630	525.5	15.3	11.5	44.9	67.5	51.2	23.9	16.5	7.6	13.0
Colomera (El León)	860	645.6	14.4	13.2	56.0	89.3	56.7	23.7	16.1	6.6	11.2
Padul	753	423.2	15.8	7.9	42.4	53.0	23.5	23.9	16.7	8.9	13.9
Dúrcal Central	890	509.1	15.3	8.5	49.2	62.3	46.8	23.9	16.1	8.0	13.0
Nigüelas	931	489.6	14.0	9.6	48.4	62.0	43.0	21.7	14.9	7.6	12.0

MAP = mean annual precipitation; MAT = mean annual temperature.

3. Materials and methods

We have studied soils that developed on three alluvial fans in the Granada Basin that have remained relatively stable over time (Fig. 1). Two of these, Dúrcal (DUR) and Llano de la Perdiz (LLP), date from the Early Pleistocene (Aguirre, 1957; Ruiz Bustos et al., 1992; Sanz de Galdeano and López-Garrido, 1999), and consist of gravels with mica schists and quartzites from the Nevado–Filabride Complex and a small proportion of limestones and dolomites from the Alpujarride Complex. The third alluvial fan, Colomera (COL), also dates from the Early Pleistocene (Fernández and Soria, 1986–1987) but has gravel clasts of limestones and dolomites from the External Zones.

Similarly, in a sector adjacent to the area that rose during the Middle?–Late Pleistocene, Nigüelas (NIG), we studied a vertical section approximately 11.5 m high where four depositional episodes of gravelly materials, equivalent to those from DUR and LLP (mica schists and quartzites from the Nevado–Filabride Complex with a small proportion of limestones and dolomites from the Alpujarride Complex), were distinguished. These depositional episodes alternated with pedogenic episodes. We identified and studied four buried soils and designated them (from bottom to top) as NIG-1, NIG-2, NIG-3 and NIG-4 (Fig. 2). The original surfaces of soils NIG-1 and NIG-2 were tilted before the deposition of the parent material of soil NIG-3.

Field descriptions of the soils were based on procedures of the Soil Survey Staff (1990). The micromorphological study was based on thin sections (Bullock et al., 1985). The Munsell soil colour chart was used to describe the soil colours. Particle size distribution was determined by the pipette method after the removal of organic matter with H_2O_2 and dispersion by shaking with sodium hexametaphosphate (Loveland and Whalley, 1991). The organic carbon content was determined using the method of Tyurin (1951). The pH was measured potentiometrically in a 1:2.5 soil/water suspension. The $CaCO_3$ equivalent was determined according to Williams (1948). For the determination of the cation exchange capacity (CEC), 1 N Na–acetate was used at pH 8.2. Exchangeable bases were extracted

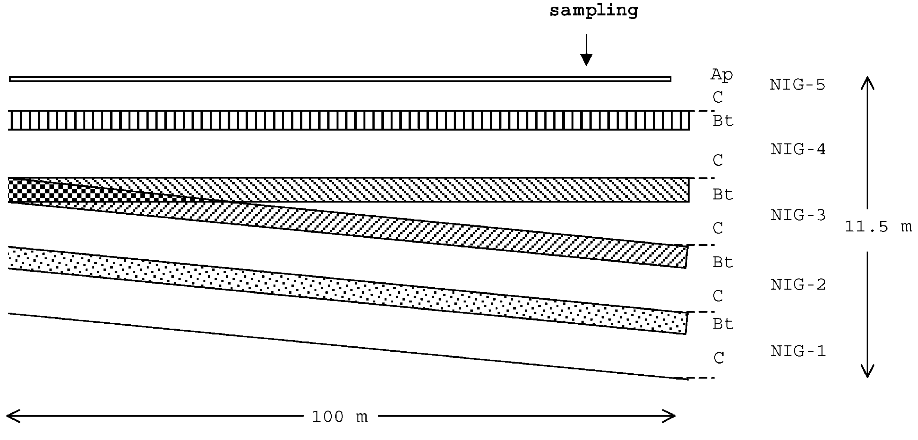

Fig. 2. Vertical section of gravel units and soils at NIG.

with 1 N NH_4–acetate at pH 7.0 and measured by atomic absorption spectroscopy (Ca and Mg) and flame photometry (Na and K). Discs of soil and lithium tetraborate (0.6:5.5) were prepared and the total contents of Si, Fe and Al were measured by X-ray fluorescence using a Philips PW-1404 instrument. X-ray diffraction patterns for the clay fraction were obtained with a Philips PW-1700 instrument using CuKα radiation, and the diffraction intensities used in the quantitative analysis were taken from Schultz (1964) and Barahona (1974). Total iron oxides (Fe_d) were extracted with citrate–dithionite (Holmgren, 1967), and the amorphous forms (Fe_o) with ammonium oxalate (Schwertmann and Taylor, 1977). Iron in the extracts was measured by atomic absorption spectroscopy. A redness index (R_r) was calculated as (hue $\times$ chroma)/value (Hurst, 1977). For this index, hue is converted to the following values: 10YR = 0.0, 7.5YR = 2.5, 5YR = 5.0, 2.5YR = 7.5 and 10R = 10.0.

