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Initial soil development and carbon accumulation on moraines of the rapidly retreating Werenskiold Glacier, SW Spitsbergen, Svalbard archipelago

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ABSTRACT

The rapid retreat of the Werenskiold Glacier (the Svalbard archipelago, High Arctic) is leading to an extensive broadening of the proglacial zone covered with recent moraine till on older glacigenic deposits or directly on bedrock schists. To study the type and intensity of initial soil development under a harsh periglacial climate, a chronosequence of six soils was established between the glacier front and its terminal moraine on 1 to about 80 year-old moraines. Although the surface layer over the entire area of study is frost-active, the patterned features are not well developed. The succession of vegetation, mainly *Saxifraga* sp. and lichens, starts 5–6 years after deglaciation, successively covering up to 30% of the soil surface, then stagnating . Present-day soil-forming processes within the uppermost soil layer comprise initial weathering of primary minerals (chlorites and amphiboles), carbonate dissolution and base cation leaching associated with pH lowering, accumulation of organic matter and nitrogen, and an increase in pedogenically-derived Fe. Soil development measures are time-related and, in general, fit a logarithmic model. The intensity of transformation, including organic carbon and nitrogen accumulation, started at high rates comparable to those reported in Low Arctic and Alpine environments; however, in the fourth/fifth decade after deglaciation it reached a quasi steady-state. Low annual precipitation is probably a crucial factor that controls plant succession and leaching of carbonates, thus limiting mineral weathering, organic matter accumulation and soil development on the Werenskiold moraines.

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1. Introduction

Global climate warming significantly influences the functioning of glacier systems on all continents. The most often reported symptoms of the ongoing changes are: the shortening of mountain glacier tongues (Barry, 2006), reduction of the mass of arctic and subarctic glaciers (Hagen et al., 2006), as well as gradual deepening of the active layer in areas of permafrost occurrence (Humlum et al., 2003; Rachlewicz and Szczucinski, 2008). Contrary to widespread popular opinion, glacier regression did not start in recent decades, but has occurred for a minimum 100–200 years and followed the last period of relatively colder climate — the Little Ice Age (LIA). Glaciers of southern Alaska reached their maximum extent at the end of 18th century, documented by the oldest moraines on the forefield of the Mendenhall Glaciers, assessed to be no younger than 240 years old (Alexander and Burt, 1996). The recession of the Hailogou Glacier in the Gongga Mountains (Sichuan province, south China) and formation of neoglacial moraines started in 1820, as reported by He and Tang (2008). Regular recession of subarctic glaciers on Spitsbergen (the largest island of the Svalbard Archipelago, the High Arctic) began at the end of 19th century or in first decades of 20th century. Unequivocal assessment of the starting point of glacial retreat in this area is highly problematic as the tongues of many glaciers reach the tidewater and deglaciation rates were initially low. Significant shortening of glacier snouts in the High Arctic was confirmed in the second and third decades of 20th century (Ziaja, 2001). The rate of deglaciation in the Svalbard archipelago has accelerated rapidly in the last two decades, up to 700% in some cases. Based on new original calculations, Rachlewicz et al. (2007) distinguished four types of glaciers differing in deglaciation rate: very dynamic, surging tidewater glaciers with post-LIA retreat rates of between 100 and 200 m a^{-1}, other tidewater glaciers receding at a rate of 15 to 70 m a^{-1}, land terminating valley polythermal glaciers with an average retreat of 10 to 20 m a^{-1}, and small, usually cold glaciers with retreat rates below 10 m a^{-1}.

Uncovering new land is an environmentally important result of glacier retreat. According to Rachlewicz et al. (2007), the total surface of glaciers in the central part of Spitsbergen has decreased by between 5 and 53.5% during the 20th century, and the area of recently uncovered ground and rock surfaces has increased by several thousand hectares. Cannone et al. (2008) expect that most small

