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Soil development over millennial timescales – a comparison of soil chronosequences of different climates and lithologies

D Sauer¹, I Schülli-Maurer², S Wagner³, F Scarciglia⁴, R Sperstad⁵, S Svendgård-Stokke⁵, R Sørensen⁶, G Schellmann⁷

¹Institute of Geography, Dresden University of Technology, Dresden, Germany

²Institute of Soil Science and Land Evaluation, Hohenheim University, Stuttgart, Germany

³Institute of Crop Science and Resource Conservation, Soil Science Division, University of Bonn, Germany

⁴Department of Biology, Ecology and Earth Sciences, University of Calabria, Arcavacata di Rende (CS), Italy

⁵The Norwegian Forest and Landscape Institute, Ås, Norway

⁶Department of Plant and Environmental Sciences, Norwegian University of Life Sciences, Ås, Norway

⁷Institute of Geography, University of Bamberg, Bamberg, Germany

E-mail: daniela.sauer@uni-hohenheim.de

Abstract. This paper reports soil development over time in different climates, on time-scales ranging from a few thousand to several hundred thousand years. Changes in soil properties over time, underlying soil-forming processes and their rates are presented. The paper is based on six soil chronosequences, i.e. sequences of soils of different age that are supposed to have developed under the similar conditions with regard to climate, vegetation and other living organisms, relief and parent material. The six soil chronosequences are from humid-temperate, Mediterranean and semi-arid climates. They are compared with regard to soil thickness increase, changes in soil pH, formation of pedogenic iron oxides (expressed as Fe_d/Fe_t ratios), clay formation, dust influx (both reflected in clay/silt ratios), and silicate weathering and leaching of base cations (expressed as $(Ca+Mg+K+Na)/Al$ molar ratios) over time. This comparison reveals that the increase of solum thickness with time can be best described by logarithmic equations in all three types of climates. Fe_d/Fe_t ratios (proportion of pedogenic iron Fe_d compared to total iron Fe_t) reflects the transformation of iron in primary minerals into pedogenic iron. This ratio usually increases with time, except for regions, where the influx of dust (having low Fe_d/Fe_t ratios) prevails over the process of pedogenic iron oxide formation, which is the case in the Patagonian chronosequences. Dust influx has also a substantial influence on the time courses of clay/silt ratios and on element indices of silicate weathering. Using the example of a 730 ka soil chronosequence from southern Italy, the fact that soils of long chronosequences inevitably experienced major environmental changes is demonstrated, and, consequentially a modified definition of requirements for soil chronosequences is suggested. Moreover, pedogenic thresholds, feedback systems and progressive versus regressive processes identified in the soil chronosequences are discussed.



1. Introduction

Assessment of soil-forming processes and their rates is an essential scientific base for several reasons. First, it reveals the time scales on which soil-forming processes operate and thus also the time required for soil regeneration after disturbance, erosion or degeneration. Soils are a main component of terrestrial ecosystems and the base of human and animal nutrition. Their extension on the earth's surface is limited, as the earth's surface area itself is limited, and because the major part of the earth's surface is covered by water or ice. If only soils are considered that are suitable for crop production, the area is further reduced by deserts and high mountain regions. The remaining soil area can hardly be increased by any measures (with few exceptions in the Netherlands and Australia's Sunshine Coast, where extreme efforts are undertaken to gain land from the sea). In fact, the area of fertile soils steadily decreases in support of housing, logistical and industrial constructions, and due to inadequate land-use or over-use, resulting in soil degradation and erosion. At the same time, the world's population continues to increase. Since soils are an essential and limited resource it is important to understand how they form, which processes are involved in their formation, at which rates these processes proceed and lead to measurable changes in soil properties, and finally, which factors may influence the direction and rates of soil-forming processes.

Second, the quantitative assessment of the direction and rates of soil-forming processes in different climates is an important tool for the correct paleo-environmental interpretation of paleosols. The environmental conditions and the time-span in which a paleosol developed, can only be retraced, if the reconstruction is based on profound actualistic studies on soil development over time under well-defined climatic and environmental conditions. Paleosols in turn are archives, in which responses of former soils to former climatic changes are documented. Therefore, their analysis and interpretation may help us understand causal relationships between climatic shifts and the responses of soils.

Third, quantitative studies of soil development over time provide a scientific base for modelling soil formation. Modelling soil formation, in turn, is an essential requirement to take the step from description and analysis of soil development towards prediction of future soil development. Such prediction is particularly important with regard to possible soil responses to the climatic shifts that are going on at present and expected in the near future. Soil responses to these shifts will probably be inconsiderable in the central regions of ecotones, such as in most parts of the humid-temperate climate of central Europe. In contrast, serious soil responses are to be expected in the peripheral regions of more fragile ecotones, such as in dry regions of the Mediterranean (e.g. on the Iberian Peninsula), which are supposed to become warmer and drier in the coming decades and that are already at the edge of desertification. Modelling soil responses to climatic changes may identify areas at particular risk of soil degradation; hence measures (e.g. extensification and adaption of land-use) may be taken in time to prevent desertification.

Since it is impossible to directly observe soil formation over thousands of years, the direction and rates of soil-forming processes are usually assessed using soil chronosequences, i.e. sequences of soils of different ages that are supposed to have developed under similar conditions with regard to climate, vegetation and other living organisms, relief, and parent material. Important research on soil chronosequences has been done in the Mediterranean climate of California, mainly from the 1970-1980's onwards, by Harden [1, 2], who examined the Merced River soil chronosequence in central California. There, Harden also developed the profile development index (PDI) that has since been widely applied. Other significant soil chronosequence studies in California include those of McFadden and Hendricks [3] who analysed pedogenic iron forms in Holocene to middle Pleistocene soils on fluvial deposits; Busacca [4] who studied soils ranging from 600 yr to 1.6 Ma in age at Honcut Creek, Sacramento Valley; Muhs [5] who investigated a soil chronosequence on Quaternary marine terraces (< 3 ka to > 1 Ma), on San Clemente Island; and Aniku and Singer [6] who analysed pedogenic iron in a soil chronosequence on marine terraces (105-600 ka). Soil chronosequences in semi-arid climates were extensively studied in the southern Great Basin (USA), especially at Silver Lake Playa, Cima Volcanic Field, Kyle Canyon, and Fortymile Wash, by McFadden et al. [7, 8]; Reheis et al. [9]; and Harden et al. [10].

In the Mediterranean region of Europe, soil chronosequences were investigated mainly in southern Spain and Italy. Important studies were done by Alonso et al. [11] and Dorronsoro and Alonso [12] on soils on Holocene to Pleistocene fluvial terraces in Spain, and by Scarciglia et al. [13] on soils on Quaternary marine terraces in Calabria (southern Italy).

Soil chronosequences in temperate- and cool-humid climates were studied both in North America and Europe. Studies in North America include those of Singleton and Lavkulich [14] who investigated soils on beach sand on Vancouver Island; Barrett and Schaetzl [15] and Barrett [16] who examined podzolisation in sandy terraces and beach ridges at Lake Michigan; Alexander and Burt [17], and Burt and Alexander [18] who analysed soils on moraines of Mendenhall Glacier, SE Alaska. European soil chronosequence studies comprise those of Arduino et al. [19] who analysed iron oxides and clay minerals in soils on fluvial terraces in northern Italy; Bain et al. [20] who investigated soils on river terraces ranging in age from 80-13,000 years in Scotland; and Mokma et al. [21] who studied a Holocene Podzol chronosequence in Finland. Sequences of moraine ridges in mountain regions, e.g. in the Sierra Nevada (California) and in the European Alps, have also been used for studying soil chronosequences. Birkeland and Burke [22] investigated soil catena chronosequences on moraine ridges in the eastern Sierra Nevada, California, and Egli et al. [23, 24, 25] who analysed Holocene soil chronosequences in the Swiss Alps.

In contrast to the above-mentioned Mediterranean and temperate to cool climates, where numerous soil chronosequences have been studied over the last four decades, soil chronosequence studies in tropical regions and some extreme environments such as cold deserts are rare. The few existing studies include those of Pillans [26] who investigated soils on basaltic lava flows with ages ranging from 10 ka to 5.59 Ma in the tropical climate of northern Queensland (Australia), and Muhs [27] who studied soils on Pleistocene reef terraces of Barbados. These soils formed in Sahara dust, volcanic ash from the Lesser Antilles island arc, and detrital carbonate from the underlying reef limestone. A soil chronosequence in a cold desert environment was analysed by Bockheim [28] who studied soils on moraines ranging from ca. 6 ka to 250 ka in age in the Transantarctic Mountains.

This paper compares the direction and rates of soil-forming processes in humid-temperate, Mediterranean and semi-arid climates. It aims at contributing to the knowledge on natural soil-forming processes that is required for the three main purposes mentioned in the beginning of the introduction. For this purpose, six soil chronosequences are presented. In addition to the comparison of three climatically different regions, also two diverse pathways of soil formation in different parent materials in the same region in humid-temperate climate are included (Figure 1). The sites for these soil chronosequences were selected trying to avoid considerable human influence. Man has become a major soil-forming factor [29]. This factor is excluded here as far as possible, in order to ensure a good understanding of (semi-) natural soil formation first, as a base for unambiguous identification of the effect of the soil-forming factor man, separately. However, a certain extent of human influence could not be avoided, especially in the Mediterranean soil chronosequence where the soils are largely influenced by long-term agricultural use and related erosion.