To estimate the degree of development of each profile, a clay accumulation index (CI) was calculated as $\sum(B-C)T$, where B = B horizon clay content (%), C = C horizon clay content (%) and T = thickness (cm) of the B horizon (Levine and Ciolkosz, 1983). Similarly, an iron oxide accumulation index (Fe_dI) was calculated using the same equation as for the clay, where B = B horizon Fe_d content (%) and C = C horizon Fe_d content (%). To estimate the age of the soils, the equation of Levine and Ciolkosz (1983) was used:

$$\log(\text{year}) = 1.81 + 0.998 \times \log(\text{CI}).$$

4. Results

4.1. Macromorphology

The C horizons of all the soils ranged in colour from pink to light brownish grey and retained the original deposit structure although in the surface soils, the mineral particles are cemented by calcium carbonate. None of the buried soils contains a clear A horizon, suggesting that they were disturbed or truncated at the time of burial. Surface soils also seemed to be truncated, especially soils COL (under olive cultivation) and DUR (under almond cultivation), where the Ap horizons are part of the previous Bt horizons that are disturbed by ploughing. Soil LLP was the least disturbed. All soils have a well-developed Bt horizon (Table 2) characterized by a red colour and a moderate to strong angular–subangular blocky or prismatic structure. The redness indices of the most strongly developed Bt horizons in each soil ranged from 7.5 to 15, being greatest in the surface soils COL, LLP and DUR and the buried soil NIG-3, intermediate in soils NIG-1 and NIG-2 and least in NIG-4.

4.2. Analytical features

All the Bt horizons contained more clay than the C horizons (Table 3). The clay accumulation index (CI) was greatest in soils COL, DUR, LLP and NIG-3, intermediate in soils NIG-1 and NIG-2, and least in soil NIG-4 (Fig. 3). The pH in the Bt horizons was mostly alkaline (Table 3) except for soil LLP, where slightly acid pH values appeared in the

Table 2
Field soil data

Profile	Horizon	Surface age	Altitude (m)	Depth (cm)	Colour dry	R_r	Structure (type, size, grade)
NIG-5	Ap	Pleistocene?	842	0-15	10YR 6/3		sbk, f, 1
	C			15-150	10YR 6/4		sg, 0
NIG-4	Bt1	Pleistocene?		150-167	5YR 4/6	7.5	abk, m, 2
	Bt2			167-184	2.5YR 4/4	7.5	abk, m, 2
	Bt3			184-198	7.5YR 5/6	3.0	sbk, m, 1
	CB			198-255	10YR 6/4		sbk, f, 1
	Ck			255-442	2.5Y 6/2		sg, 0
NIG-3	Bt1	Pleistocene?		442-492	10R 4/6	15.0	abk, m, 3
	Bt2			492-532	10R 4/6	15.0	abk, m, 2
	Ck			532-560	10YR 6/3		sg, 0
NIG-2	Bt1	Pleistocene?		560-595	2.5YR 4/6	11.2	abk, m, 2
	Bt2			595-638	2.5YR 4/4	7.5	sbk, m, 2
	Ck			638-788	10YR 6/3		sg, 0
NIG-1	Bt1	Pleistocene?		788-868	2.5YR 4/6	11.2	abk, m, 2
	Bt2			868-898	7.5YR 6/6	2.5	abk, f, 1
	Ck			898-1150	10YR 6/3		sg, 0
DUR	Ap	Early Pleistocene	796	0-10	7.5YR 4/6	3.7	cr, f, 1
	Bt1			10-55	10R 4/6	15.0	pr, c, 3
	Bt2			55-84	2.5YR 4/6	11.2	abk, m, 2
	Ckm			> 84	10YR 6/3		m, 0
LLP	A	Early Pleistocene	985	0-2	7.5YR 4/6	3.7	cr, f, 1
	E			2-18	7.5YR 6/6	2.5	gr/sbk, m, 2
	Bt1g			18-50	10R 4/6	15.0	abk, m, 2
	Bt2g			50-85	2.5YR 4/6	11.2	abk, c, 2
	Bt3g			85-121	2.5YR 4/6	11.2	abk, c, 2
	Bt4g			121-176	2.5YR 4/6	11.2	abk, m, 2
	Ckm			>176	7.5YR 8/4		m, 0
COL	Ap1	Early Pleistocene	710	0-6	2.5YR 3/6	15.0	abk, m/gr, f, 2
	Ap2			6-20	2.5YR 3/6	15.0	pr, c, 3
	Bt			20-50	10R 4/6	15.0	pr, m, 3
	Btk1			50-90	2.5YR 4/6	11.2	abk, f, 2
	Btk2			90-120	2.5YR 7/4	4.3	abk, f, 1
	Ckm			>120	10YR 8/2		m, 0

Abbreviations for structure and consistence are from Soil Survey Staff (1990).

upper weakly calcareous horizons. In soil NIG-2, the Bt horizons seem to have been recalcified by the leaching of calcium carbonate from soil NIG-3. $CaCO_3$ has accumulated in the Ck horizons, particularly in the surface soils where the mineral particles are cemented to form a Ckm horizon. For this reason, they could be designated as Bk or Bkm, but because most or all of the original parent material structure has not been obliterated (Soil Survey