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glaciers in the Alps (80% of total glacial coverage and an important contribution to water resources) will disappear in the next few decades. Therefore, recently deglaciated areas have increasingly attracted the interest of the scientific community, because soils on glacier forefields offer unique possibilities to study biogenic and abiogenic processes and their interactions induced by accelerated climate changes, yet still under extreme conditions. The rate of glacier transgression and regression (thawing) plays a key role in postglacial landscape formation (Karczewski, 1982; Lonne and Lysa, 2005). It generally decides on the type and quantity of non-cemented deposits in the proglacial zone. Moreover, the rate of deglaciation influences the local microclimate, water drainage patterns, the intensity of erosion, occurrence of cryogenic phenomena in the ground, succession of plant communities, as well as soil evolution (Haugland, 2004; Ziaja, 2004). Ecological changes in the forefields of retreating glaciers are mostly characterized in the context of the succession of plant species or communities (Burga et al., 2010; Cannone et al., 2008; Frenot et al., 1998) and microorganisms (Kastovska et al., 2005). Reports focusing on soil development or transformations of soil properties in deglaciated areas are currently rare and concerned mainly with seasonal or long-term changes in soil respiration and its influence on regional or global carbon turnover as well as on the climate (e.g. Bekku et al., 2004). A limited number of research papers have focused on soil development in the context of weathering processes and the physical transformation of glacial deposits (e.g. Bernasconi and BigLink Consortium, 2008; Burt and Alexander, 1996; Egli et al., 2001; Haugland, 2004; Mavris et al., 2010), and only a few papers involve the problems of the morphological development of the soil profile and discuss the classification of soils currently developing in specific, near-glacial environments (Alexander and Burt, 1996; He and Tang, 2008; Jacobson and Birks, 1980). Extremely rare are reports that present soil chronosequences in the High Arctic (Mann et al., 1986), probably because of the relatively young age and superimposed weak transformation of uncovered glacial deposits.

The objectives of the present study in the forefield of the Werenskiold Glacier were therefore (i) to examine how the morphology and properties of initial soils develop during the first decades after deglaciation; and (ii) to quantify changes in soil properties, including organic carbon accumulation in the upper layer of the soil profile.

2. Study site and methods

2.1. Site characteristics

The study area is in the proglacial foreland of the Werenskiold Glacier (77°05′ N, 15°15′ E), located in Wedel-Jarlsberg Land in south-west Spitsbergen, the largest island of the Svalbard Archipelago in the High Arctic. The Werenskiold Glacier is a typical land-based, polythermal glacier of approximately 27.4 km^2 in area and 9.5 km in length (Hagen et al., 1993). Previous studies focused mostly on mass balance and geometry of the glacier (e.g. Baranowski, 1977), and hydrological (e.g. Krawczyk, 1992), climatic (e.g. Migała et al., 2008), geologic (e.g. Czerny et al., 1993), geomorphologic (e.g. Karczewski et al., 1984), and geochemical (e.g. Bukowska-Jania, 2007) aspects, while the study of soil development on the glacier's foreland and adjacent areas still remains at an initial stage (Piroznikow and Gorniak, 1992; Szerszen, 1968).

The southern part of its proglacial zone is covered with a fluted moraine, changing towards the west into a flat moraine. Moraine tills rest immediately on bedrock schists or glacio-fluvial sands and gravels. The thickness of the neoglacial till does not in general exceed 50 cm. The northern part of the foreland is predominantly covered with glaciofluvial sediments forming a mosaic of plains and fans. The proglacial zone of the glacier is delimited in the north and south by lateral moraines, and in the west by a terminal (frontal) moraine, formed probably during a large and rapid glacial surge at the beginning of the 20th century. Both within the frontal and lateral moraines there are well-preserved glacial cores, making the moraine surface thermoactive in many places and causing frequent landslides on their slopes (Karczewski et al., 1984).

The Werenskiold Glacier basin is located in the contact zone of three tectonic blocks of the Caledonian Hecla Hoek formation (Czerny et al., 1993). The southern neighborhood of the glacier is built of metamorphic Proterozoic amphibolites, quartzites, chlorites and amphibolite-quartzite schists. The eastern surroundings of the accumulation field consist of phyllites, laminated schists and quartzite conglomerates with dolomite precipitations. The Jens Erikfjellet formation, which surrounds the glacier basin from the north-west, is built predominantly of greenschists and mica-calcite-quartzite schists. The rock bed at the glacier terminus, in the area exposed from ice cover, consists of Precambrian chlorite-mica-quartzite schists, marbles and marble–quartzite conglomerates (Czerny et al., 1993). Due to the glacial mixing of these materials, the soil parent material is considered to be polymineral and relatively homogenous in particular sedimentary zones of the proglacial area. Petrographical analyses of pebbles in the southern part of the forefield documented 5–20% of quartz and quartz–mica aggregates, 20–30% of quartzites, 12–21% of chlorite schists, 5–12% of mica schists, 1–9% of amphibolites, and 23–38% of marbles and marble–quartzite conglomerates (Kowalska and Sroka, 2008). Quartzites are not abundant in the source areas, thus — as the most resistant to weathering — are probably over-represented in the pebble fraction. Due to the regular occurrence of marbles in bedrock, the "fresh" glacial sediments always contain calcium carbonate and have an alkaline reaction (Bukowska-Jania, 2007).

