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Authors: Egli, Markus, Wernli, Michael, Kneisel, Christof, and Haeberli, Wilfried

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Melting Glaciers and Soil Development in the Proglacial Area Morteratsch (Swiss Alps): I. Soil Type Chronosequence

Markus Egli‡*
*Michael Wernli**
Christof Kneisel† and
*Wilfried Haeberli**

*Department of Geography, University of Zürich, Winterthurerstrasse 190, 8057 Zürich, Switzerland.
†Department of Physical Geography, University of Würzburg, Am Hubland, 97074 Würzburg, Germany.
‡megli@geo.unizh.ch

Abstract

Proglacial areas in the Alps usually cover a time span of deglaciation of about 150 years (time since the end of the "Little Ice Age" in the 1850s). In these proglacial areas soils have started to develop. In view of the foreseeable climate change, the time factor is of growing interest with respect to the landscape and consequently the soil development. We investigated soil changes (primarily on the basis of soil types) in the proglacial area Morteratsch (Swiss Alps) to derive time trends that can be used as a basis for spatial modeling. Differences in the soil development could be primarily interpreted in view of the time scale and topography (landscape shape, slope, aspect). Data was managed with GIS and regression analyses. Input data sets were the digital soil map, the glacial states, and the digital elevation model. The calculations were done raster based (GRID, 20 m resolution). After about 20 years the first signs of soil development could be found. Around 25% of the area of the valley floor is covered with weakly developed Skeletic/Lithic Leptosol after about 30 years of deglaciation. One hundred years of soil development led to a strong decrease of the Skeletic/Lithic Leptosol in favor of the Humi-Skeletic Leptosol and Ranker. Fluvisols and Cambisols play a subordinate role also after 100–150 years. Undisturbed and fast soil evolution was measured in flat positions and on slopes up to about 14°. In general, the various landforms also correlated well with soil evolution. One of the most surprising facts was that the weathering between south- and north-facing sites differed distinctly, with the north-facing sites having the higher weathering rates. Soil moisture seems to be a decisive factor in weathering. Thicker snow packs probably inhibit or reduce soil frost and allow larger fluxes of snowmelt water to infiltrate into already moist profiles. Slope, exposure and to a lesser extent also the landform determined the soil development: these influences could be quantified using regression analyses. These analyses serve as a basis for further spatio-temporal modeling.

Introduction

Mountainous ecosystems are likely to be especially sensitive to changing environmental conditions such as global warming, acid deposition, or nutrient cycling (Theurillat et al., 1998; Arn, 2002; Hosein et al., 2004). Although almost negligible on a worldwide scale, the Alps are an essential element of the landscape of Central Europe. Glaciers and discontinuous permafrost in such ecosystems react sensitively to atmospheric warming because the year-round temperature of their surroundings is not greatly below their melting point (Haeberli and Beniston, 1998; Maisch et al., 2003; Haeberli, 2004). The direct response of glaciers to climate change occurs through changes in the mass balance, and ultimately through variations in glacier length and size (Jóhannesson et al., 1989; Hoelzle et al., 2003; Haeberli et al., 2004). The Alpine landscape may respond very noticeably and differentially to climate change as it integrates all ecological and historical factors (Theurillat et al., 1998).

A key or "interface" function must be attributed to soils. Soil-landscape patterns result from the integration of short- and long-term pedogeomorphic processes (Friedrich, 1996; Klingl, 1996). Many of the soil properties change continuously with time.

Soil sequences may give an insight into the influence of a factor on the weathering rates. Jenny (1980) differentiated the following sequences: lithosequences (differences in parent mineralogy), climosequences (differences in precipitation and/or temperature), toposequences (lateral variations in slope and topography), chronosequences (effect of time on chemical weathering), and biosequences (variation in biota and its influence on chemical weathering). Precipitation and temperature, in particular, distinctly influence soil properties by affecting types and rates of chemical, biological, and physical processes (Dahlgren et al., 1997). Currently occurring worldwide climate changes are fuelling a growing interest in the effect that the factors of climate and time are having on the landscape and consequently soil evolution. Global warming due to anthropogenic emissions of greenhouse gases is predicted to increase the earth's average surface temperature during the next 50–100 years (IPCC, 2001).

Soils play a major role in the biogeochemical cycle including storage of nutrients and carbon. Carbon dioxide is converted to bicarbonate, and nutrients are released during carbonic acid weathering of silicate minerals, thus contributing to both carbon and nutrient cycling. Climate change can have significant impacts on the global biogeochemical cycle by altering the type and rate of soil processes and the resulting soil properties (Bain et al., 1994;

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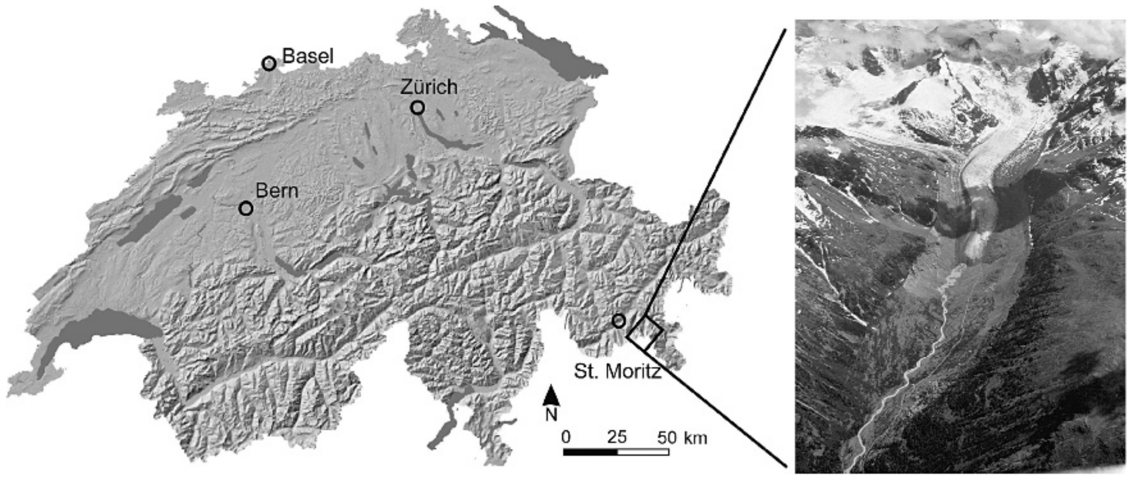


FIGURE 1. Location of and overview over the investigation site Morteratsch (Morteratsch glacier photographed by Christine Rothenbühler, 2004).