Table 3
Analytical soil data for the fine-earth (< 2 mm) fraction

Profile	Horizon	Gravel (%)	Sand (%)	Silt (%)	Clay (%)	pH	$CaCO_3$ (%)	OC (%)	Exchangeable bases ($cmol_c\ kg^{-1}$)					Base saturation (%)
									Na^+	K^+	Ca^{2+}	Mg^{2+}	CEC	
NIG-5	Ap	34.8	62.5	28.7	8.8	8.5	3.7	0.62	0.11	0.69	1.97	3.02	5.59	100
	C	35.6	60.3	30.3	9.4	8.1	4.1	0.55	0.15	0.38	2.14	3.17	5.42	100
NIG-4	Bt1	36.5	42.1	22.2	35.7	7.3	0.5	0.32	0.10	0.16	10.00	2.70	12.46	100
	Bt2	37.1	46.3	26.7	27.0	7.4	0.6	0.25	0.14	0.13	6.51	2.72	9.02	100
	Bt3	42.6	52.2	29.8	18.0	7.1	0.6	0.16	0.14	0.07	4.28	2.76	6.44	100
	CB	43.7	55.0	31.8	13.2	7.4	0.6	0.18	0.19	0.04	3.57	2.90	6.01	100
	Ck	44.0	60.0	32.4	7.6	8.3	17.5	0.20	0.20	0.04	2.22	3.17	5.01	100
NIG-3	Bt1	37.0	30.5	14.4	55.1	8.3	0.4	0.14	0.12	0.39	10.23	5.00	14.20	100
	Bt2	43.5	32.1	12.7	55.2	8.3	0.5	0.14	0.10	0.36	9.90	4.70	14.50	100
	Ck	49.3	64.8	29.4	5.8	9.1	6.4	0.14	0.02	0.04	1.22	3.57	4.30	100
NIG-2	Bt1	39.3	37.1	17.7	45.2	8.6	2.2	0.16	0.09	0.26	7.24	9.18	14.18	100
	Bt2	39.1	41.9	29.6	28.5	8.3	2.8	0.21	0.07	0.18	4.24	7.31	10.31	100
	Ck	47.6	57.5	32.5	10.0	8.5	2.3	0.20	0.07	0.05	3.74	5.93	8.59	100
NIG-1	Bt1	38.7	41.7	22.7	35.6	8.5	0.4	0.20	0.09	0.35	2.99	7.31	9.88	100
	Bt2	43.3	65.5	22.5	12.0	8.8	0.7	0.23	0.05	0.07	4.24	3.75	7.73	100
	Ck	49.3	58.4	30.7	10.9	8.9	1.2	0.25	0.02	0.07	2.24	2.68	4.73	100
DUR	Ap	33.4	31.2	20.6	48.2	8.5	1.2	0.55	0.13	0.76	7.15	7.08	15.12	100
	Bt1	29.8	17.3	10.3	72.4	8.3	1.0	0.46	0.15	0.51	13.97	6.90	17.19	100
	Bt2	36.8	26.1	17.0	56.9	8.2	0.5	0.45	0.26	0.40	10.23	5.60	16.33	100
	Ckm	50.9	65.0	26.2	8.8	8.4	30.7	0.33	0.03	0.15	2.52	1.22	3.44	100
LLP	A	32.7	74.4	14.4	11.1	7.4	1.1	3.77	0.06	0.26	14.00	2.25	24.12	68.7
	E	38.0	66.2	21.3	12.5	7.5	0.61	1.18	0.02	0.15	8.00	1.67	12.99	75.7
	Bt1g	31.7	32.5	13.3	54.2	6.6	0.45	0.65	0.06	0.15	8.70	3.67	16.23	77.8
	Bt2g	35.0	45.3	15.6	39.1	6.5	0.56	0.35	0.12	0.11	13.00	5.25	14.38	100
	Bt3g	46.4	53.2	15.5	31.3	6.7	0.68	0.26	0.15	0.11	9.70	4.42	12.99	100
	Bt4g	48.8	59.9	18.0	22.1	6.6	0.9	0.28	0.21	0.11	10.70	4.58	12.06	100
	Ckm	53.4	74.4	18.5	7.1	8.7	36.0	0.31	0.21	0.04	2.53	4.62	6.03	100
COL	Ap1	12.9	21.3	21.0	57.6	8.0	1.9	0.10	0.11	1.85	29.00	1.25	24.33	100
	Ap2	6.3	22.4	18.9	58.7	8.2	2.0	0.10	0.09	1.66	27.70	1.00	29.34	100
	Bt	0.2	5.0	3.4	91.6	7.7	0.8	0.05	0.14	0.87	33.20	1.67	38.65	93.0
	Btk1	15.1	12.7	42.6	44.7	8.3	51.1	0.03	0.09	0.49	23.21	0.92	22.90	100
	Btk2	29.0	15.4	62.9	21.7	8.5	71.4	0.02	0.06	0.23	10.92	0.50	11.45	100
	Ckm	44.3	46.3	38.8	14.9	8.7	80.0	0.02	0.05	0.10	6.32	0.50	6.30	100

OC = organic carbon; CEC = cation exchange capacity.