Similarities and differences between the soil chronosequences in terms of identified soil-forming processes and their rates are discussed. Using the example of the Mediterranean soil chronosequence from southern Italy, comprising ca. 730 ka, the fact that soils of long chronosequences inevitably experienced major environmental changes is demonstrated, and, consequentially a modified definition of requirements for soil chronosequences is suggested.

2. Material and methods

Three soil chronosequences were studied in southern Norway. Two of them were on marine loamy sediments in the provinces Vestfold and Østfold, on both sides of the Oslo Fjord, and one was on beach sand in Vestfold (Tables 1-3). These areas have been subject to continuous glacio-isostatic upward movement, since the ice sheet in the southern Oslo Fjord region melted between ~13000 and 12000 calendar years ago. As a consequence of the steady uplift since the weight of the ice had gone,

more and more parts of formerly submerged areas have risen above sea-level. Thus, the age of the land surface continuously increases from the coast towards the inland. Since the two chronosequences on loamy sediments in Vestfold and Østfold are very similar with respect to all soil properties discussed here, they have been merged into one sequence comprising 12 pedons in this paper. More details on soil properties and analytical data of the Norwegian soil chronosequences were previously reported [30, 31, 32, 33, 34]. A Mediterranean soil chronosequence was studied at the Gulf of Taranto near Metaponto, on the Ionian coast of Basilicata, southern Italy (Table 4). This area is subject to tectonic uplift due to the collision of the African and Eurasian plates. The interplay between land uplift and sea level fluctuations due to alternating glacial and interglacial periods during the Pleistocene led to a sequence of 11 marine terraces [35]. Details on the age estimates of the terraces and on the soils were previously published [36]. Two soil chronosequences in semi-arid climate were located on Holocene beach ridges along the eastern coast of Patagonia, in southernmost South America, specifically in the northern part of Golfo San Jorge. Soil data of these sequences were reported in [37].

This paper thus compares soil chronosequences that were individually described in a number of publications with regard to observed changes in soil properties over time, including soil thickness increase, changes in soil pH, formation of pedogenic iron oxides (expressed as Fe_d/Fe_t ratios), clay formation, dust influx (both reflected in clay/silt ratios), and silicate weathering and base leaching (expressed as $(Ca+Mg+K+Na)/Al$ molar ratios) over time.

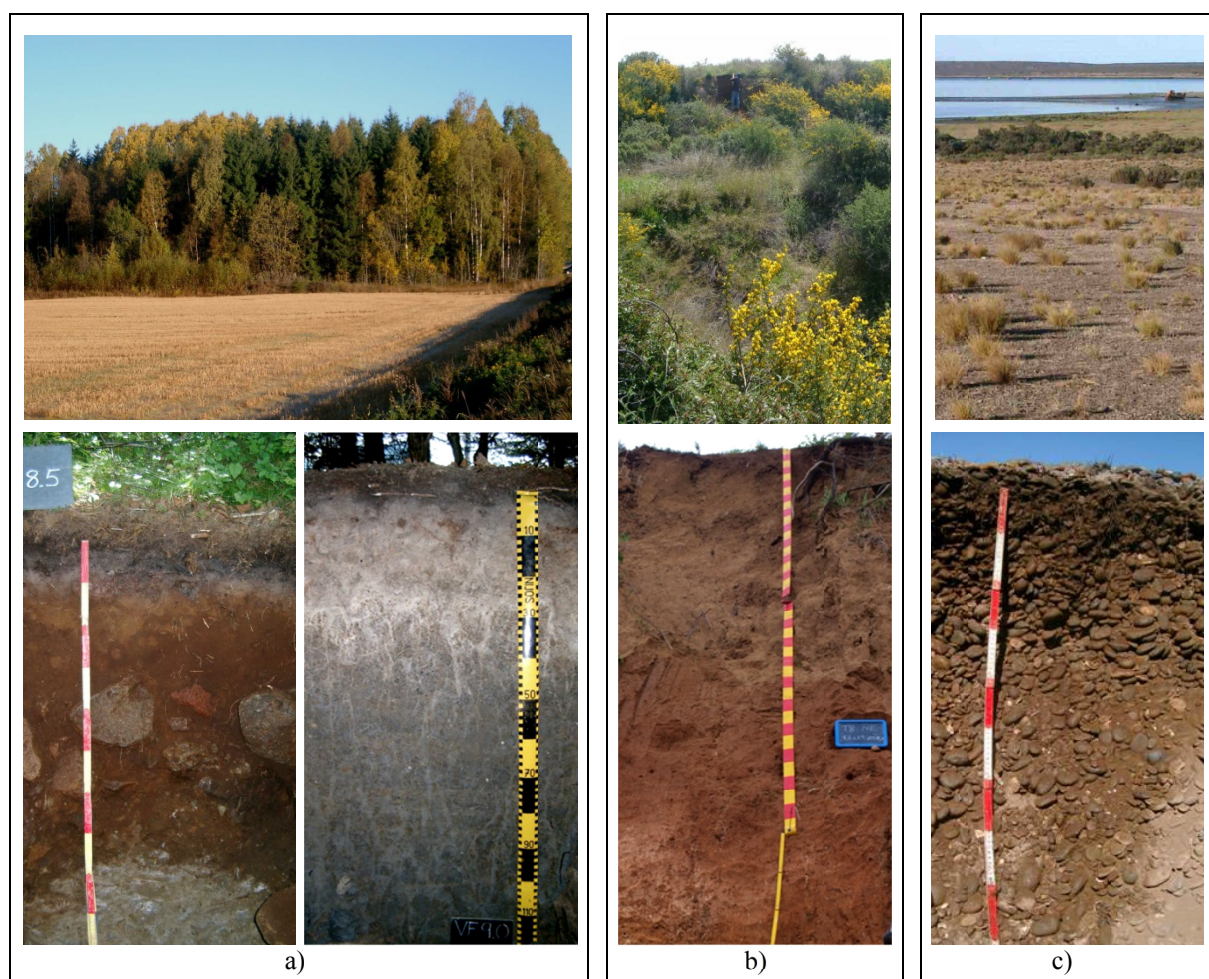


Figure 1. Soil development a) in the humid-temperate climate of S Norway (left: Podzol development on beach sand, right: Albeluvisol development on loamy marine sediments), b) in the Mediterranean climate of southern Italy and c) in the semi-arid climate of eastern Patagonia.

All soils were described according to FAO [38] and classified according to WRB [39] (Tables 1-6).

Table 1. Soil chronosequence on beach sand in Vestfold, southern Norway (MAT 6°C; MAP 975 mm)

Pedon (no., name, location)	Age¹	Horizon sequence (depth in cm, horizons according to FAO [38])	Classification according to WRB [39]
Pedon 1 SVF2.4 (Gleåbukta)	2 300 ± 150	organic layer: 0-+5 Oi; 0-8 Ah1; 8-25 Ah2; 25-46 Bw; 46-56 BC; 56-115 Cg1; 115-135 2Cg2	Endoskeletal Brunic Endostagnic Umbrisol (Arenic, Colluvic)
Pedon 2 SVF3.0 (Gleåbukta)	3 800 ± 150	organic layer: 0-+3 Oa; +3-+8 Oe; +8-+9 Oi; 0-7 AE; 7-16 Bw1; 16-21 Bw2; 21-37 BC; 37-47 Ahb; 47-77 BC; 77-124 BCg; 124-138 2Cg; 138-180 Cr	Brunic Arenosol (Colluvic, Protosodic)
Pedon 3 SVF4.0 (Haraldsrød)	4 300 ± 200	organic layer: 0-+0.5 Oe; +0.5-+3.5: Oi 0-4 Ah; 4-9 AE; 9-22 Bh; 22-36 Bs; 36-50 Bw; 50-90 BCg; >90 2R	Endostagnic Endoleptic Cambisol (Protosodic)
Pedon 4 SVF6.0 (Jåberg)	6 600 ± 170	organic layer: 0-+3.5 Oa; 3.5-+9 Oe; +9-+13 Oi; (0-5 AE) ² ; 0/5-13 EA; 13-26 Bhs; 26-54 Bs; 54-74 BC; 74-85 Cg	Endostagnic Folic Podzol
Pedon 5 SVF7.2 (Grønneberg)	7 650 ± 130	organic layer: 0-+2 Oa; +2-+6.5 Oe; +6.5-+7.5 Oi; 0-10 AE; 10-24 Bsh; 24-40 Bs1; 40-65 Bs2; 65-105 BCg; 105-140 2Cg	Endostagnic Podzol
Pedon 6 SVF8.5 (Rauan)	9 650 ± 100	organic layer: 0-+3.5 Oa; +3.5-+8.5 Oe; +8.5-+10 Oi 0-5 AE; 5-7 E; 7-30 Bs; 30-76 2Bs2; 76-105 3Cg	Endoskeletal Endostagnic Folic Podzol (Ruptic)

¹ ages derived from calibrated ¹⁴C datings, given in calendar years before sampling, not BP (i.e. before 1950)

² discontinuous horizon

Table 2. Soil chronosequence on loamy marine sediments in Vestfold, southern Norway (MAT 5.3-6.6 °C; MAP 880-1075 mm)