Dahlgren et al., 1997). Much of the work on climosequences has been summarized by Birkeland (1999). Earlier studies documented the effect of differences in climate along an altitudinal gradient—generally a decrease in temperature and an increase in precipitation—on plant communities and soil taxa. Common trends reported in these studies included changes in soil types, soil organic matter, clay content, soil acidity, and exchangeable ions (e.g. Whittaker et al., 1968; Mahaney, 1978; Laffan et al., 1989; Bäumler and Zech, 1994; Bockheim et al., 2000). Several studies (chrono- and climosequences) have been carried out in the past several years on soils developing in Alpine environments of northeastern Italy and southern Switzerland. These studies have elucidated the chemical and mineralogical processes leading to the formation of podzols (Mirabella and Sartori, 1998; Righi et al., 1999; Egli et al., 2001a, 2001b; Mirabella et al., 2002; Egli et al., 2003).

Time scales involved with the formation of the presently visible landscape reach as far back as the Alpine orogeny but mainly relate to the late glacial ice retreat at the end of the last Ice Age (20,000–10,000 yr BP) and embrace the entire Holocene time period. Soil chronosequences in the Alps (and also climosequences) cover periods up to about 15,000 years.

Soils in proglacial areas in the Alps are, however, young and were formed over a time span of about 150 years (the time since the "Little Ice Age" in the 1850s; Fitze, 1982). Proglacial areas are in most cases defined as the areas between present-day glaciers and the distinct moraines deposited in the 1850s.

As a consequence of warming, additional areas will become ice-free and subject to weathering and soil formation. The most evident soil changes in the Alps will occur in proglacial areas where already-existing young soils will continuously develop and new soils will form due to glacier retreat. The rate of reactions is of fundamental interest in the understanding of the soil system and its interaction with the surrounding environmental conditions. The aim of our research was to determine soil changes (primarily on the basis of soil types) in a proglacial area in order to derive time-trends that can be used for a further modeling. As a soil type reflects a specific soil evolution, soil types can be used to derive further properties if certain random conditions are considered (Egli et al., 2006 [this issue]).

Investigation Area

The soils studied lie within the proglacial area of Morteratsch in the Upper Engadine (Switzerland). The border of the proglacial area is given by distinct moraines deposited in the 1850s during the "Little Ice Age" (Figs. 1 and 2). The actual length of this proglacial area is approximately 2 km, and it has an area of 1.8 km^2. The proglacial area is in a valley that runs N–S. The altitude ranges from 1900 m a.s.l. to about 2150 m a.s.l. Alpine glaciers have fluctuated during the past 12,000 years near the borders of the moraines formed in the year 1850 indicating more

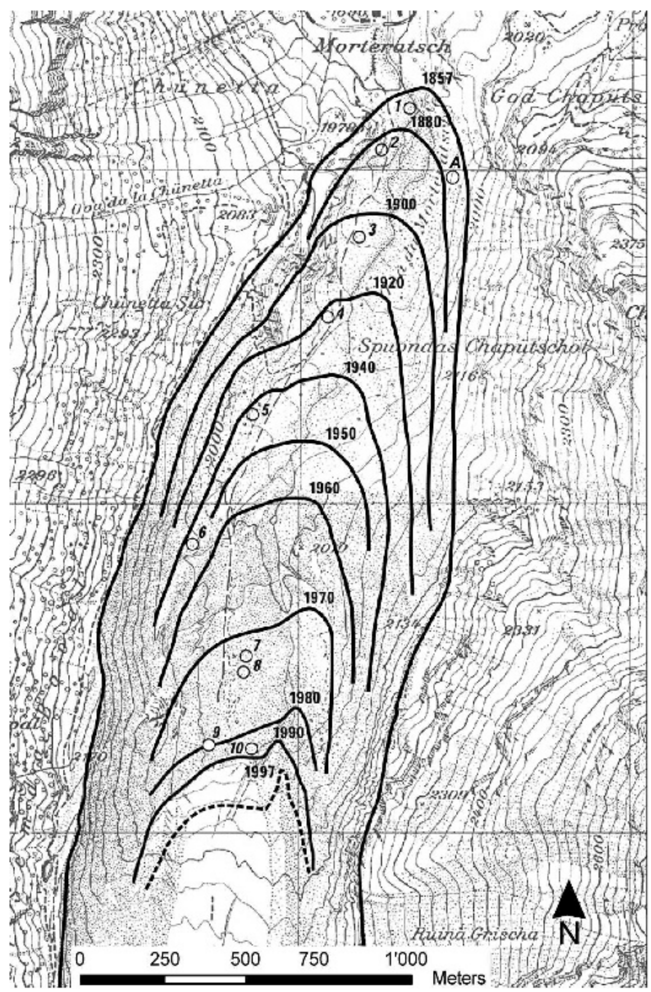


FIGURE 2. The proglacial area of Morteratsch with isochrones of deglaciation and monitoring sites (after Burga, 1999).

TABLE 1

Geochemical and mineralogical characteristics of the parent material (Egli et al., 2003).

Compounds	Mean chemical composition (SD)	Minerals	Mean composition (wt.%)
SiO_2 (g kg^{-1})	734.7 (29.4)	Albite	25–28
CaO (g kg^{-1})	7.6 (9.3)	Orthoclase	25–29
MgO (g kg^{-1})	6.3 (6.1)	Anorthite	3–6
Na_2O (g kg^{-1})	27.7 (5.8)	Muscovite	1–4
K_2O (g kg^{-1})	45.1 (8.9)	Biotite–Illite	3–7
MnO (g kg^{-1})	0.4 (0.4)	Chlorite	1–2
Fe_2O_3 (g kg^{-1})	29.1 (10.6)	Quartz	30–35
Al_2O_3 (g kg^{-1})	133.9 (5.9)	Amphiboles	1–3
TiO_2 (g kg^{-1})	6.5 (4.2)		

SD = standard deviation.

or less similar climatic (±1–1.5°C) and hydrologic conditions within that period. This has been shown by many geomorphologic and climatic studies (cf. Burga and Perret, 1998; Keller, 1994; Patzelt, 1977; Magny, 1992; Renner, 1982; Gamper, 1985; Maisch, 1992).