Staff, 1990), they are labelled as C horizons. The gravel content was similar in all C horizons (Table 3). The Bt horizons of all the soils contained very little organic C, indicating that the mineralization of organic matter predominated during the development of these soils.

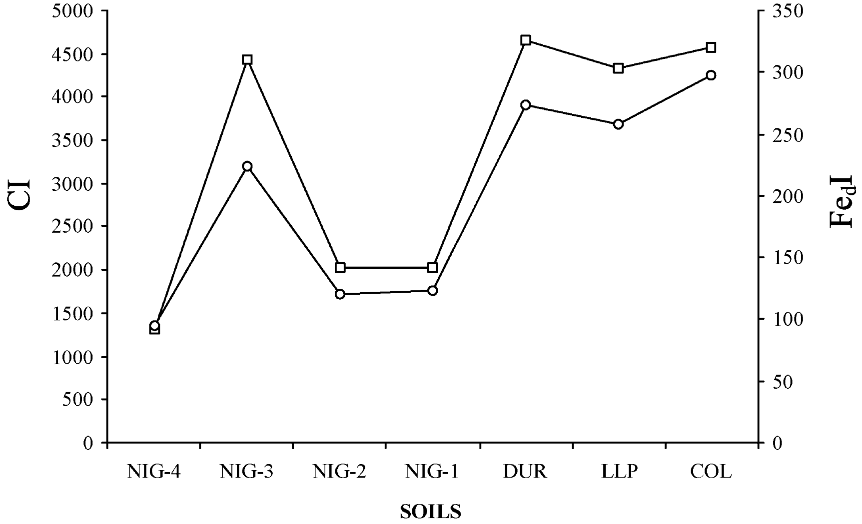

Fig. 3. Clay accumulation (CI) (squares) and iron oxide accumulation (Fe_dI) (circles) indices for all the soils.

The cation exchange capacity (CEC) was related to the clay and organic C contents by the multiple-regression equation:

$$\mathrm{CEC}\ (\mathrm{cmol_c\ kg^{-1}}) = 5.008 \times \mathrm{OC}\ (\%) + 0.345 \times\ \mathrm{Clay}\ (\%) \qquad (r = 0.965)$$

The regression coefficients show that the influence of organic C on the CEC values was roughly 15 times greater than that of the clay. Exchangeable bases are dominated mainly by Ca^{2+} and Mg^{2+}, with lesser amounts of Na^+ and K^+. Only the Ap horizons of soils DUR and COL show relatively high contents of K^+ which is attributable to fertilizing. All soils are eutric. The high base saturation (equal or close to 100%) can be attributed to basification by the runoff of waters rich in Ca^{2+} and Mg^{2+} originating from the surrounding terrain of limestone and dolomite. Only soil LLP, whose surface was isolated from the surrounding terrain by an incision of the rivers probably during the Middle?–Late Pleistocene (Sanz de Galdeano and López-Garrido, 1999), was less affected by this runoff and has a base saturation of less than 80% in its upper horizons. This basification must have occurred subsequent to soil formation and the original pH of the Bt horizons should have been more acidic than that at present. The neutral or nearly neutral pH values and high base saturation of these red soils have also been explained by the occurrence of dry periods during which there was a capillary rise of bases (Lamouroux, 1971). Whatever the mechanism causing basification, the red soils are usually less acidic than the brown soils that developed over the similar parent material (Duchaufour, 1977).

The contents of the total iron (Fe_t) that was extracted by dithionite (Fe_d) and that extracted by oxalate (Fe_o) were all greater in the Bt horizons than the C horizons (Table 4). The values of the iron oxide accumulation index (Fe_dI) showed a similar pattern to those of the clay accumulation (Fig. 3) and redness indices (Fig. 4). It was greatest in the Bt horizons of soils COL, LLP, DUR and NIG-3, least in soil NIG-4 and intermediate in soils NIG-1 and NIG-2. The values of the Fe_o/Fe_d ratio in the Bt horizons were very small

Table 4
Total silica (Si_t), aluminium (Al_t) and iron (Fe_t), forms of iron and clay minerals of the soils