Pedon (no., name, location)	Age¹	Horizon sequence (depth in cm, horizons according to FAO [38])	Classification according to WRB [39]
Pedon 1 VF2.4 (Sem)	1 650 ± 120	0-5 Ah; 5-13 BE; 13-20 E1; 20-40 E2; 40-56 Btg1; 56-95 Btg2; 95-130 Ctr; 130-155 Cr	Luvic Endogleyic Stagnosol (Siltic)
Pedon 2 VF4.5 (Ramnes)	4 600 ± 70	0-9 Ah; 9-21 Ap; 21-27 E; 27-40 Beg; 40-70 Btg1; 70-100 Btg2; 100-135 Btg3; 135-170 Cg; 170-230 C	Luvic Stagnosol (Siltic)
Pedon 3 VF6.6 (Fossan)	6 200 ± 100	0-9 Ah; 9-15 Ap; 15-30 BE; 30-42 E; 42-56 E/Bg; 56-91 Btg1; 91-195 Btg2; 195-230 C	Luvic Glossic Albic Fragic Stagnosol (Siltic)
Pedon 4 VF8.8 (Holmen)	6 550 ± 200	0-8 Ah; 8-20 Ap; 20-36 BE; 36-40 E; 40-66 E/Bg; 66-90 B/Eg; 90-113 Btg1, 113-155 2Btg2; 155-175 3Cg	Luvic Glossic Albic Fragic Stagnosol (Siltic)
Pedon 5 VF7.3 (Gjein)	8 100 ± 120	0-4 Ah; 4-18 BE; 8-27 E; 27-40 E/Bg; 40-58 Btg1; 58-105 Btg2; 105-190 Cg	Luvic Glossic Albic Fragic Stagnosol (Endeutric)
Pedon 6 VF9 (Torp)	9 000 ± 150	0-6 AE; 6-9 Bs; 9-24 BE; 24-27 E; 27-49 E/Bg; 49-85 Btg; 85-156 Bg; 214 Cg	Luvic Glossic Albic Fragic Stagnosol (Protosodic)

¹ ages derived from calibrated ¹⁴C datings, given in calendar years before sampling, not BP (i.e. before 1950)

Table 3. Soil chronosequence on loamy marine sediments in Østfold, southern Norway (MAT 4.6-6.4 °C; MAP 770-880 mm)

Pedon (no., name, location)	Age¹	Horizon sequence (depth in cm, horizons according to FAO [38])	Classification according to WRB [39]
Pedon 1 ØF3 (Løkkevika)	3 000 ± 250	0-8 Ah; 8-30 BE; 30-40 E; 40-80 Btg; 80-110 Bg; 110-155 Cr1; 155-210 Cr2	Hyperdystric Alic Endogleyic Stagnosol (Siltic)
Pedon 2 ØF4 (Tomb)	3 500 ± 200	0-8 Ah; 8-25 Ap; 25-40 E; 40-77 Btg; 77-116 Btrg; 116-150 Cr	Luvic Albic Endogleyic Stagnosol (Endosiltic)
Pedon 3 ØF7.5 (Husevja)	6 550 ± 150	0-3 Ah; 3-12 AE; 12-26 Bs; 26-29 BE; 29-38 E; 38-66 E/Bg; 66-80 Btg; 80-105 Cg	Luvic Glossic Albic Fragic Stagnosol (Episiltic, Protospodic)
Pedon 4 ØF5 (Navestad)	6 650 ± 150	0-6 Ah; 6-16 Ap; 16-25 E; 25-40 E/Bg; 40-66 Btg1; 66-90 Btg2; 90-170 Cr1; 170-185 Cr2	Luvic Glossic Albic Endogleyic Fragic Stagnosol (Endoeutric)
Pedon 5 ØF8 (Os Kirke)	9 750 ± 150	0-5 Ah; 5-13 EB; 13-30 E1; 30-40 E2; 40-70 E/Bg; 70-127 Btg; 127-175 BCg; 175-195 Cg	Luvic Glossic Albic Fragic Stagnosol (Endoeutric, Siltic)
Pedon 6 ØF11 (Båstad)	11 050 ± 150	0-6 Ah; 6-10 AE; 10-21 Bs; 21-37 E; 37-60 E/Bg; 60-77 Btg; 77-90 BCg1; 90-100 BCg2; 100-175 BCg3; 175-195 C	Luvic Glossic Albic Fragic Stagnosol (Endofluvic, Siltic, Protospodic)

¹ ages derived from calibrated ¹⁴C datings, given in calendar years before sampling, not BP (i.e. before 1950)

Table 4. Soil chronosequence on marine terraces near Metaponto, Gulf of Taranto, southern Italy (MAT 16.2 °C; MAP 456 mm at Marina di Ginosa)

Pedon (no., name, location)	Age¹	Horizon sequence (depth in cm, horizons according to FAO [38])	Classification according to WRB [39]
Pedon 1 T0; Lido di Metaponto	190 ¹	0-2 C; 2-13 AC; 13-31 CB1; 31-46 2CB2; 46-67 Cw1; 67-95 Cw2; 95-158 C	Calcaric Brunic Arenosol (Ochric)
Pedon 2 T1; Petrulla (Loess)	16 000 ¹	0-30 Ap; 30-60 Bw; 60-120 Bk1; 120-185 Bk2; 185-240 2Bk3; 240-450 3C	Endocalcic Luvisol (Loamic)
Pedon 3 T2; San Teodoro I	100 000	0-23 Ap; 23-60 Bt1; 60-93 Bt2; 93-130 Ck1; 130-180 Ck2; 180-280 2Bk1; 280-325 3Bk2; 325-350 4Bk3; 350-400 Ckm; >400 5Cw	Endocalcic Luvisol (Cutanic)
Pedon 4 T3; San Teodoro II	120 000 ²	0-25 Ap; 25-70 Ap/Bt (tilted); 70-100 Bt/Ap (tilted); 100-112 Bt; >112 Bk	Bathicalcic Luvisol (Cutanic)
Pedon 5 T4; Marconia	195 000	0-25 Ap1; 25-42 Ap2; 42-58 Bt1; 58-116 2Btk1; 116-133 Btk2; 133-151 Bt2; 151->223 3Bt3	Endocalcic Alisol (Clayic, Cutanic)
Pedon 6 T5-SE	310 000	0-30 Ap; 30-54 Bt1; 54-83 Bt2; 83-108 Btk1; 108-174 2Bt3; 174->206 Btk2	Endocalcic Alisol (Cutanic)
Pedon 7 T5-NW	330 000	0-33 Ap; 33-80 Bt; 80-127 2Btk1; 127-168 Btk2; 168->205 Btk3	Endocalcic Alisol (Cutanic)
Pedon 8 T6; Tinchi I	405 000	0-20 Ap; 20-36 Bt1; 36-50 2Bt2; 50-108 Bt3; 108-165 Bw1; 165-182 3Bw2; 182-215 Cw; 215->250 C	Chromic Luvisol (Clayic, Cutanic)
Pedon 9 T7; Tinchi II	500 000	0-33 Ap; 33-54 Bt1; 54-73 Bt2; 73-160 2Bt3; 160-290 Bt4; 290-330 Bw; 330-338 Bg; 338-380 Ck/Ckm; 380-410 3C	Chromic Luvisol (Cutanic)
Pedon 10 T8; Pisticci	575 000 ³	0-22 Ap; 22-47 AB; 47-65 2Btk1; 65-109 Btk2; 109-200 Btk3; 200-239 Btk4; 239-259 Btk5; 259-286 CB1; 286-323 CBk; 323->380 CB2	Chromic Alisol (Cutanic)
Pedon 11 T9; Bernalda	670 000	0-10 Ah; 10-26 Ah/Bw; 26-70 Bt; 70-100 Btk1; 100-117 Btk2; 117-140 Btk3; 140-220 Btk4; 220->230 2Bw	Chromic Luvisol (Cutanic)

¹ OSL dating from [40]; ² age based on occurrence of Senegalese fauna (Eemian); age based on tephra chronology; all other ages estimated according to Pleistocene glacial-interglacial cycles and related sea level fluctuations

Table 5. Soil chronosequence on beach ridges with substantial dust accumulation along the Patagonian East coast (MAT 12.6 °C; MAP 287 mm)

Pedon (no., name, location)	Age ¹	Horizon sequence (depth in cm, horizons according to FAO[38])	Classification according to WRB [39]
Pedon 1 R1 (Bust.); Pa'04/12b	2241 ± 46	0-1 (dp); 1-4 Ah; 4-8 AC1; 8-18 AC2; 18-30 C(h); 30-42 C(h)k; 42-50 2Ck1; 50-60 2Ck2; 60-73 2Ck3; 73-100 3C	Calcaric Skeletic Regosol (Protocalcic, Yermic)
Pedon 2 R2 (Cam.); Pa'04/8	3372 ± 48	0-1 (dp); 1-4 Ah; 4-26 AC; 26-55 C(h)k; 55-65 2Ck; 65-75 2C	Calcaric Skeletic Regosol (Protocalcic, Yermic)
Pedon 3 R3 (Bust.); Pa'04/4e	4052 ± 50	0-1 (dp); 1-4 Ah1; 4-17 Ah2; 17-44 AC; 44-65 C(h)k1; 65-80 2C(h)k2; 80-87 2Ck1; 87-100 2Ck2; 100-140 2C	Calcaric Skeletic Regosol (Protocalcic, Yermic)
Pedon 4 R4 (Bust.); Pa'04/2	6238 ± 51	0-1 (dp); 1-5 Ah; 5-20 AC; 20-60 C(h); 60-95 2C1; 95-100 2Ck1; 100-135 2C2; 135-139 2Ck2; 139-160 2C3; 160-164 2Ck3; 164-176 2C4	Calcaric Skeletic Regosol (Protocalcic, Yermic)