Generally, three plant species groups were distinguished (Burga, 1999): (1) pioneer species, starting early in the chronosequence and reaching their optimum in early or medium stages; (2) a selection of subalpine forest and dwarf-shrub/heat species, most starting in later stages; and (3) a selection of species occurring mainly in alpine grassland with different distributions and optima along the gradient. The glacial till consists of granite and gneissic material (Table 1). The morainic material was produced through glacial transport within a small area of relatively homogeneous parent material. The lithostratigraphic units are mainly the Bernina- and Stretta-crystalline (Spillmann, 1993; Büchi, 1994). The parent material was altered and reached the state of the greenschist facies (Trommsdorff and Dietrich, 1999). Present climatic conditions for the Morteratsch site are approximately 0.5°C mean annual temperature and approximately 1000–1300 mm mean annual precipitation as calculated by using data from the nearby Samedan and Bernina meteorological stations. According to the WRB (FAO, 1998) soil classification, the soils vary from Lithic Leptosols to Dystric Cambisols.

Materials and Methods

SOIL CARTOGRAPHY

Soil mapping in the proglacial area was made by means of aerial photographs and field investigations. The soil map had a scale of 1:10,000. Soil cartography and classification was performed according to the FAL system (Brunner et al., 1997). Soil mapping already includes a certain generalization as small-scale variations (less than approximately 100 m^2) could not be considered. The items included the soil type, soil depth relevant for plant growth (= soil volume – skeleton volume – groundwater volume; result is related to depth instead of volume), parent material, vegetation, topography, soil hydrology, terrain form, pH-value, organic C content, soil skeleton, granulometry, aggregates, and humus form. In total, 47 soil pits and core drillings were made and described in more detail and used as reference profiles for mapping. According to the occurrence of the soil units, the number of profiles varied (1 for Gleyic Cambisol, 3 for Dystric Cambisol, 18 for Humi-Skeletic Leptosol (including Ranker), 3 for Fluvisol, and 22 for Skeletic/Lithic Leptosol). Ten pits (that are related to a botanical monitoring; Burga, 1999) were

TABLE 2

Classification of the slope.

Class	Slope (°)
1	0–3
2	3–6
3	6–9
4	9–14
5	14–19
6	19–27
7	27–37
8	37–42
9	>42

excavated for the chemical and physical analysis of the soil material. Around 2–4 kg of soil material were collected per soil horizon. In order to yield reasonable results, large soil sampling volumes are needed for soils in alpine areas (Hitz et al., 2002). Soil bulk density was determined by a soil core sampler (or by excavated holes with a volume of about 500–2000 mL that were backfilled with a measurable volume of quartz sand). Taking advantage of the profile pits, undisturbed soil samples were taken down to the C horizon.

SOIL CHEMISTRY AND PHYSICS

Element pools in the parent material (Ca, Mg, K, Na, Fe, Al, Mn, Si, and Ti) were determined by a method of total dissolution (using a mixture of HF, HCl, HNO_3, and H_3BO_3) in a previous investigation (Egli et al., 2003). Total C and N contents of the soil and parent material were measured with a C/H/N analyzer (Elementar Vario EL). Soil pH (in 0.01 M $CaCl_2$) was determined on air-dried fine earth samples using a soil:solution ratio of 1:2.5.

The particle size distribution was determined on some selected soil samples. After a pretreatment of the samples with H_2O_2 (3%), particle size distribution of the soils was measured by a combined method consisting of sieving the coarser particles (2000–32 μm) and the measurement of the finer particles (<32 μm) by means of an X-ray sedimentometer (SediGraph 5100).

AREA STATISTICS

Area calculations (proportion of different soil types between two isochrones and in relation to either time or topographic features) and statistics (regression analyses) were performed using ArcGIS 8.3 (ESRI) with modules programmed in Visual Basic for Applications (VBA). Input data sets were the digital soil map, the glacial states (Burga, 1999), and the digital elevation model (raster of 20 m) within the proglacial area. The calculations were done raster-based (GRID, 20 m resolution).

Soil types were not only related to the state factor time but also to the landscape forms, slope, and north and south exposure (see Tables 2 and 3). North exposure is related to >270–90°N and south exposure to >90–270°N. Relative area calculations refer to the area between two isochrones of deglaciation (Fig. 2). In total, 11 different glacial states (with corresponding isochrones) could be distinguished (Burga, 1999).

Results

SOIL CHARACTERISTICS

According to the FAL classification (translated into the FAO–UNESCO [1990] and WRB system [FAO, 1998]), primarily

TABLE 3

Description of landscape forms derived from the digital elevation model.

Landscape form	Landscape code	Landscape form: perpendicular to slope (planform curvature)	Landscape form: direction of slope (profile curvature)
Depressions	10	concave	concave
Foot of the slope, flattening slope	20	planar	concave
Flattening slope ridge	30	convex	concave
Valley shape	40	concave	planar
Flat slope	50	planar	planar
Steepening slope ridge	60	convex	planar
Steepening valley	70	concave	convex
Steepening slope	80	planar	convex
Ridges	90	convex	convex

the following types of sites can be differentiated: Endoskeletic Fluvisols, Skeletic or Lithic Leptosols, Humi-Skeletic Leptosols (including some sites with Ranker (FAO–UNESCO) that have a weak B horizon, and Dystric and Gleyic Cambisols (endoskeletic) and sites having no soil. The young soils that are closer to the glacier show almost no obvious signs of chemical weathering and alteration products. They are usually characterized by a very thin and often discontinuous humus layer (Table 4; Fig. 3). The oldest soils (150 years), however, have a spatially continuous humus layer (O or A horizon) and partially signs of weathering product formation (formation of Fe- and Al-oxyhydroxides; start of clay mineral formation/transformation) and, thus, a weakly pronounced B horizon. This could be shown by a higher chroma (but the same hue and value) with a Munsell soil color chart when compared to the A or C horizon (10YR7/4 in the AB horizon; Egli et al., 2003). The oxalate extractable Fe (442–1115 mg kg^{-1}) and

TABLE 4

Selected chemical and physical properties of 11 typical soil profiles in the proglacial area.