Profile	Horizon	Al_t (%)	Si_t (%)	Fe_t (%)	Fe_d (%)	Fe_o (%)	Clay minerals (%)			
							K	I	S	P
NIG-5	Ap	15.17	58.43	5.49	1.37	0.06	+	+++	++	+
	C	15.32	58.92	5.42	1.52	0.04	+	+++	++	+
NIG-4	Bt1	18.60	62.54	7.58	2.77	0.09	++	+++	(+)	+
	Bt2	17.51	61.94	6.83	2.82	0.06	++	+++	+	+
	Bt3	16.91	65.65	6.78	2.50	0.06	+	+++	+	+
	CB	16.34	65.89	6.27	2.27	0.04	+	+++	++	+
	Ck	17.15	67.04	6.24	1.57	0.03	+	+++	++	+
NIG-3	Bt1	22.35	53.12	9.59	3.70	0.06	+++	+++	nd	+
	Bt2	22.00	53.41	9.45	3.58	0.06	+++	+++	nd	=
	Ck	14.25	64.81	5.46	1.16	0.03	+	+++	++	+
NIG-2	Bt1	20.39	58.03	8.60	3.65	0.06	++	+++	nd	+
	Bt2	18.79	58.71	8.21	3.23	0.07	++	+++	nd	+
	Ck	16.31	62.79	6.89	1.87	0.03	+	+++	++	+
NIG-1	Bt1	18.34	61.59	8.03	1.99	0.10	++	+++	nd	+
	Bt2	14.50	69.09	6.06	0.74	0.04	++	+++	nd	+
	Ck	15.90	66.36	6.73	0.53	0.04	+	+++	++	+
DUR	Ap	19.06	58.76	7.32	3.22	0.07	++	+++	nd	++
	Bt1	22.90	53.03	8.93	4.02	0.07	+++	+++	nd	++
	Bt2	22.12	56.30	8.52	3.16	0.04	+++	+++	nd	++
	Ckm	10.91	48.08	4.18	0.37	0.03	+	+++	+	+
LLP	A	7.41	78.12	4.10	1.02	0.06	+++	+++	nd	nd
	E	9.62	76.79	4.86	0.97	0.06	+++	+++	nd	nd
	Bt1g	20.37	57.30	9.65	3.00	0.11	+++	+++	nd	nd
	Bt2g	17.33	62.70	8.61	2.82	0.10	+++	+++	(+)	nd
	Bt3g	17.16	62.96	8.49	2.59	0.11	+++	+++	+	nd
	Bt4g	16.02	65.46	8.11	1.57	0.07	++	+++	+	nd
	Ckm	7.09	40.52	4.65	0.74	0.01	+	+++	++	nd
COL	Ap1	17.81	60.69	7.75	3.42	0.14	+++	+++	nd	nd
	Ap2	17.95	60.41	7.63	3.56	0.14	+++	+++	(+)	nd
	Bt	23.92	51.52	9.65	4.56	0.14	+++	+++	(+)	nd
	Btk1	12.08	25.46	5.20	2.06	0.03	+	+++	++	nd
	Btk2	4.95	13.85	2.40	0.97	0.01	+	+++	+++	nd
	Ckm	2.14	12.78	1.53	0.17	0.09	+	+	+++	nd

Clay minerals: K = kaolinite, I = illite, S = smectite, P = paragonite, tr = traces.
nd = not detected; (+) = traces; + = 5-15%; ++ = 15-40%; +++ = >40%.

(< 0.05), indicating an almost total crystallization of the hydrous Fe oxides that were formed by the weathering of silicates (Arduino et al., 1986).

In all the soils, the Fe_t/Si_t and Al_t/Si_t ratios were greater in the Bt horizons than the C horizons (Table 4), indicating that Si was more mobile than Fe and Al. As the original pH

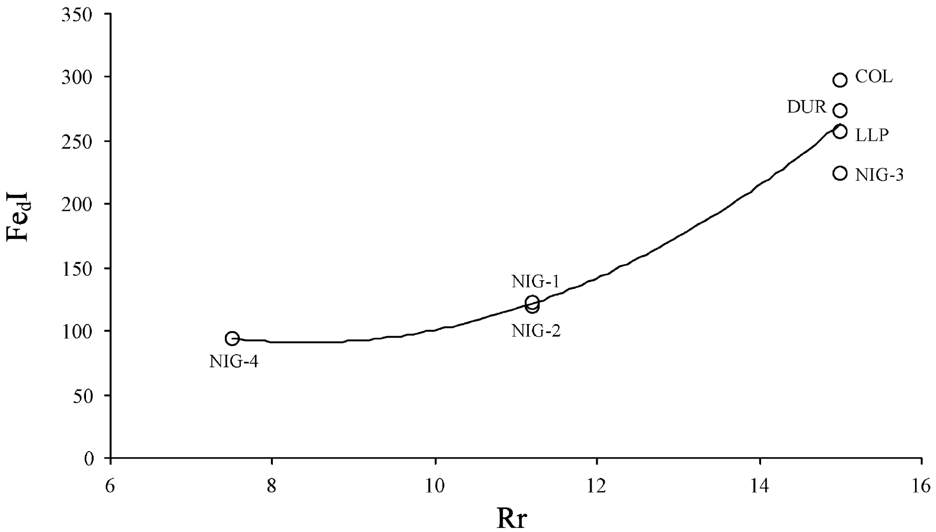

Fig. 4. Relationship between the redness index (R_r) and the iron oxide accumulation index (Fe_dI) of the Bt horizons.

of these soils should not have been < 5.0 (Loughnan, 1969), the difference in the $Fe_t + Al_t/Si_t$ ratio between the Bt horizons and C horizons should increase with greater weathering and leaching. These differences show approximately the same patterns as the CI, Fe_dI and R_r indices (Fig. 5).