Long-term climate data for the Hornsund area have been collected by the permanent polar observatory (Polish Polar Station) at Isbjornhamna, located ca. 12 km south-east of the Werenskiold Glacier. Mean annual precipitation (water equivalent) in the period 1979–2006 was about 430 mm, relatively high when compared to other records from Spitsbergen. Mean monthly precipitation ranges from 20 mm in May to 63 mm in September (Fig. 1). The mean annual air temperature was −4.4 °C, with mean monthly air temperatures ranging from −11.3 °C in January to 4.4 °C in July. A distinct upward trend of mean annual air temperature as high as +0.095 °C/year has been detected during the last three decades of measurements (Marsz and Styszynska, 2007). Soils have a gelic temperature regime, with a mean annual temperature −4.7 °C at a depth of 50 cm below the soil surface. Mean summer (June–August) and winter (December–February) soil temperatures are 1.8 °C and −10.3 °C, respectively (Leszkiewicz and Caputa, 2004). Climate conditions in the foreland of the Werenskiold Glacier are considered to be slightly milder than those in the direct proximity of the polar station at Isbjornhamna (Migała et al., 2008).

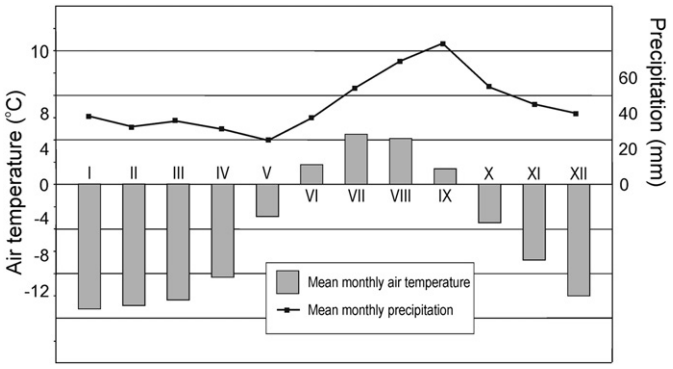

Fig. 1. Mean monthly air temperature and precipitation at the Polish Polar Station (Isbjornhamna-Hornsund, south-west Spitsbergen) in the period 1979–2006. Based on: Marsz and Styszynska, 2007.

2.2. Soil sampling

A preliminary investigation, preceding the final selection of sampling sites, revealed heterogeneous site conditions which are typical on glacier forelands and demonstrated by several soil parameters: proportion of rock fragments, soil texture and wetness, in particular. It was decided to include only the sites on typical moraine till into the present study, while soils on outwash sands and gravels, as well as glaciofluvial and limnic sediments were rejected. The soil pits were placed on relatively well-drained sites, never in local depressions, channels, or on steep slopes. Finally, six sites with increasing distance from the glacier were selected, considering that each sampling site represents a different period and stage of soil development. Soil formation on the moraines began at a time of stabilization of the deposits and has been continued to the present. The age of parent material is therefore considered to approximate the period of soil development. The estimated ages of moraine materials in the proglacial zone were assessed based on numerous previous delineations of the glacier snout position (e.g. Baranowski, 1977; Bukowska-Jania, 2007; Karczewski et al., 1984; Piroznikow and Gorniak, 1992).

The soil chronosequence was located on the southern foreland of the Werenskiold Glacier (Fig. 2). Four sampling sites were placed on a typical fluted moraine (Table 1), one site (WER10) on a flat moraine, and the last one (WER17) in the western (marginal) part of the end (terminal) moraine. All the studied sites were located within a direct distance of 1650 m. The difference in altitude between sites did not exceed 50 m, in the range 25–75 m asl.