Site/Soil	Soil age (y)	Horizons	Depth (cm)	Bulk density (g cm^{-3})	Skeleton (wt.–%)	Sand* (g kg^{-1})	Silt* (g kg^{-1})	Clay* (g kg^{-1})	pH ($CaCl_2$)	Org. C (g kg^{-1})	Org. N (g kg^{-1})
A/Humi–Skeletic Leptosol (Ranker)											
	140	A	0–7	1.4	65	77	19	4	4.7	34.3	2.5
		(Bw)A	7–25	1.9	59	78	19	3	5.2	3.7	0.3
		C	>25	1.9	64	78	19	3	5.5	2.5	0.2
1/(Endoskeletic) Fluvisol											
	140	(C)A1	0–2	1.0	11	55	36	9	4.8	16.4	0.6
		(C)A1	2–8	1.0	11	55	36	9	4.8	16.4	0.6
		2C	8–14	1.3	19	49	42	9	4.9	8.4	n.d.
		3C	14–23	n.m.	70	n.m.	n.m.	n.m.	n.m.	n.m.	n.d.
2/Humi–Skeletic Leptosol											
	120	A	0–3	n.m.	40	n.m.	n.m.	n.m.	5.5	176.1	7.6
		C	3–13	n.m.	65	n.m.	n.m.	n.m.	4.9	16.7	1.4
3/Skeletic Leptosol											
	100	A	0–2	n.m.	40	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.
		C	2–26	1.5	70	n.m.	n.m.	n.m.	5	1.7	n.d.
4/Skeletic Leptosol											
	80	A	0–2	n.m.	15	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.
		AC	2–6	1.5	70	85	12	3	4.5	5.2	n.d.
		(B)C	6–8	1.5	75	85	12	3	4.5	5.2	n.d.
		C	8–14	1.4	75	84	14	2	4.8	1.7	n.d.
5/Skeletic Leptosol											
	70	A	0–3	n.m.	40	n.m.	n.m.	n.m.	n.m.	n.m.	n.m.
		C	3–22	1.3	68	n.m.	n.m.	n.m.	4.5	4.4	n.d.
6/Skeletic Leptosol											
	60	CA	0–12	1.5	55	85	13	2	5.1	6	n.d.
7/Skeletic Leptosol											
	30	AC	0–6	1.5	53	92	6	2	5.2	1.9	n.d.
		C	6–25	1.4	70	83	14	3	5.4	0.7	n.d.
8/Skeletic Leptosol											
	30	(A)C	0–15	1.6	66	82	17	1	5.7	2	n.d.
9/Skeletic Leptosol											
	20	(A)C	0–18	1.8	77	n.m.	n.m.	n.m.	5.7	2	n.d.
10/Lithic Leptosol											
	10	C	0–9	1.7	75	75	22	3	5.8	0.9	n.d.

n.d. = not detectable.

n.m. = not measured.

* size fractions: sand (2000–63 μm), silt (<63–2 μm), clay (<2 μm).

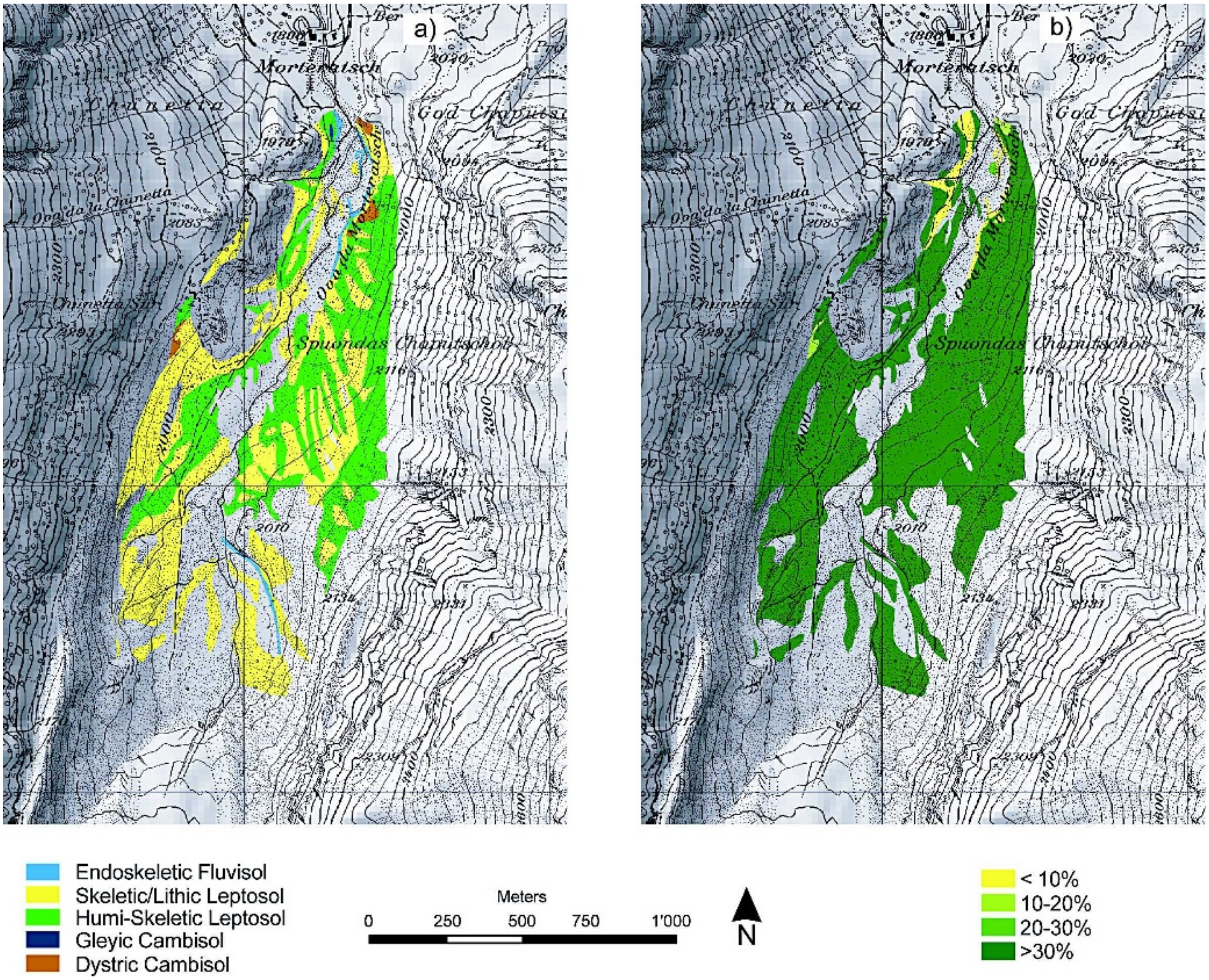


FIGURE 3. Soil types (a) and skeleton content (in vol.-%) of the parent material (b) in the proglacial area. DEM25 reproduced by permission of swisstopo (BA067583).

Al (122–287 mg kg^{-1}) in the A and B horizon showed higher values than in the C horizon (Fe: 427 mg kg^{-1}, and Al: 115 mg kg^{-1}; Egli et al., 2003).