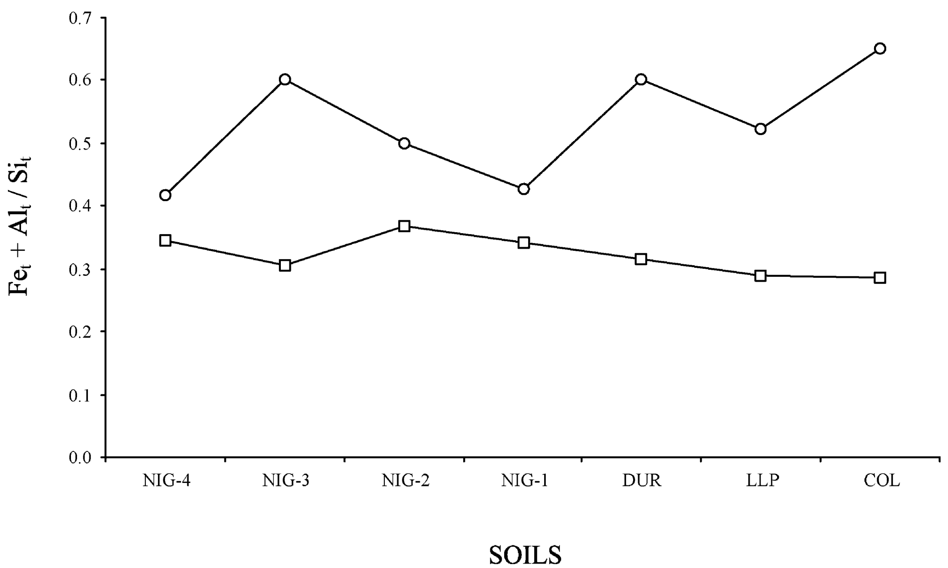

Fig. 5. The $Fe_t + Al_t/Si_t$ ratio of the Bt (circles) and C (squares) horizons.

The semi-quantitative analysis of the clay minerals (Table 4) also revealed differences between the soils. In soils COL, LLP, DUR and NIG-3, smectite is less abundant and kaolinite was more abundant in the Bt horizons than in the C horizons. However, in soils NIG-1, NIG-2 and NIG-4, the upward increase in kaolinite was weaker than in other soils. The kaolinite neoformation in the Bt horizons must have occurred before basification when the soils were more acidic.

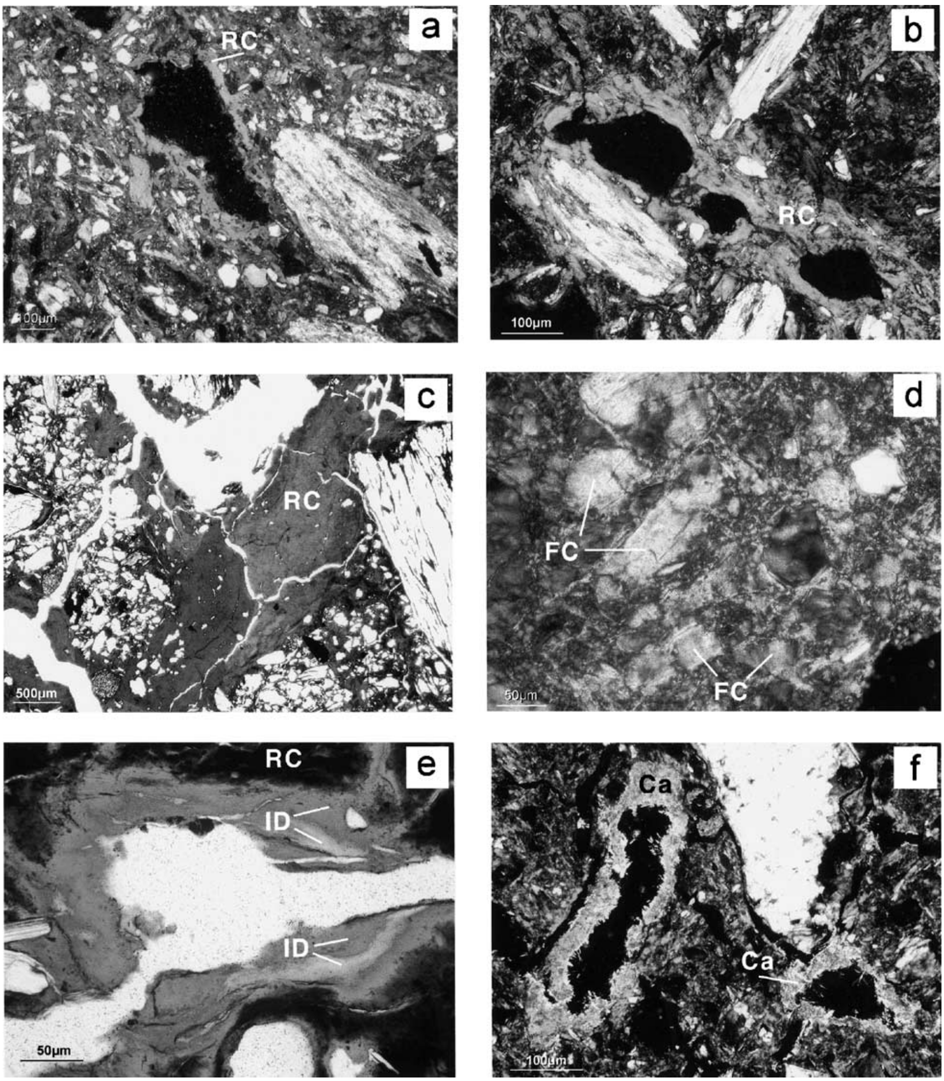

Fig. 6. Micromorphological features in the Bt horizons of the soils: (a) thin red clay coatings (RC) in NIG-4, (b) thicker red clay coatings in NIG-2, (c) the thickest red clay coatings in NIG-3, (d) fragments of red clay coatings (FC) embedded in the soil matrix in DUR, (e) red clay coatings with iron depletion zones (ID) in LLP and (f) calcitic coatings (Ca) in NIG-2.