Basic geomorphic information and vegetation cover were recorded at each site. All the soil pedons were described according to FAO guidelines (FAO, 2006). Soil texture, structure, consistence, color, redoximorphic features and root abundance in all the derived layers of the soil profile were noted (Table 2). An attempt to classify the soils was made based on the FAO World Reference Base (IUSS, 2007). Sampling depth was different in particular profiles depending on the individual pedon features; however, the first, uppermost sample was always taken at the depth of 0–3 cm. Before sampling, the soil surface was carefully cleaned of lichens and higher plants to avoid an overestimation of organic matter content. All samples were air-dried and crushed to sieve >2 mm and ≤2 mm size fractions.

2.3. Soil analysis

Soil samples were analyzed for texture and basic chemical properties using laboratory methods applied to soil classification purposes (Van Reeuwijk, 2002). Particle size distribution of the ≤2 mm fraction, after removing the organic matter and sample dispersion with heksametaphosphate–bicarbonate solution, was conducted using sand separation on sieves and the hydrometer method to fine earth fractions (<0.1 mm). Soil pH was potentiometrically measured in a 1:2.5 (soil: distilled water) suspension. Calcium carbonate was measured by the gas volumetric method. Total content of organic carbon was determined by dry combustion using an automatic analyzer (Ströhlein CS-mat 5500) and total nitrogen — by the standard Kjeldahl technique. Exchangeable ions (Ca^{2+}, Mg^{2+}, K^{+}, Na^{+}) were extracted initially with 1 M ammonium acetate at standard pH = 7. However, due to unrealistically high calcium and magnesium concentrations (resulting from carbonate dissolution), extraction with 1 M NH_4Cl at pH = 8.2 was additionally applied. Iron in non-silicate forms ("free" iron, Fe_d) was extracted with a bicarbonate–dithionite–citrate buffer (BDC) according to the standard Mehra–Jackson method, and iron in amorphous oxides and hydroxides ("active" iron, Fe_o) was extracted with acid ammonium oxalate according to the standard Tamm method (Van Reeuwijk, 2002). The concentration of metallic elements in all extracts was measured by atomic absorption spectroscopy (Phillips-Unicam AAS).