The fine earth of the soils is very sandy (sand content 50–90%) and contains partially some silt (eolian attribution? or grinding due to fluvial transport). The Fluvisols had partially a slightly increased silt content. In all other soils, granulometry was rather uniform (Fig. 3, Table 4). On specific sites, Dystric Cambisols (with a clearly differentiated A-Bw-C profile sequence and endoskeletic characteristics) can be found that are not a further step in the observable soil evolution in the proglacial area but clear evidence of the impact of geomorphodynamics on soil development in this Alpine environment. Close to the lateral moraines, there are a few sites where debris flows penetrated the moraine and deposited pre-weathered material of older soils from outside the proglacial area.

Finally, the vegetation also reflects the soil status. The first flowering plants invading young deglaciated surfaces are scattered individuals of mostly sterile *Epilobium fleischeri* and *Linaria alpina* that appear after about 7 years. *Epilobietum fleischeri* obtain a higher cover-abundance after ca. 27 years. First plants of *Oxyrietum digynae* appear after ca. 12 years and disappear after ca. 27 years. The establishment of *Larici–Pinetum cembrae* takes place after about 77 years (Burga, 1999) on sites where the soil has been more intensely developed.

According to the soil map, a great part of the very young soils have a thickness of <10 cm (after the subtraction of soil skeleton) and a skeleton content of >50% (weight). Soil skeleton (defined as the fraction >2 mm) was constantly high in all soils. The size of the soil skeleton, however, varied considerably, giving rise on especially very young surfaces to a patterned soil distribution and development. Older soils (on sites with a tendency of accumulation) had a thickness of up to 40 cm and a skeleton content of <50%. The humus content usually varied between <2% and 10%. The soils are generally weakly to strongly acidic, which depends also on their age. Generally, the closer the site is located to the present-day glacier margin, the higher is the pH value.

REGRESSION ANALYSES

On a small scale (less than approximately 100 m^2), the parent material varies due to changing physical properties (e.g. size of the soil skeleton) and minor lithological variations. This results in fine-scale soil heterogeneity and variations of soil development. The patterned distribution of the soils (especially at its earliest stage) shows that the soil forming conditions were not identical

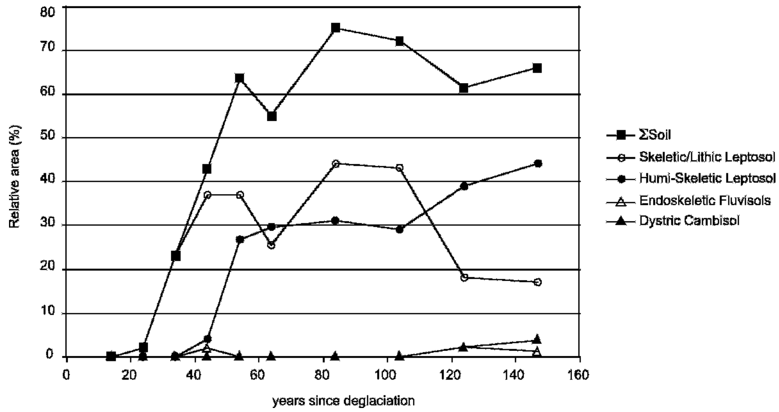


FIGURE 4. Relative area of soil types in the proglacial area as a function of time.

everywhere. On a larger scale, the parent material in the proglacial area can, however, be considered more or less constant (see Tables 1 and 4, Fig. 3) and thus a negligible factor according to Jenny (1980) if forming conditions were to be taken into account in the explanation of the different state of a specific soil.

Climate, furthermore, does not vary greatly in the area of interest and can therefore also be considered as a negligible factor. The state factor vegetation is in this case not a really independent one. The quality of the substratum such as the grain size of the parent material, the distribution of moisture, and the availability of nutrients are essential factors for the establishment of the vegetation (Burga, 1999). In a following step, plants influence also further soil development. According to Jenny's (1980) paradigm, only the state factors time and topography remain and consequently determine differences in soil evolution. This means that changes in the soil can be primarily interpreted in view of the time scale and topography and we tried, therefore, to model soil development in the proglacial area of Morteratsch using only these two state factors (see also Egli et al., 2006 [this issue]). Topography pertains to the configurations of the landscape and may refer to inclination, length of slope, concavity or convexity, and (north and south) exposure (Jenny, 1980). The time factor finally needs a dynamic simulation. Dynamic simulation of soils is based on the assumption that the state of each system at any moment can be quantified, and that changes in the state can be described by rate or differential equations (Hoosbeek et al., 2000). The derivation of rates in changes was obtained in this work indirectly by regression analyses (and not by process models).

The relative distribution of the different soil types as a function of time is shown in Figure 4. As expected, the weakly developed soils (Leptosols) dominate. The Skeletic/Lithic Leptosols develop earlier than the Humi-Skeletic Leptosol. First signs of soil development can be found after about 20 years. Around 25% of the area of the valley floor is covered with weakly developed Skeletic/Lithic Leptosol after about 30 years of deglaciation. One hundred years of soil development led to a strong decrease of Skeletic/Lithic Leptosols in favor of Humi-Skeletic Leptosols and Rankers (not evidenced in Fig. 4 because of its marginal occurrence). Fluvisols and Cambisols after 100–150 years also play a subordinate role. The sum of the soil area increases steadily in the first 70 years and then reaches a kind of asymptotic value that is, however, due to a partial hindrance of further soil development caused by a rock outcrop on the orographically left and northern part of the proglacial area (Fig. 3a).

The form of the landscape influences soil formation. On sites with a tendency to accumulation, such as depressions and at the foot of slopes as well as on concave slopes or flat slopes, a very significant correlation of the soil types with time and furthermore a distinct soil sequence with time from less to better evolved soils can be measured (Fig. 5). Sites receiving deposition are, however, not ideal for quantifying time-dependent processes under "undisturbed" situations. Under such circumstances, soil development is, in its initial stages, slightly enhanced. In most cases the correlations between landform and soil type are similarly significant. A higher variability and less pronounced correlation can be seen regarding the landforms "valley shape" and "steepening valley." A similar behavior can be detected if the soils are correlated to the slope classes. Generally Skeletic/Lithic Leptosols develop first and tend to be replaced by the Humi-Skeletic Leptosols after a certain time. The fastest soil development can be measured—as expected—in flat or only moderately steep slopes. The faster soil development in flat slopes and in depressions might, theoretically, also be due to finer rock material deposited in such areas. The steeper the slope the later is the transition from Skeletic/Lithic Leptosol to Humi-Skeletic Leptosol (Fig. 6). For very steep sites, however, only a small number of relative areas were available.