4.3. Micromorphological features

The Bt horizons show a porphyric-related distribution, with stipple-speckled b-fabric in soils NIG-1, NIG-2 and NIG-4 and mono-granostriated b-fabric in NIG-3 and the surface soils. In NIG-4, the Bt horizons show many thin red clay coatings in the channels and other voids (Fig. 6a). They are somewhat thicker and more common in NIG-1 and NIG-2 (Fig. 6b), and even thicker and more abundant in NIG-3 and in the surface soils (Fig. 6c). The surface soils contain many fragments of these red clay pedofeatures embedded in the matrix (Fig. 6d). The red clay coatings in soil LLP have scattered yellowish zones (Fig. 6e), indicating weak hydromorphic iron depletion. Distinct and rather frequent calcitic coatings appear in the Bt horizon of NIG-2 (Fig. 6f). Similar features also appear, though with less clarity and less frequency, in the surface soils.

5. Discussion

The main pedogenic processes that affected these soils were the mineralization of organic matter, leaching of carbonates, strong weathering of smectite to kaolinite, clay illuviation and rubification, which formed strongly developed red Bt horizons of clay texture with abundant clay coatings. These soil properties must have developed under a wetter climate than that at present. In addition, the $pH > 7.0$, the high contents of exchangeable bases and the presence of $CaCO_3$ in the Bt horizons suggest subsequent calcification, the latter also being evident in the micromorphological study of soils NIG-2, DUR and COL.

5.1. Buried soils

Continuous deep oceanic sedimentary records can be used as a chronological and paleoclimatic reference for long-term climatic fluctuations (Kukla, 1977; Bradley, 1985). Because the oxygen isotopic record of the oceanic sequences provides an integrated summary of global ice-volume changes, it has been argued that the isotopic stages should be used as standard reference units for both marine and terrestrial deposits (Shackleton and Opdyke, 1973). Various authors have used this marine record to date and correlate the episodes of soil development (Bronger and Heinkele, 1989; Bronger et al., 1998a,b; Markewich et al., 1998; Stremme, 1998; Olsen, 1998; Frechen, 1999; Dearing et al., 2001; Antoine et al., 2001).

According to the ages of the isotopic events in the low-latitude oxygen-isotope sequence (Bassinot et al., 1994), the deposit on which the heavily eroded soil NIG-5 developed probably formed during the last cold episodes between 11,000 and 71,000 BP (stages 4–2). Consequently, the parent material of soil NIG-4 should have formed in the former cold episode, between 127,000 and 186,000 BP (stage 6), and soil NIG-4 during the warm periods between 71,000 and 127,000 BP (stage 5). In addition, the formation of soil NIG-4, estimated by the clay accumulation index (Levine and Ciolkosz, 1983), must have begun around 85,000 BP or even earlier, given that erosion decreased the thickness of the Bt horizons. This supports the suggestion that this soil was formed during stage 5. The deposit on which soil NIG-3 developed probably formed during the cold episode between 242,000 and 301,000 BP (stage 8) and soil NIG-3 during the warm episode between 186,000 and 242,000 BP (stage 7). The Bt horizons of soil NIG-2 probably formed during the warm period between 301,000 and 334,000 BP (stage 9) and its parent material dates from the cold episode between 334,000 and 364,000 BP (stage 10). Finally, soil NIG-1 probably formed during the warm period between 364,000 and 427,000 BP (stage 11) and its parent material was probably deposited during the cold episode between 427,000 and 474,000 BP (stage 12). Consequently, the tilting of both deposits and soils NIG-2 and NIG-1, which were related to an uplift of Sierra Nevada, must have occurred around 300,000 BP in the Middle Pleistocene.

Based on the CI and Fe_dI indices (Fig. 3), the differences in the $Fe_t + Al_t/Si_t$ ratio between Bt and C horizons (Fig. 5), the extent of kaolinite neoformation (Table 4) and the micromorphological features, soils NIG-2 and NIG-1 show similar degrees of development although less than that of soil NIG-3 and greater than that of NIG-4. In addition, the duration of the warm periods in which these soils developed was around 63,000 years (NIG-1), 33,000 years (NIG-2) and 56,000 years (NIG-3 and NIG-4). Therefore, the time factor appears not to account for the different degrees of development of the buried soils, especially NIG-1, NIG-3 and NIG-4. The greater development of soil NIG-3 may therefore be attributed to a different, probably moister, climate. Greater moisture would also account for the leaching of carbonates from the Bt horizons of NIG-3 through the C horizon to form calcitic coatings in the Bt horizons of soil NIG-2. Consequently, in our region, the different degrees of soil development during the last 474,000 BP indicate that the wettest climate of the later Quaternary warm periods dates from between 186,000 and 242,000 BP (stage 7), and the driest from 71,000–127,000 BP (stage 5). The warm periods older than 242,000 BP (stages 9 and 11) probably had climates with intermediate wetness.