¹ ages are uncalibrated radiocarbon ages of molluscs from each beach ridge

Table 6. Soil chronosequence on beach ridges with minor dust accumulation along the Patagonian East coast (MAT 12.6 °C; MAP 287 mm)

Pedon (no., name, location)	Age ¹	Horizon sequence (depth in cm, horizons according to FAO [38])	Classification according to WRB [39]
Pedon 1 L1 (Cam.); Pa'04/6	1376 ± 47	0-3 (dp) ³ ; 3-5 Ah; 5-8 AC; 8-22 C(h); 22-50 2C	Calcaric Hyperskeletal Leptosol (Yermic)
Pedon 2 L2 (Cam.); Pa32	2618 ± 92	0-1 (dp); 1-4 Ah; 4-25 AC; 25-30 2Ck; 30-60 2C	Calcaric Hyperskeletal Leptosol (Protocalcic, Yermic)
Pedon 3 L3 (Bust.); Pa'04/1b	5400 ± 25	0-2 (dp); 2-7 Ah; 7-20 C(h)1; 20-50 C(h)2; 50-80 Ck; 80-100 C	Calcaric Hyperskeletal Leptosol (Protocalcic, Yermic)

¹ ages are uncalibrated radiocarbon ages of molluscs from each beach ridge

2. Comparison of soil property changes in the different chronosequences

2.1. Soil thickness increase

Soil thickness was calculated according to [12] as thickness of A horizons + thickness of B horizons + ½ thickness of transitional AC/CA/BC/CB horizons. Soil thickness increase over time can be best described by logarithmic functions in all studied soil chronosequences (Figure 2). The rates of increase in soil thickness are clearly influenced by climate and parent material. The highest rate is observed for the Mediterranean climate (Figure 2; green triangles). The humid-temperate climate of southern Norway allows for medium rates of soil thickness increase over time, whereby rates in the loamy marine sediments (Figure 2; red squares) are higher than those in the sandy beach deposits (Figure 2; orange squares). The lowest rates of soil thickness increase are found in the semi-arid climate of Patagonia (Figure 2; green diamonds), where the mean annual temperature (12.6 °C) is in between those of the two other regions, but the low amount of precipitation (MAP = 287 mm) limits the rates of weathering and primary plant and animal productivity, and increase in soil thickness.

In general, it was attempted to establish all soil profiles comprised in the chronosequences in flat positions, showing minimum erosion. This approach worked well in the Holocene chronosequences in Norway and Patagonia, whereas in the Holocene-Pleistocene chronosequence in southern Italy, spanning ~730 ka, it was impossible to include only soils that were not or only slightly eroded because erosion was ever-present. A theoretical function of increase in soil thickness over time was obtained by including only maximum soil thicknesses (Figure 3). However, very few of the soils included in

this curve were really not eroded, hence without erosion, still higher rates of soil thickness increase would be observed.

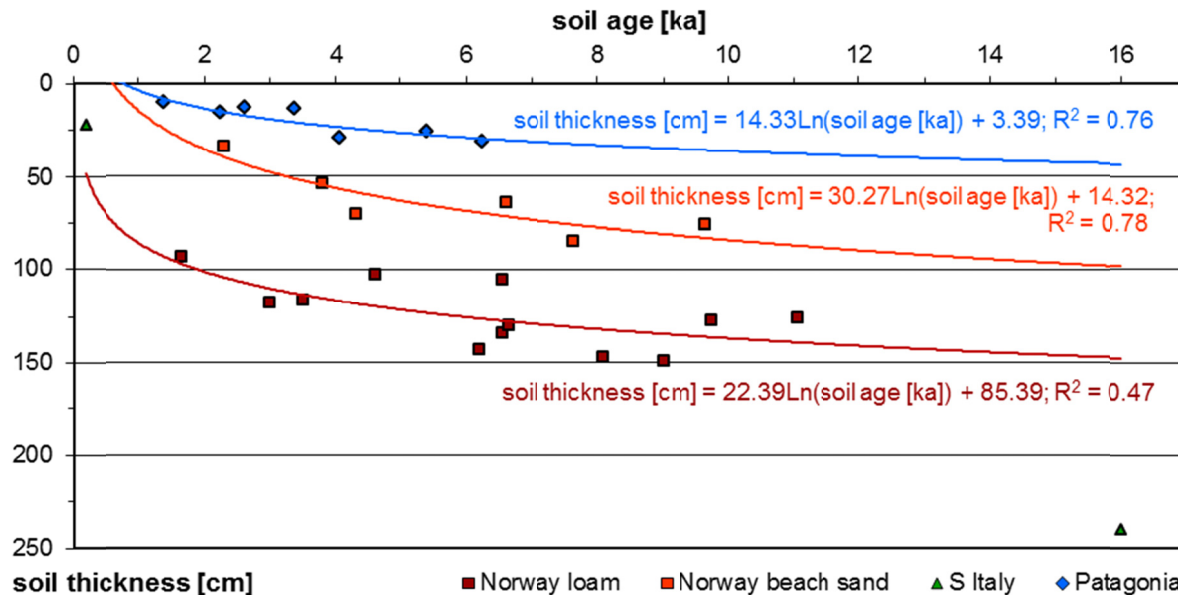


Figure 2. Soil thickness increase over a time-span of 16 ka in the studied soil chronosequences. The two soil chronosequences on loamy sediments in Norway and on gravelly beach ridges in Patagonia could be merged with regard to soil thickness.

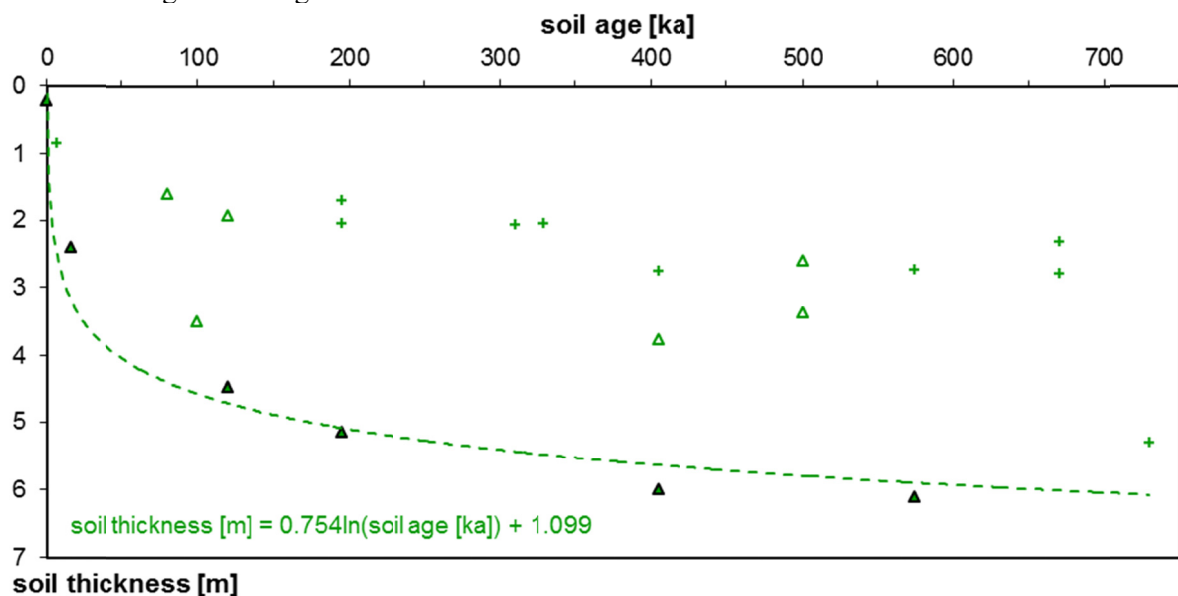


Figure 3. Soil thickness increase observed in the Italian soil chronosequence over a time-span of 730 ka. Open triangles = eroded soils; + = C horizon not reached.

2.2. Soil pH

Weighted mean soil pH values of the upper 50 cm of the soils were calculated by taking the anti-log of the pH value for each horizon down to 50 cm depth and multiplying the resulting proton concentration by horizon thickness; the products were summed up and divided by 50.

The different soil moisture regimes of the study areas are clearly reflected in the pH levels (Figure 4). The study areas in the humid-temperate climate in Norway are characterised by distinct leaching

environments. Although pedogenesis follows different pathways in the loamy marine sediments and sandy beach deposits, the pH values are very similar in the Norwegian soil chronosequences (Figure 4; orange and red squares). Within a very short period of time carbonates, which may be present in the sediments in the beginning (mainly in the form of shell fragments), are dissolved. This process is favoured by the presence of sulphides in the loamy sediments, and by the high water conductivity in the sandy sediments, respectively. After complete removal of carbonates, pH drops rapidly to around pH 4 and stays rather constant thereafter, due to buffering by adsorption of protons to clay minerals and proton consumption in weathering and clay mineral formation. Due to these buffering systems, after ~2 ka, pH in the Norwegian soils is influenced by the local vegetation rather than soil age.