Very distinct differences between north- and south-facing sites can be differentiated (Fig. 7). The development of Skeletic/Lithic Leptosols begins earlier on the north-facing than on the south-facing sites. At about 80 years, a greater percentage of the area is covered with this soil on south-facing sites . Skeletic/Lithic Leptosols transform much earlier into Humi-Skeletic Leptosols on north-facing sites where a continuous and quite distinct increase in the relative area of this soil type can be already found after 30 years. On south-facing sites, Humi-Skeletic Leptosols cover an area of about 20% after 150 years of soil development (60% on north-facing sites).

Discussion

Soil development proceeds quickly in the proglacial area. One hundred fifty years of soil development lead to Ranker (Humi-Skeletic Leptosols) that will be transformed later into Dystric Cambisol. Our soil cartography revealed that within 150 years significant soil-forming processes took place such as accumulation of organic matter, the beginning of parent material alteration, and the formation of weathering products (weak B-horizons in Ranker). Glaciers and periods of glaciation may have a significant impact on weathering, changing the interplay between physical and chemical weathering processes, by putting large volumes of dilute meltwaters and fine-grained sediment in contact with each other.

There exist various indications in literature about the rate of soil formation and weathering in cold environments. Especially on young surfaces, Arn (2002), Egli et al. (2003), and Hosein et al.

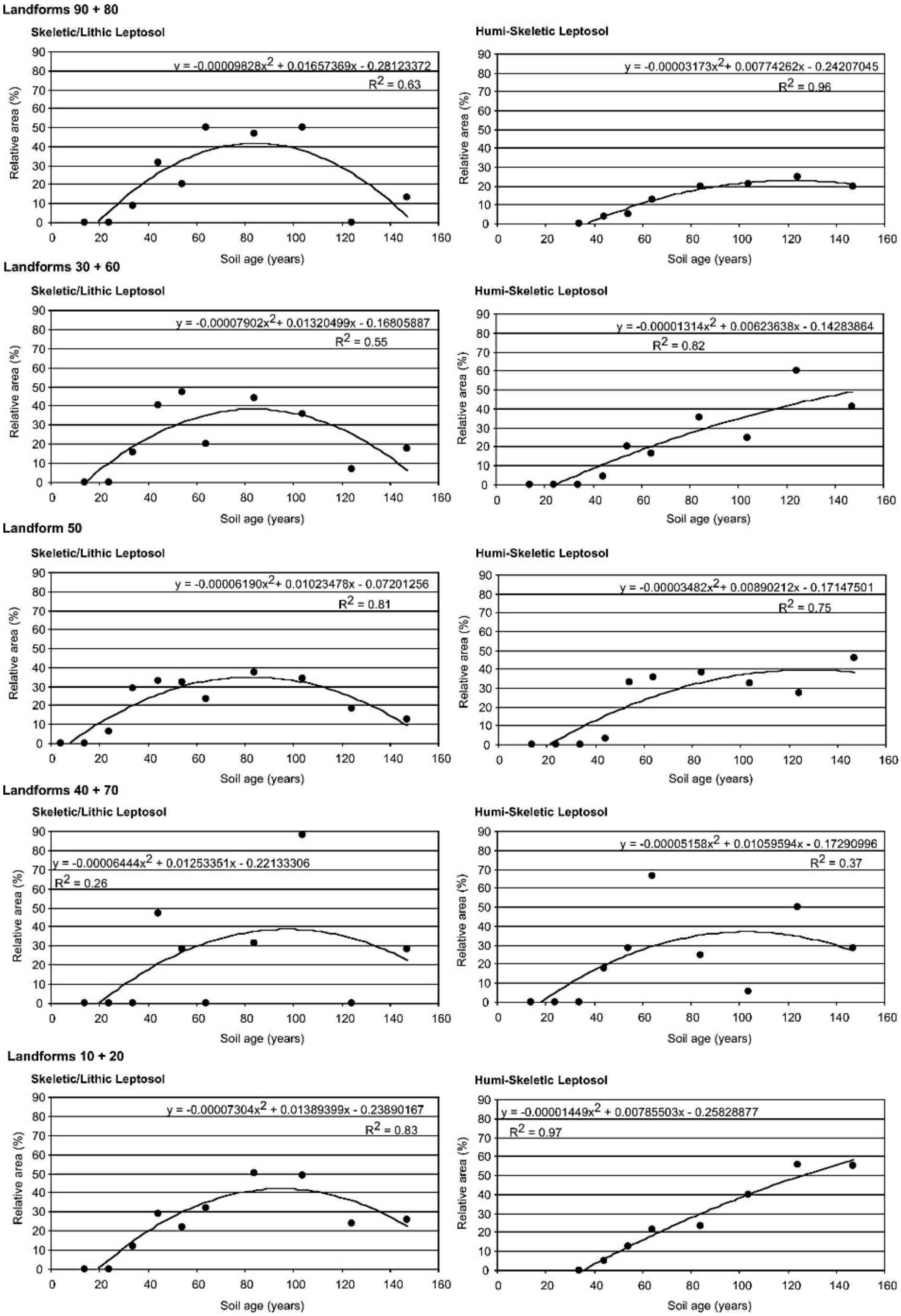
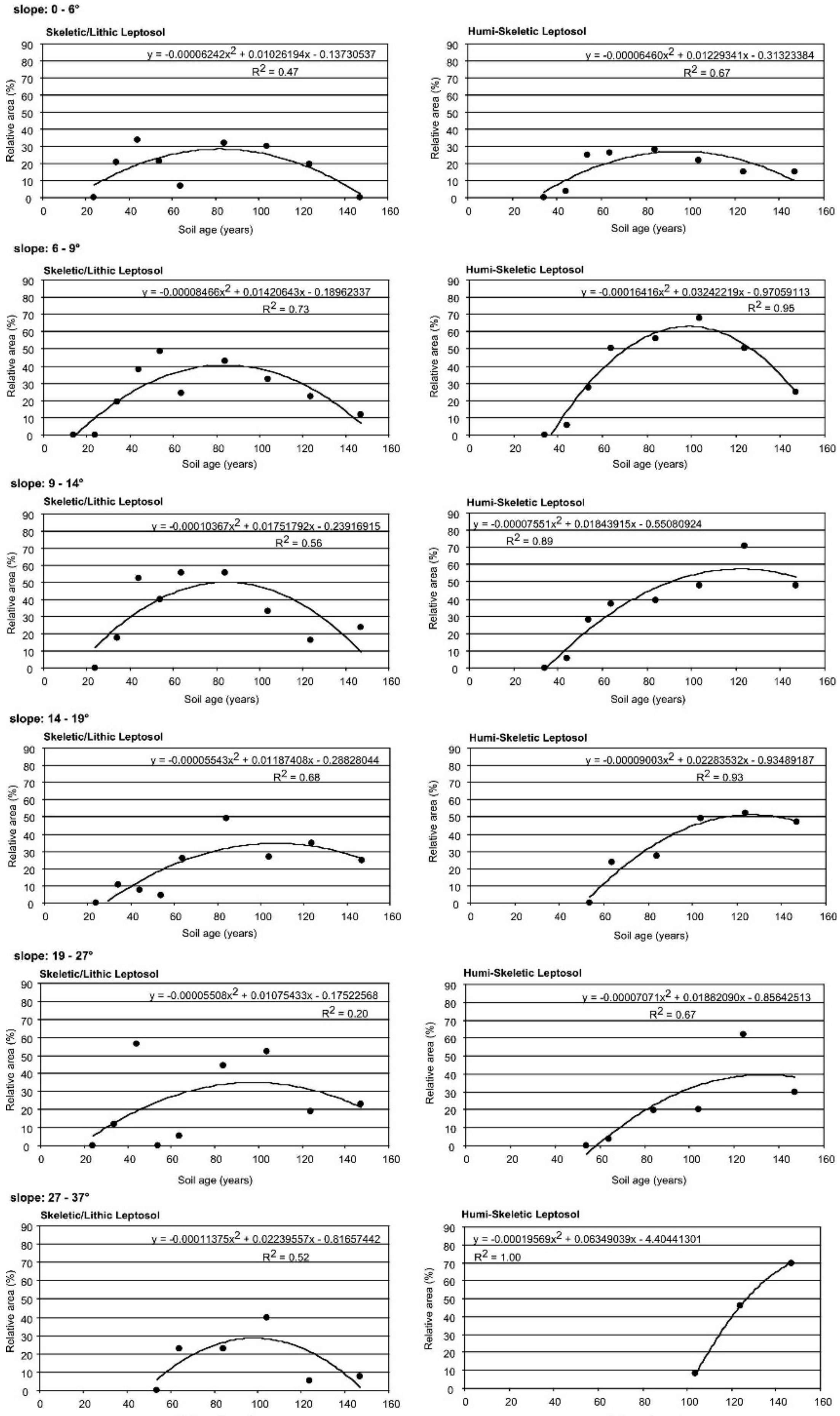