5.2. Surface soils

The parent materials of the surface soils, dating from the Early Pleistocene (between 788,000 and 1,650,000 BP; Birkeland, 1999), must have been deposited during one of the cold episodes before 788,000 BP, and the soils on them were formed during subsequent warm periods. The CI and Fe_dI indices (Fig. 3), the differences in the $Fe_t + Al_t/Si_t$ ratio between Bt and C horizons (Fig. 5), the extent of kaolinite neoformation (Table 4) and the micromorphological features were similar in all of the soils, indicating an equivalent degree of weathering and development. The minor differences in the indices of these soils could be attributed to parent-material differences or to waterlogging in some profiles. The greater weatherability of carbonate materials (limestones mainly) compared with metamorphic materials (mica schists and quartzites) could account for the slightly stronger development of soil COL, and the hydromorphic processes that affected soil LLP could explain its slightly weaker development.

The extent of development of the surface soils is similar to that of NIG-3, but the Fe_dI index of the latter is slightly less (Fig. 3). Nevertheless, the surface soils present two basic differences from soil NIG-3. First, most of the red clay coatings are fragmented and incorporated into the soil matrix, and second, a strong accumulation of $CaCO_3$ in the C horizons cements the mineral particles, forming the Ckm horizon. The fragmentation of the clay coatings suggests frost disturbance (Catt, 1987) and may be attributed to the cold episodes (Kemp, 1985; Van Vliet-Lanoë, 1985) following the formation of the Bt horizons. The origin of the large carbonate contents of the Ckm horizons of soils LLP and DUR, which were formed on a parent material similar to NIG-3 and also have decalcified Bt horizons, cannot be explained purely by leaching from the upper horizons, rather, this carbonate content must be attributed to the infiltration by the runoff water that is rich in Ca^{2+} and HCO_3^- ions. The presence of calcitic coatings in the Bt horizons of the surface soils indicates recalcification after the formation of the Bt horizons. This implies that $CaCO_3$ accumulation and cementation in the Ckm horizons increased over time as described in the soils of fluvial terraces (Dorronsoro and Alonso, 1994). However, the accumulation of $CaCO_3$ in the Ckm horizons was far greater in COL because this soil was surrounded by and formed over carbonate materials. Consequently, in periods after their formation, these soils were partially truncated, disturbed and recalcified to form polygenetic soils (Tarnocai and Valentine, 1989).

Because the extent of development of the surface soils is similar to that of NIG-3, we cannot rule out that it occurred during stage 7. Nevertheless, it could also have taken place during earlier warm episodes with similar climatic conditions to stage 7, such as stages 13 (between 474,000 and 528,000 BP) and 15 (between 568,000 and 621,000 BP). Bt horizons with clay illuviation are known to have formed elsewhere in these early interglacials (Bronger et al., 1998a).

5.3. Soil–time relationships

As the CI and Fe_dI indices, the differences in Fe_t+Al_t/Si_t ratio between Bt and C horizons, the extent of kaolinite neoformation and the micromorphological features of the soils formed during stage 7 (186,000–242,000 BP) are similar to surface soils formed on deposits of the Early Pleistocene, these features cannot be used to date surfaces older than 242,000 BP. In contrast, the degree of development of the soils formed in stage 5 and later is less than that in stage 7 and decreases progressively towards the youngest surfaces (Simón et al., 2000), showing a clear relationship between the degree of development and the age of the surfaces on which they formed. Consequently, these soils can be used for the approximate dating of landforms.

6. Conclusions

The depositional and soil development episodes during the Pleistocene were not continuous but were governed by pulses of tectonic uplift, giving rise to sedimentation, separated by periods of relative quiescence with soil development. From the Early to the early Late Pleistocene, the main pedogenic processes were the leaching of carbonates, weathering, illuviation and rubification, but the degree of development of the Bt horizons varied over time. The surface soils that formed over the deposits from the Early Pleistocene show the strongest development although in periods after their formation, they were partially truncated, disturbed and recalcified, resulting in polygenetic soils. The different degrees of development of the buried soils during the last 474,000 years indicate that the wettest warm period was stage 7 (186,000–242,000 BP), and the driest, stage 5

(71,000–127,000 BP). Stages 9 (301,000–334,000 BP) and 11 (364,000–427,000 BP) had climates with intermediate wetness. Given that the CI and Fe_dI indices, the differences in $Fe_t + Al_t/Si_t$ ratio between Bt and C horizons, the extent of kaolinite neoformation and the micromorphological features of the soils that were formed during stage 7 are similar to the surface soils that were formed on deposits of the Early Pleistocene, these features cannot be used to date surfaces older than 242,000 BP. However, from stage 7, the degree of soil development progressively declines with the decreasing age of the surfaces so that these soils can be used to estimate the age of landforms.

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