The Mediterranean climate, in which the Italian soil chronosequence is located, corresponds to a xeric soil moisture regime. This means that leaching conditions prevail during the moist winter months, whereas there are no leaching conditions in the dry summers. Such an environment should, over a long time-span, gradually lead to decarbonatisation (removal of carbonates) and pH drop but at distinctly lower rates than in a humid climate. However, there is another important factor counteracting decarbonatisation; the deposition of considerable amounts of dust characteristic in the Mediterranean region, which leads to accumulation of carbonates and re-carbonatisation of the soils. The rates of dust influx must have varied through time. Well-developed Bt horizons with thick illuvial clay films, which are typical for Mediterranean soils on calcareous parent materials (including those comprised in the Italian chronosequence), indicate that there must have been periods with low dust influx, allowing for complete decarbonatisation and clay illuviation. On the other hand, Mediterranean soils often show accumulation of secondary carbonates within previously decarbonated horizons, indicating phases of increased influx of calcareous dust (Figure 5).

pH in the upper 50 cm

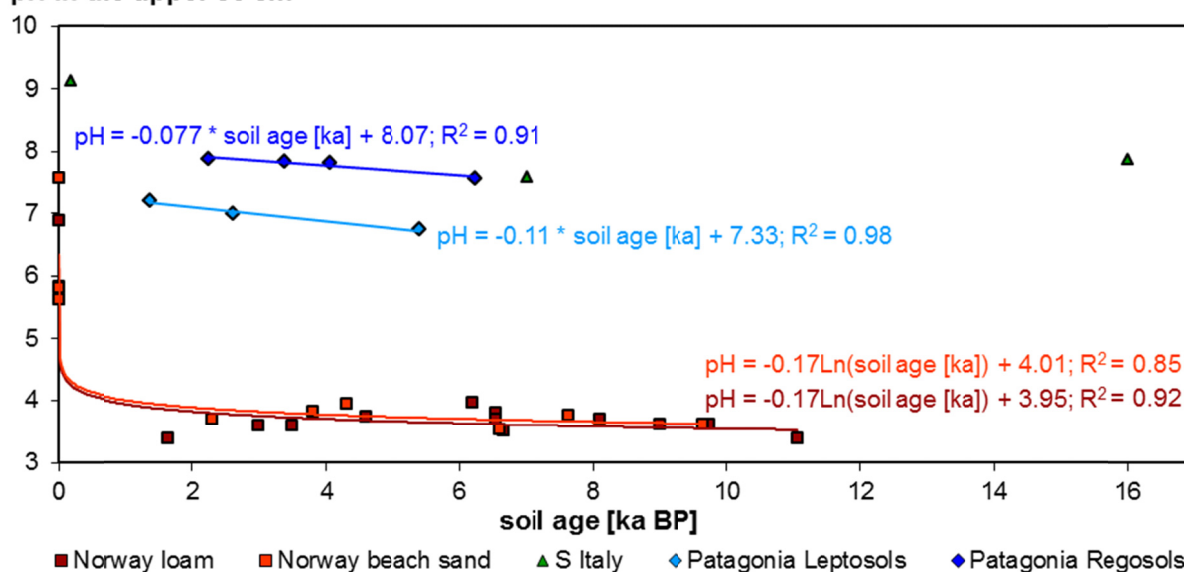


Figure 4. Soil pH, weighted means of the upper 50 cm (soil:solution ratio = 1:2.5; measured in KCl for Norway and Italy and in CaCl₂ for Patagonia). The two soil chronosequences on loamy sediments in Norway could be merged with regard to soil pH.

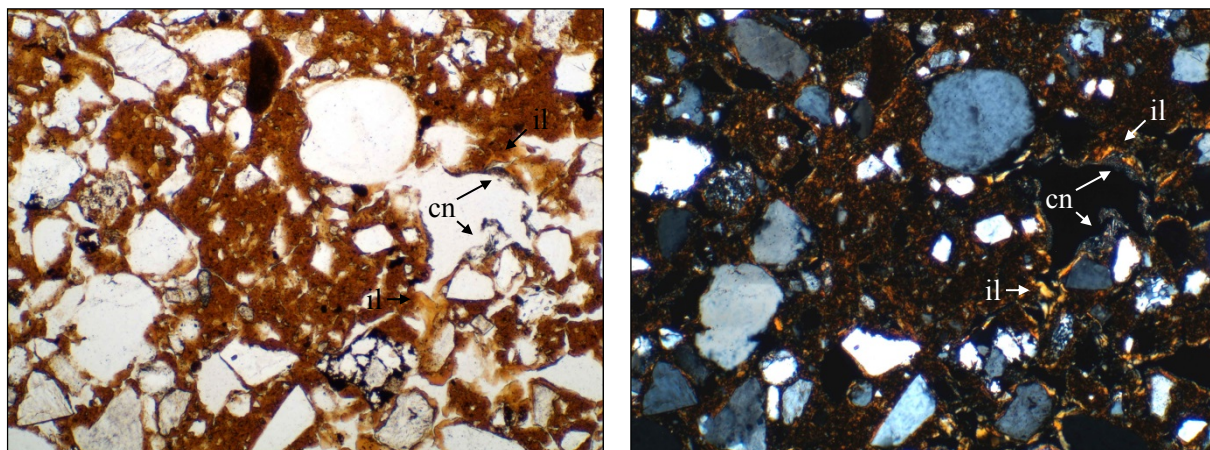


Figure 5. Thin section photograph of profile Bernalda, terrace T9, 82.5-86.5 cm depth (width = 2.2 mm; left with plane polarised light, right with crossed polarisers). The middle Pleistocene soils of the Italian chronosequence are characterised by neo-formed reddish brown clay and illuvial clay (il). This soil (ca. 670 ka-old) moreover shows accumulation of secondary calcite needles (cn) in many voids.

Because of the xeric moisture regime and variable re-carbonatisation through dust influx, the general pH trend with time in the Metaponto chronosequence shows a slow decrease with some scatter and even increased pH values in the older soils of the sequence (Figure 6). Pleistocene alternations between Mediterranean forest (during interglacial periods) and steppe environments (during glacial periods) in southern Italy [41, 42] must have been accompanied by alternating leaching and non-leaching or slight leaching conditions. However, the influence of paleo-environmental changes on the time courses of soil chemical parameters is difficult to reconstruct. It is likely that the rates of decarbonatisation and pH decrease in the upper 50 cm were considerably decreased and even temporarily reversed during glacial periods with steppe environments and enhanced dust influx rates. Also the environmental conditions within glacial periods were not constant but included, for example shifts between *Artemisia* steppe and forest steppe, which must have been related to different soil moisture regimes and hence different carbonate dynamics.

The Patagonian soil chronosequences are located in the driest environment included in this work. However the gravelly beach ridges, on which the soil chronosequences in Patagonia were established, are highly water permeable. Therefore, decarbonatisation is taking place in the upper 20 to 40 cm of these soils, and pH decreases towards the soil surface, indicating that some leaching processes are active. Both chronosequences show a slow but distinct pH decrease with time, whereby a clear difference in pH levels is observed between the soil chronosequence on beach ridges with > 10 % fine earth, i.e. the Regosol sequence (Figure 4; dark blue diamonds), and the sequence on beach ridges with \leq 10 % fine earth in the upper 75 cm, i.e. the Leptosol sequence (Figure 4; light blue diamonds). The pH values of the Regosols are 0.8 to 0.9 pH units higher than those of the Leptosols. The higher amounts of fine earth in the Regosols point to greater dust accumulation at these sites, compared to the Leptosol sites; it can thus be concluded that the higher pH level of the Regosol sequence is also due to the higher input of calcareous dust.

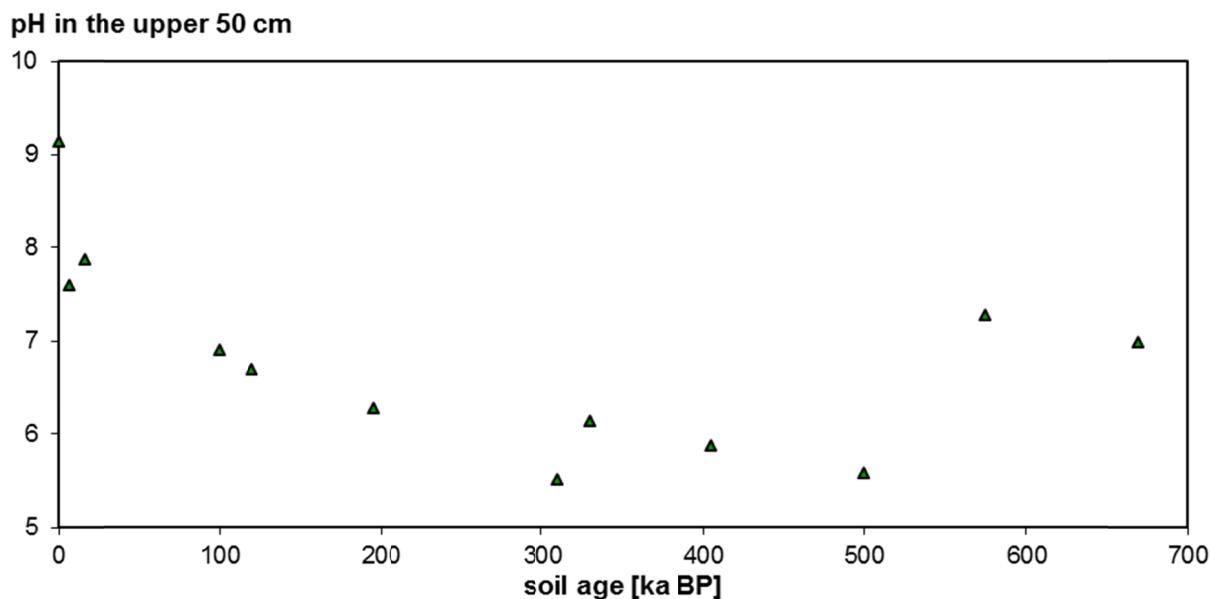


Figure 6. Soil pH in KCl (weighted means of the upper 50 cm) over the whole time-span of the Italian soil chronosequence.