FIGURE 5. Relative area of the Skeletic/Lithic Leptosols and Humi-Skeletic Leptosols as a function of landscape form and time span of deglaciation with 10 = depressions, 20 = foot of the slope, 30 = flattening slope ridge, 40 = valley shape, 50 = flat slope, 60 = ridge slope, 70 = steepening valley, 80 = steepening slope, 90 = ridges.


slope: 0 - 6°
Skeletic/Lithic Leptosol
y = -0.00006242x2 + 0.01026194x - 0.13730537
R2 = 0.47
Humi-Skeletic Leptosol
y = -0.00006460x2 + 0.01229341x - 0.31323384
R2 = 0.67
slope: 6 - 9°
Skeletic/Lithic Leptosol
y = -0.00008466x2 + 0.01420643x - 0.18962337
R2 = 0.73
Humi-Skeletic Leptosol
y = -0.00016416x2 + 0.03242219x - 0.97059113
R2 = 0.95
slope: 9 - 14°
Skeletic/Lithic Leptosol
y = -0.00010367x2 + 0.01751792x - 0.23916915
R2 = 0.56
Humi-Skeletic Leptosol
y = -0.00007551x2 + 0.01843915x - 0.55080924
R2 = 0.89
slope: 14 - 19°
Skeletic/Lithic Leptosol
y = -0.00005543x2 + 0.01187408x - 0.28828044
R2 = 0.68
Humi-Skeletic Leptosol
y = -0.00009003x2 + 0.02283532x - 0.93489187
R2 = 0.93
slope: 19 - 27°
Skeletic/Lithic Leptosol
y = -0.00005508x2 + 0.01075433x - 0.17522568
R2 = 0.20
Humi-Skeletic Leptosol
y = -0.00007071x2 + 0.01882090x - 0.85642513
R2 = 0.67
slope: 27 - 37°
Skeletic/Lithic Leptosol
y = -0.00011375x2 + 0.02239557x - 0.81657442
R2 = 0.52
Humi-Skeletic Leptosol
y = -0.00019569x2 + 0.06349039x - 4.40441301
R2 = 1.00
Relative area (%)
Soil age (years)