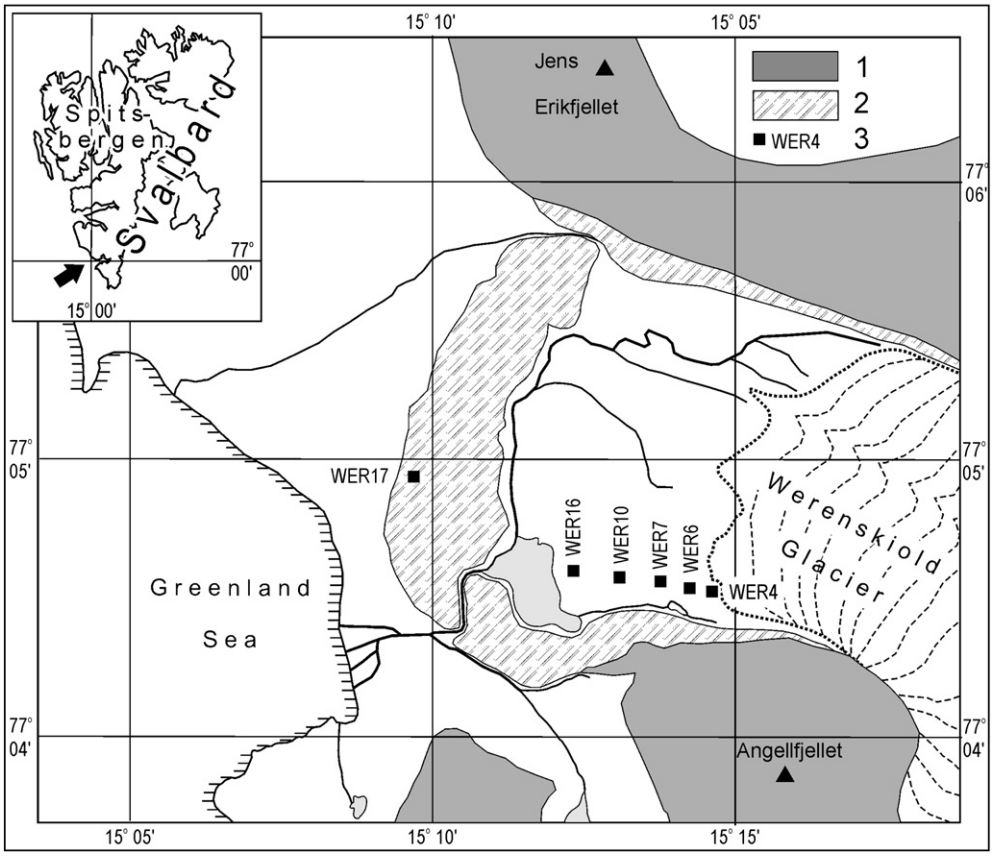

Fig. 2. Location of soil profiles in the forefield of the Werenskiold Glacier. 1 — Mountains without ice cover, 2 — terminal and lateral moraines, 3 — soil pits.

Table 1
Localization of pedons; landforms, ages, kind of parent materials, and vegetative cover of the soils.

Site	Distance to glacier (meters)	Surface age (years)	Landform	Parent material	Bare ground (%)	Structure of vegetation
WER4	20	1	Fluted moraine	Thin layer of till directly on bedrock schists	100	No vegetation
WER6	150	6	Fluted moraine	Thin layer of till on bedrock schists	95	*Saxifraga oppositifolia* (50%[a]), *Saxifraga caespitosa* (50%), single individuals of *Poa arctica*
WER7	280	12	Fluted moraine	Thin layer of till on bedrock schists	90	*S. oppositifolia* (40%), *S. caespitosa* (10%), lichens (50%)
WER10	450	30	Flat moraine	Moraine till on older stratified outwash gravels; permafrost at 90 cm	75	*S. oppositifolia* (50%), *S. caespitosa* (5%), lichens (45%)
WER16	750	45	Fluted moraine	Till on older stratified till and outwash sands; permafrost at 95 cm	70	*S. oppositifolia* (60%), *S. caespitosa* (20%), lichens, single individuals of *P. arctica* and *Cerastium alpinum*
WER17	1650	70–80	Terminal moraine	Deep moraine till; permafrost at 125 cm	70	*S. oppositifolia* (70%), *S. caespitosa* (10%), lichens, single individuals of *P. arctica* and *C. alpinum*

[a] Percentage of surface covered with vegetation.

Soil organic carbon and nitrogen stocks in the upper 15 cm of soil were calculated according to the following equation (Mavris et al., 2010):

$$C_{stock}(N_{stock}) = \sum c_i h_i d_i (1-RM),$$

where C_{stock} (N_{stock}) denotes organic C (total N) abundance (kg m^{-2}), c_i — concentration of C or N (kg t^{-1}), h_i — thickness of i layer (m), d_i — bulk soil density (t m^{-3}), and RM the mass proportion of rock fragments.

Mineralogical composition of the clay fraction (<0.002 mm) was determined by x-ray diffractometry (XRD). The clay fraction was separated by centrifugation, followed by ultrasonic sample dispersion. To remove organic matter, all clay samples were treated with a 10% H_2O_2 solution. Initially, air dried specimens were x-rayed. The Mg^{2+}-saturated slides were treated with glycerol and rescanned. The K^+-treated slides were subsequently heated to 300°C and rescanned. The slides were finally heated again to 550°C and rescanned. Diffraction patterns were obtained using CuKα radiation and step-scanning between 3 and 40° 2θ, using 0.05° 2θ increments with a 3-s counting time. The differential thermal gravimetry (DTA-DTG) method was additionally used to allow for the differentiation of primary versus pedogenic minerals (detailed results were not shown, only the final determination).

The Statistica 8.0 software system (StatSoft Inc., Tulsa, OK) was used to test the statistical significance of trends in changes of soil properties. Various linear and nonlinear models were tested, and the model yielding the highest correlation coefficients (at least 0.05 confidence level) was chosen as the best chronofunctions.