2.3. Formation of pedogenic iron oxides

The formation of pedogenic iron oxides in the different soil chronosequences is compared by calculating the weighted average Fe_d/Fe_t ratio (ratio of dithionite-extractable iron, analysed by ICP-OES and total iron, analysed by X-ray fluorescence analysis) in the upper 50 cm of the soils. For this purpose, the Fe_d/Fe_t ratio of each horizon down to 50 cm depth was multiplied by horizon thickness, the results were summed up and divided by 50.

The Norwegian soil chronosequences show increasing trends of pedogenic iron oxides over time, which can be described by linear functions (Figure 7; orange and red squares). The Italian chronosequence shows a clear increase in Fe_d/Fe_t ratios in the first ~120 ka, whereas Fe_d/Fe_t ratios in the older soils are highly variable (Figure 8). Decreased Fe_d/Fe_t values (open triangles in Figure 8) are interpreted to correspond to severely eroded soils, since erosion removes the uppermost, most weathered parts of the profile, thus exposing lower, less weathered parts of the soil having lower Fe_d/Fe_t ratios. A logarithmic curve can be fitted to the maximum Fe_d/Fe_t values of the Italian chronosequence. It is concluded that: 1) the contents of pedogenic iron oxides in the middle Pleistocene soils of the Italian chronosequence are influenced to variable degree by erosion; and 2) the time course of the progressive formation of pedogenic iron oxides can be described by a logarithmic equation. However, it has to be considered that this equation averages over higher rates of pedogenic iron oxide formation in interglacial periods and lower rates in glacial periods.

The Patagonian soil chronosequences exhibit a reverse development of pedogenic iron oxide contents; their Fe_d/Fe_t ratios decrease with soil age (Figure 7; blue diamonds). This unusual time course of Fe_d/Fe_t probably reflects proceeding accumulation of dust in the semi-arid environment of these soils. Central eastern Patagonia is characterised by strong winds and sparse vegetation cover. This combination of environmental factors leads to enhanced aeolian activity. Dust that settles down on the gravelly beach ridges, on which the soil chronosequences were established, is easily trapped in the interstitial voids between the pebbles and accumulates in the soils. Dust accumulation leads to dilution of the soils with regard to Fe_d/Fe_t ratios.

2.4. Clay/silt ratios

Clay/silt ratios can be used in soil chronosequences on homogenous parent materials to assess progressive clay formation with time. However, variations in the primary clay/silt ratio of sediment

layers due to slight changes in the depositional conditions (e.g. water velocity) may limit the applicability of this approach. The soil chronosequences comprised in this paper are compared with regard to the weighted mean clay/silt ratios in the upper meter where possible (Norway, Italy), because the depth of 50 cm that has been used for pH and Fe_d/Fe_t ratios is not useful for the clay/silt ratio of soils that are characterised by clay illuviation. The clay/silt ratios of the Patagonian chronosequences, however, refer to the upper 50 cm because not all pedons of these sequences were sampled down to one meter depth.

Fe_d/Fe_t in the upper 50 cm

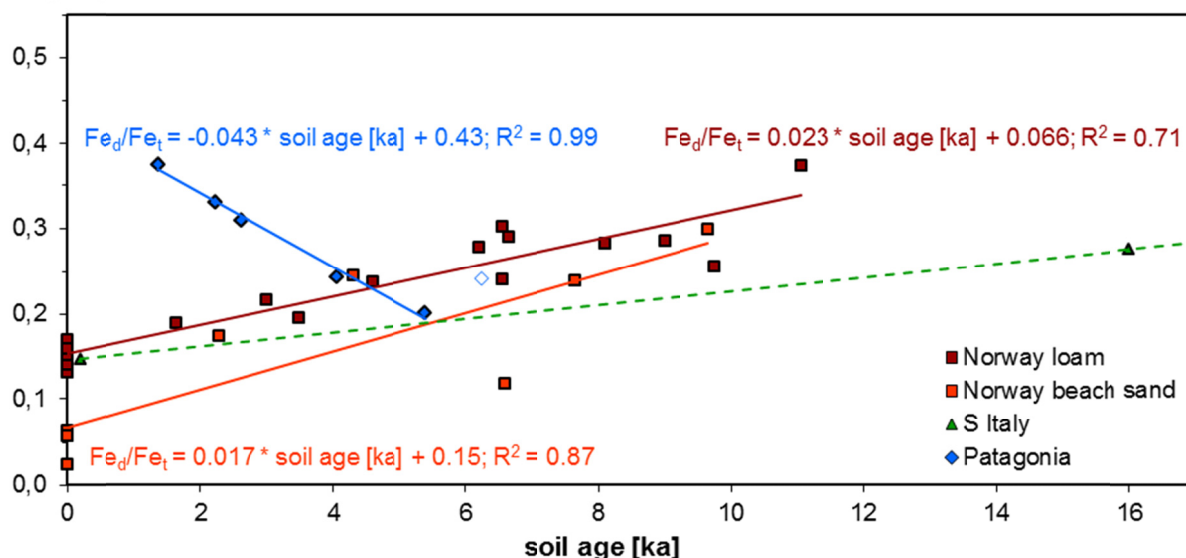


Figure 7. Trends of Fe_d/Fe_t in the studied soil chronosequences (weighted means of the upper 50 cm of each pedon). The two soil chronosequences on loamy sediments in Norway and on gravelly beach ridges in Patagonia could be merged with regard to Fe_d/Fe_t ratios.

Fe_d/Fe_t in the upper 50 cm

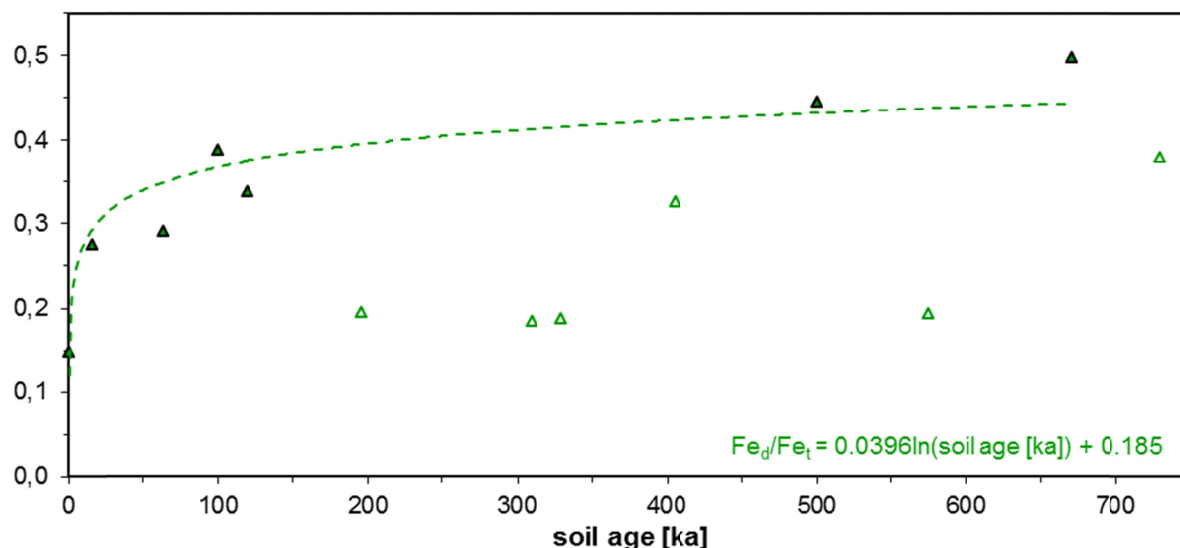


Figure 8. Fe_d/Fe_t ratios (weighted means of the upper 50 cm of each pedon) over the whole time-span of the Metaponto soil chronosequence. Open triangles represent Fe_d/Fe_t ratios of strongly eroded soils and were not included in the curve.

The soil chronosequences compared in this paper have all developed on marine, littoral and alluvial sediments. The parent materials of the pedons of the same chronosequence usually have the same source and thus similar mineralogical and geochemical composition, however, they vary with regard to texture. This variability is reflected in a large scatter of clay/silt ratios, particularly in the sequences on marine loamy sediments in Norway (Figure 9, red squares). The sandy beach deposits in Norway also show some scatter, but nevertheless a trend of increasing clay/silt ratios with time can be observed (Figure 9, orange squares). The Italian sequence exhibits a rough trend of increasing clay/silt ratios, but at the same time shows considerable scatter. The Patagonian soil chronosequences exhibit a distinct linear decrease in their clay/silt ratios (Figure 8; blue diamonds). This trend is caused by the high rate of dust influx and related silt accumulation, compared to the low rate of clay formation in the semi-arid climate of this area. This observation confirms that the proceeding dust accumulation is a main soil-forming process in this area, as has already been suggested by the decreasing Fe_d/Fe_t ratios of the Patagonian chronosequences.