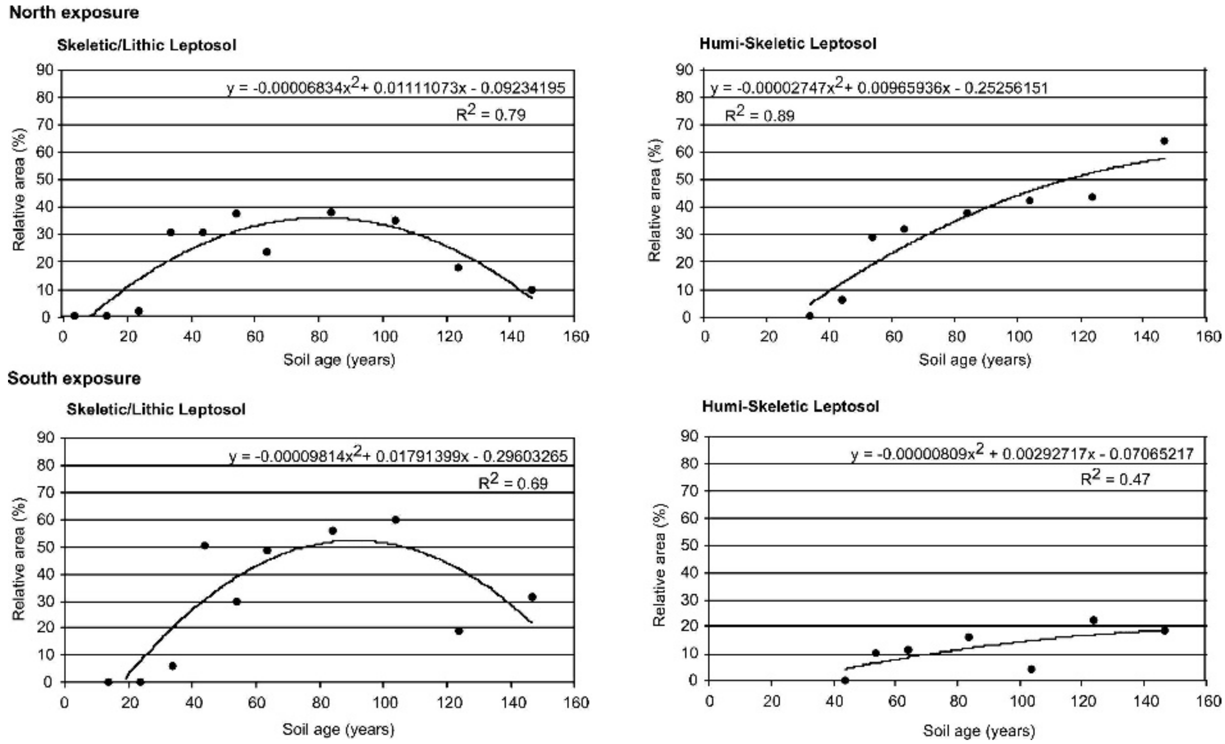


FIGURE 7. Relative area of the Skeletic/Lithic Leptosols and Humi-Skeletic Leptosols as a function of exposure and time span of deglaciation.

(2004) measured high formation or transformation rates of minerals as well as chemical denudation rates (Anderson et al., 2000). Egli et al. (2001a, 2001b) found Humi-Skeletic Leptosols after 150 years of soil formation and Dystric Cambisols after about 350 years of soil formation in a proglacial area at a similar altitude in the Alps. Fitze (1982) and Patzelt (1973) described Rankers that have been formed in the time span of 200–350 years after deglaciation. Zech and Wilke (1977) found already after 200 years of soil evolution a Dystric Cambisol and after 600 years a Podzol in a proglacial area consisting of granitic, glacial deposits and having a mean annual precipitation of 1800 mm and mean annual temperature of 0°C. Haugland (2004), furthermore, described in Norway (Jotunheimen region; characterized by a slightly colder climate when compared to Morteratsch) a Ranker after 47 years of deglaciation and a beginning brunification on sites with about 120 years of soil evolution. The rate of soil evolution is, in Alpine areas, distinctly determined by the parent material and the climate (especially the amount of precipitation). A very fast soil evolution was found by Alexander and Burt (1996) in southeast Alaska where already after about 240 years an E horizon and thus a podzolization could be seen on moraines. Soil evolution was, however, enhanced there by very high precipitation (>250 cm yr^{-1}) and higher temperatures.

The rate of soil formation in the Morteratsch area is in the same order of magnitude that was found in studies in Norway (Haugland, 2004). Dystric Cambisols usually form after about 250–300 years of deglaciation (Egli et al., 2001a) in Central Alpine areas. Podzols require a minimum duration of soil formation of about 1200 years (Egli et al., 2003).

Except for the landforms steepening valley and valley shape, the various landforms correlate well, in general, with soil evolution. The individual regression curves, however, differ only slightly, which means that differences in the evolutionary sequence are existent but small. Undisturbed and fast soil evolution can be expected in flat positions and on slopes of up to about 14° (Fig. 6).

The differences in weathering between sites with northern and southern exposure were surprisingly distinct (with higher weathering rates on sites with northern exposure; see Fig. 7). North exposure includes in this context NW-, N-, and NE-facing sites, and south exposure SE-, S-, and SW-facing sites. Physical weathering in Alpine areas is thought to be more intense on south-exposing sites or rock walls due to repeated influence of freeze-thaw cycles (e.g. Gruber et al., 2004). Several studies show the influence of slope aspect and the resulting microclimate on soil weathering and development (Cooper, 1960; Klemmedson, 1964; Macyk et al., 1978; Carter and Ciolkosz, 1991; Rech et al., 2001). Higher temperatures on south-facing slopes should theoretically increase rates of chemical weathering (Rech et al., 2001). Other factors that influence weathering include the number of freeze-thaw cycles and the availability of moisture (Rech et al., 2001). Several studies (e.g. Cooper, 1960) found more advanced stages of soil development on south-facing slopes. In contrast, other authors (e.g. Macyk et al., 1978; Carter and Ciolkosz, 1991) measured thicker solums and higher clay concentrations on north-facing slopes. In general, higher mean annual air and soil temperatures can be expected on south exposure, while a higher moisture content is usually associated with north-facing slopes.

←

FIGURE 6. Relative area of the Skeletic/Lithic Leptosols and Humi-Skeletic Leptosols as a function of slope classes and time span of deglaciation.

The higher moisture content most probably leads to a more intense leaching of ions and to an enhanced transformation of primary into secondary minerals (cf. Hunckler and Schaetzl, 1997). Thick snow packs inhibit or reduce soil frost and allow large fluxes of snow meltwater to infiltrate into already moist profiles (Hart and Lull, 1963; Sartz, 1973; Isard and Schaetzl, 1995; Schaetzl and Isard, 1996).

Conclusions

Although the soil distribution has partially a patterned character, the statistical and geographical analysis of a soil map in a proglacial area has shown several significant trends in soil types that can be used for a further modeling of their evolution. Using the statistical analyses a modeling of soil dynamics in the proglacial area can be made possible. Several stages of soil development could be distinguished. After about 20 years of deglaciation, Skeletic/Lithic Leptosols begin to develop. Humi-Skeletic Leptosols (including Ranker) start to replace them after about 100 years. Dystric Cambisols will start to develop only after about 250–300 years of deglaciation. The WRB classification system (FAO, 1998) did not always fully match the required soil description (e.g. Ranker).

Slope, exposure, and to a lesser extent also landform determine soil development. The influence of individual parameters (and according classes) could be described with regression analyses. Despite the generally cold conditions in the proglacial area, the climate is favorable enough for rapid soil development with organic matter accumulation, soil acidification, etc. (see Burt and Alexander, 1996). Soil distribution is, however, also patterned. Patterned structures may be associated with abrupt thresholds that either enhance or stop/hinder soil formation. This might be due to several causes such as microclimatic conditions, microrelief, deposition of physically inhomogeneous parent material (sites with a more fine-grained deposit close to rock debris) and probably also to brief periglacial activity (cf. Haughland, 2004). The regression analyses and area statistics were raster based (20 m) and therefore took such microvariations into consideration only to a limited degree.

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