3. Results

3.1. Morphology features and classification of soils

The youngest till in the forefield of the Werenskiold glacier is, in general, only 20–40 cm thick and rests directly on bedrock schists or on older glaciofluvial gravels and sands. The permafrost layer started at a depth of 90–100 cm below ground level, provided no hard rock occurred. The thickness of till was greater on the lateral and frontal moraines, reaching 100–130 cm to contact with the permafrost layer. Recent till in the proximity of the glacier snout is greenish gray (10GY 5-6/1), massive (structureless in a pedological sense) and gleyed throughout due to prolonged saturation with melting water (Table 2, Pedon WER4). The first morphological features of soil development are changes in the soil color to dark greenish gray (10Y 4/1) due to an initial accumulation of organic matter, and formation of a weak subangular blocky (pedogenic) structure. These transformations lead to the distinguishing of the initial humic A horizon having a thickness of up to 3 cm in soils 12–45 years old and 5–6 cm in soils older than 70 years. The initial subsurface B horizon may be distinguished only in better drained soils on the frontal moraine, on materials older than 70 years. Initial B horizons are slightly more brown colored than the subsoil (have a 10Y Munsell color hue instead of 5GY, and chrome 2 instead of 1), have a weak to moderate subangular blocky structure, and are usually interlaced with plant roots throughout. These B layers, however, do not fulfill the requirements of diagnostic horizon *cambic* (IUSS, 2007), due to insufficient thickness and too little differentiation of color when compared to the subsoil.

Soils derived from very shallow till on hard rocks in the forefield of the glacier were classified as Leptosols (Table 2, pedon WER 4). More problematic is classification of soils developed from till resting on hard rock at the depth of 45–50 cm below ground level. The entire volume of till thaws during the polar summer, thus diagnostic horizon *cryic* does not occur (Bockheim et al., 2006). However, features of cryoturbation are visible, particularly as separation of coarse particles within the soil and on its surface. Soils are seasonally saturated with melting water and are "gleyed", however, they do not meet the requirements of "gleyic color pattern" (IUSS, 2007) as the dominant hue of soil color is 10GY. Finally, soils represented in pedons WER6 and WER7 (Table 2) were classified as Leptic Regosols (Calcaric, Turbic, Oxyaquic, Skeletic). All other soils, developed of deep till or till on sand both having a permafrost layer, were classified as Turbic Cryosols (Table 2: pedons WER10, WER16, and WER17).

In the pedon WER16, under the thin cover of young moraine, a unique, complete profile of buried soil was found. It consists of well developed A and B horizons whose cumulative thickness exceeds 25 cm. The subsurface Bw layer meets the requirements of diagnostic horizon *cambic* (IUSS, 2007), and the buried soil represents typical "arctic brown earths" (Tedrow and Hill, 1955) or "rusty tundra soils" (Kowalkowski, 1998), which corresponds to Haplic Cambisols (IUSS, 2007). The present soil (complete pedon WER16) must be, however, classified as Cryosols due to the occurrence of permafrost in subsoil and cryoturbation features at the soil surface (Bockheim et al., 2006; IUSS, 2007). Due to unfavorable climate conditions, Cambisols are not expected to occur in the High Arctic (Goryachkin et al., 2004). Thus, if identified on older Holocene surfaces, are probably relic soils (Kabala and Zapart, 2009).

3.2. Particle-size distribution

The share of individual particle-size fractions within the profiles of soils developed from till varies only insignificantly in terms of space and depth. The skeleton fraction (fine and medium gravel) usually forms 40–60% of the overall soil mass. The content of the silt fraction (0.002–0.05 mm) is in the range of 40–45% of the fine-earth fractions

and the clay fraction (<0.002 mm) is usually in the range of 9–15% (Table 3). The common texture class of soils developed on till is therefore loam (according to the USDA classification). The only exception is pedon WER17 located on the end moraine, where sandy loam was found in the upper soil layers due to the lower content of clay (6–7%) and silt (25–36%) fractions. Decrease in the soil skeleton due to the duration of physical weathering was weak, if any. The mean content of rock fragments within the layer 0–15 cm changed from 42% in WER4 soils to 38% in WER10, however, it exceeded 50% in older soils. Despite nearly uniform texture, a decrease in the clay fraction in the uppermost soil layer (0–3 cm) is distinctly visible (Fig. 3a) when compared to the chronosequence from the glacier snout (15% of clay in pedon WER4) to the terminal moraine (5% of clay in pedon WER17).