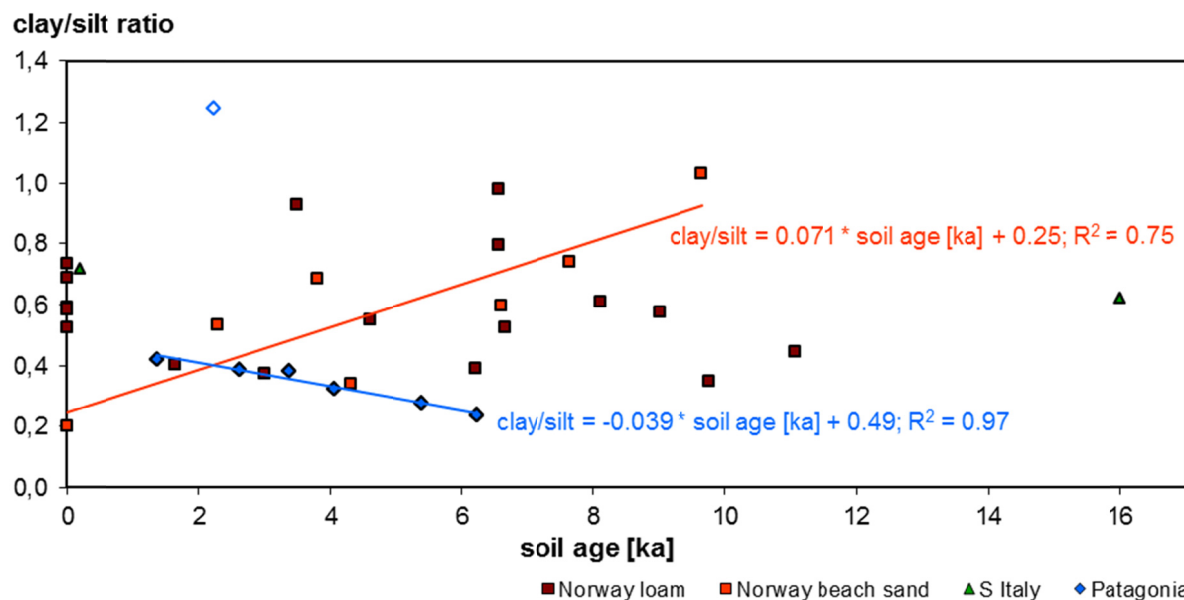


Figure 9. Clay/silt ratios, weighted mean values of the upper 100 cm (Norway and S Italy) or the upper 50 cm, respectively (Patagonia). Open blue diamond = outlier in Patagonian sequence.

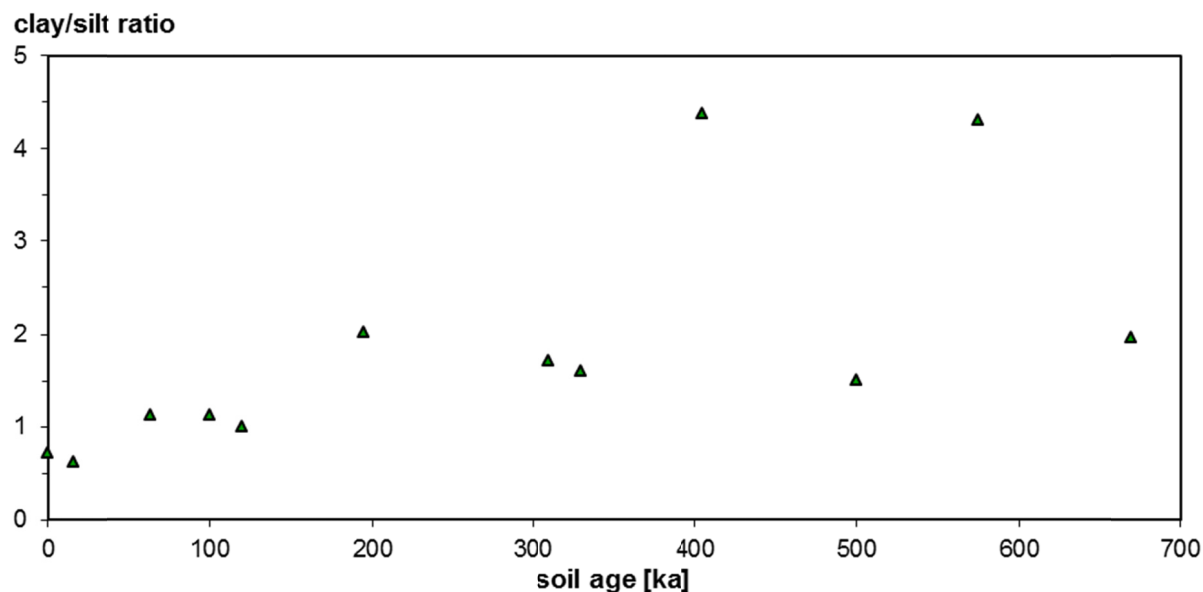


Figure 10. Clay/silt ratios (weighted mean values of the upper 100 cm) over the whole time-span of the Italian sequence.

2.5. Silicate weathering and base leaching

In general, one would expect that the main factors enhancing silicate weathering are:

- 1) soil moisture: moisture is required for important chemical weathering processes, e.g. for hydrolysis and for formation of acids as important weathering agents, and for leaching of weathering products from the weathering minerals from which they were released, which then allows for continuation of weathering.
- 2) soil temperature: temperature should increase weathering rates, since, according to the Arrhenius equation a temperature increase of 10 K induces doubling to triplication of the rates of any chemical reaction.

Several indices have been used in the literature to express the degree of silicate weathering and base leaching. Here, the molar ratio of $(Ca+Mg+K+Na)/Al$ is used. The weighted mean values of the $(Ca+Mg+K+Na)/Al$ molar ratios in the upper 50 cm of the soils of all chronosequences were calculated after subtracting the molar proportion of Ca from calcium carbonates.

The Norwegian soil chronosequences show large scatter of $(Ca+Mg+K+Na)/Al$ ratios (Figure 11; orange and red squares). The pedons on beach sand generally have higher ratios than those on marine loam, because of the higher quartz and lower silicate contents of the beach sand, involving lower Al contents. Despite the scatter, it is evident that the rates of $(Ca+Mg+K+Na)/Al$ decrease in the Norwegian sequences (Figure 11; slopes of the red and orange lines) are below the Holocene rate of $(Ca+Mg+K+Na)/Al$ decrease of the Italian chronosequence which is located in a drier moisture regime but warmer climate (Figure 11; slope of the green dashed line). The $(Ca+Mg+K+Na)/Al$ ratios in the Italian sequence show a clear decrease over 200 ka (Figure 12). Afterwards, decreasing rates indicate exhaustion of easily weatherable silicates. A minimal further decrease is assumed but cannot be proved due to large scatter.

The Patagonian soil chronosequences show distinct decreases in their $(Ca+Mg+K+Na)/Al$ ratios (Figure 11; slopes of the blue lines). It is, however, unlikely that these decreases are caused by strong silicate weathering and leaching in the semi-arid environment of these chronosequences. It is more probable that this decrease is related to the proceeding dust accumulation in these soils. This assumption is confirmed by the fact that the $(Ca+Mg+K+Na)/Al$ ratios of the Regosols, i.e. the soils with stronger dust accumulation, are overall on a lower level than those of the Leptosols. Probably, the minor amounts of sandy beach sediments in the interstitial voids between the pebbles of the beach

ridges have high quartz and low silicate contents, whereas the accumulated silty dust comprises comparatively high amounts of silicates, thus strongly increasing the Al contents and decreasing the (Ca+Mg+K+Na)/Al ratios with time.

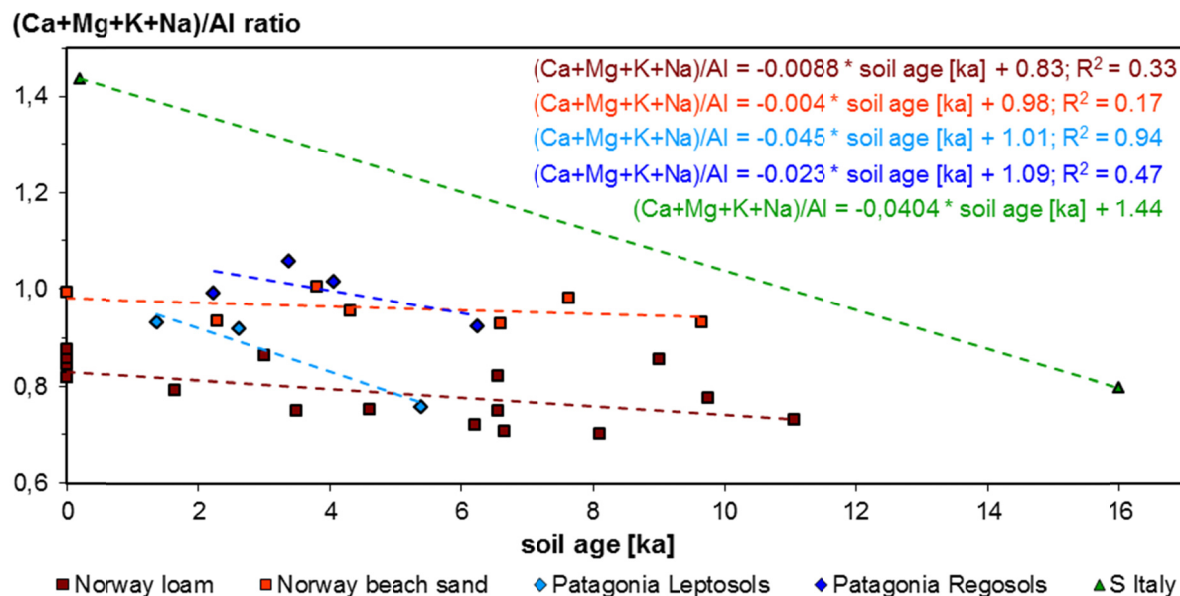


Figure 11. (Ca+Mg+K+Na)/Al molar ratio (weighted mean values of the upper 50 cm) as index for feldspar weathering and base leaching. Ca from calcium carbonate has been subtracted.

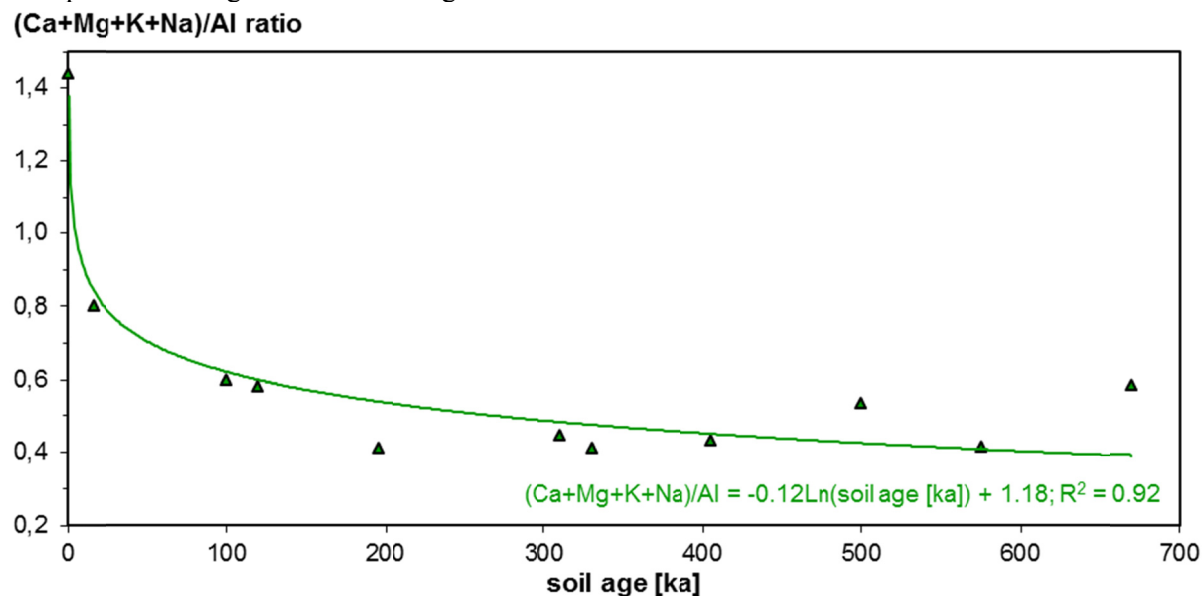


Figure 12. (Ca+Mg+K+Na)/Al molar ratio (weighted mean values of the upper 50 cm) of the Italian chronosequence over the whole time-span of the sequence.

3. Conclusions

The comparison of the six soil chronosequences shows that increase of soil thickness with time can be best described by logarithmic equations in all climates and lithologies included in this paper. Rates of soil thickness increase are influenced by both climate and parent material. Soil pH values in the upper

50 cm show decreasing trends in all climates and parent materials, whereby the general level of pH is higher in drier climates and in areas with substantial influx of calcareous dust than in humid climates. Formation of pedogenic iron oxides proceeds with time, usually leading to increasing Fe_d/Fe_t ratios, except for regions, where the influx of dust (having low Fe_d/Fe_t ratios) prevails over the process of pedogenic iron oxide formation, which is the case in the Patagonian chronosequences presented here. Dust influx has also a substantial influence on the time courses of clay/silt ratios and on element indices for silicate weathering. Variable dust influx and sediment inhomogeneity of pedons may lead to large scatter of data so that chronofunctions are difficult to obtain.

Five of the six soil chronosequences presented here are Holocene soil chronosequences, comprising time-spans of ca. 7 ka (Patagonia) to ca. 11 ka (Norway). These soils experienced some climatic shifts such as the mid-Holocene climatic optimum, and the Little Ice Age, but this climatic variation was within a range that appears to have no measureable effect on the soil chronofunctions.

In contrast, the Mediterranean soil chronosequence comprises ca. 730 ka, thus spanning a number of glacial-interglacial cycles. This means that the soil-forming factors climate and organisms were subject to major changes over the residence time of most of the soils comprised in the soil chronosequence. In the case of southern Italy, the environmental conditions alternated between Mediterranean conditions during interglacial periods and open *Artemisia* steppe to forest steppe environments during glacial periods [41, 42]. It is evident that rates of base leaching and formation of pedogenic oxides must have been significantly lower under steppe conditions than under Mediterranean conditions. It seems most likely that existing Mediterranean soils of several meters depth did not show any further deepening during periods of steppe environments. Instead, it can be assumed that enhanced dust influx and increased accumulation of soil organic matter took place in the uppermost part of the existing Mediterranean soils, and steppe soils such as Phaeozems and Chernozems formed. Each following shift to an interglacial period with Mediterranean conditions led to soil organic matter decomposition and hence degradation of the steppe soils. Mediterranean soil formation that had been interrupted during glacial periods continued, characterised by carbonate leaching, clay migration, and rubification. These conclusions are in agreement with Scarciglia et al. [13] who studied chronosequences of typical Mediterranean soils on Early to Late Pleistocene marine terraces along the northwestern coast of Calabria (southern Italy). A combination of chemical, mineralogical and micromorphological analyses revealed that soil-forming processes in those soils did apparently not proceed continuously but polycyclically. The authors thus concluded that the soils mainly developed during interglacial periods.

The problem of major changes in climate and vegetation that is discussed here, using the example of the Italian soil chronosequences, holds true for all soil chronosequences that extend back to pre-Holocene periods. Hotchkiss et al. [13], for example, discuss the impact of Pleistocene environmental changes on soil and ecosystem development in Hawaii. In such long soil chronosequences, the factor climate per se cannot be kept constant over time, as demanded in the original definition of soil chronosequences, specifying soil chronosequences as sequences of soils of different age that are supposed to have developed under similar conditions with regard to climate, vegetation and other living organisms, relief and parent material (see introduction). We thus need to accept that climate and vegetation varied over the time-spans comprised in these long soil chronosequences and that (at least) all soils of pre-Holocene age are polygenetic [29]. Hence, the requirements for these long soil chronosequences need to be modified as follows: soil chronosequences are sequences of soils of different age that are supposed to have developed under similar environmental conditions, or have experienced similar environmental changes through time, e.g. through glacial-interglacial cycles, with regard to the soil-forming factors climate, vegetation and other living organisms, relief and parent material.

In addition to the inevitable environmental variability in time discussed above, it is important to consider that, even under constant environmental conditions pedogenesis would not proceed uniformly over long time-spans. Instead, pedogenic thresholds and feedback systems naturally occur in the course of soil development. Examples for major pedogenic thresholds that occur in the soil

chronosequences comprised in this paper include the: threshold of acidification and availability of organic complexing agents that has to be reached before podzolisation starts in the soil chronosequence on beach sand in Norway; and the threshold of calcium leaching that has to be reached before clay illuviation starts in the soil chronosequences on loamy sediments in Norway and in southern Italy. Feedback systems are operating in the chronosequences involving clay illuviation (in Norway and southern Italy), since clay accumulation in the developing Bt horizons will lead to a progressively finer pore system in the Bt horizons, which in turn will favour further accumulation of clay. Feedback systems are also effective in the Norwegian soil chronosequence on beach sand, where the precipitation of the first metal-organic colloids that precipitate in the subsoil will attract other colloids and thus self-enhance the development of spodic horizons. Similarly, the first carbonate crystals, precipitating in the Patagonian soils will form nuclei, on which further carbonates will precipitate.

Besides progressive pedogenic processes such as silicate weathering and formation of clay minerals and pedogenic oxides, also regressive processes need to be taken into account. The main regressive process, particularly in the Mediterranean study area, is erosion. Johnson [44] explained soil thickness as a result of the interplay of: 1) deepening through weathering; 2) upbuilding (e.g. by dust influx); and 3) erosion. According to Johnson and Watson-Stegner [45], upbuilding can be regressive if unweathered fresh material accumulates at a rate that exceeds the weathering rate, or it can be progressive if weathering keeps pace with sedimentation and the sediments do not lead to profile rejuvenation or simplification. In the case of the Italian chronosequence, upbuilding by dust deposition represents a regressive process, because dust introduces fresh carbonates to previously decarbonated and clay illuviated soils. In the Patagonian chronosequences, upbuilding by dust deposition can be regarded as a regressive process with respect to Fe_d/Fe_t and clay/silt ratios because both ratios are decreased by dust accumulation, whereas they should generally increase with proceeding soil development including neoformation of clay minerals and pedogenic iron oxides. On the other hand, the dust that accumulated in the interstitial voids between the pebbles in the Patagonian beach ridges in many sites, represents the only fine earth at all. Dust accumulation in these cases is an essential process to increase the water holding capacity of the extremely gravelly soils and allow for the succession of higher plants. From this point of view, dust accumulation in this particular environment may be rather regarded as a progressive, than a regressive, process.

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