3.3. Carbonates, pH and exchangeable cations

The dissolution and leaching of carbonates is reported to be the first feature of the chemical transformation of glacial sediments (Bukowska-Jania, 2007). All soils developed on the recent glacial sediments contain variable, but considerable amounts of $CaCO_3$ (up to 5%). Significant diversification of $CaCO_3$ content was observed in sediments which are 10–12 years old (pedon WER7) but only in the upper 0–3 cm layer, while in the 45–70 year old formations (WER16) the leaching of $CaCO_3$ reaches the depth 12–14 cm from the soil surface (Fig. 4). However, a temporal trend in $CaCO_3$ leaching is not evident, and excluding the uppermost layer of pedon WER16, the $CaCO_3$ content was at least as high as 2.3%. Only the buried

Table 2
Soil morphology and profile descriptions.

Horizon	Depth (cm)	Color (moist)	Texture, rock fragments	Structure	Consistence (moist)	Carbonates	Roots	Boundary
Pedon WER4: Haplic Leptosol (Calcaric, Oxyaquic, Skeletic) age of glacial till — 1 year								
Cg1	0–3	10GY 6/1	L, MC — A	MA	FR	MO	N	G
Cg2	3–18	10GY 5/1	L, MC — M	MA	FR	MO	N	A
R	18+	Hard, fractured bedrock schists						
Pedon WER6: Leptic Regosol (Calcaric, Turbic, Oxyaquic, Skeletic) age of glacial till — 6 years								
Cg1	0–3	10GY 5/1	L, MC — A	PM	FR	MO	VF	G
Cg2	3–12	10GY 5/1	L, MC — M	PM/PL, WE	FR	MO	VF	G
Cg3	12–30	10GY 5/1	L, ME — M	PL,WE	FR	MO	N	C
Cg4	30–45	10GY 5/1	L, MC — A	MA	FR	MO	N	C
R	45+	Hard, fractured bedrock schists						
Pedon WER7: Leptic Regosol (Calcaric, Turbic, Oxyaquic, Skeletic) age of glacial till — 12 years								
A	0–3	10Y 4/1	L, MC — A	SB, WE	FR	MO	F	G, W
AC	3–6	5GY 5/1	L, MC — M	SB, WE	FR	MO	F	G
Cg1	6–15	10GY 5/1	L, MC — M	AB, WE	FR	MO	VF	G
Cg2	15–30	10GY 5/1	L, MC — M	AB/PL, MO	FR	MO	N	G
Cg3	30–45	10GY 5/1	L, MC — A	AB, MO	FR	MO	N	C
R	45+	Hard, fractured bedrock schists						
Pedon WER10: Turbic Cryosol (Calcaric, Oxyaquic, Skeletic) age of glacial till — 30 years (glacial till on older stratified outwash gravels)								
A	0–3	5 G 4/1	L, MC — M	SB, WE	FR	MO	F	G, W
AC	3–8	5 G 5/1	L, MC — M	SB/PL, WE	FR	MO	VF	G
Cg	8–20	5 G 5/1	L, MC — M	PL, WE	FR	MO	N	A
2C	20–90	2.5Y 3/2–3	S, MC — A/D	SG	LO	MO	N	C, W
		Stone lines at the depths 35–44 cm and 60–70 cm						
I	90+	Permafrost layer						
Pedon WER16: Thaptocambic Turbic Cryosol (Calcaric, Oxyaquic, Skeletic) age of glacial till — 45 years (glacial till on older stratified glacial till and outwash gravels)								
A	0–3	10Y 4/1	L, MC — A	SB, WE	FR	SL	C	G
AC	3–10	10GY 5/1	L, MC — A	PL, WE	FR	MO	F	G
Cg	10–14	2.5Y 5/3	SiL, FI — V	MA	FI	MO	VF	A, S
		No pavement at the boundary of horizons						
2Ab	14–20	2.5Y 4/3	SL, MC — M	SB, ST (fine)	FR	SL	C	G
2ABb	20–26	2.5Y 3/3	SL, MC — M	AB, MO	FR	SL	C	G
2Bwb	26–40	2.5Y 5/4	SL, MC — M	AB/PL, MO	FR	SL	F	C
		Thin iron pan at the depth ca 35 cm						
		Stone pavement at the boundary of horizons						
3C	40–70	2.5Y 6/6	SL/S, MC — D	MA/SG	FI/LO	SL	VF	C
4Cx	70–90	2.5Y 5/5	SL, MC — A	AB/PL, ST	FI	SL	N	C,W
I	90+	Permafrost layer						
Pedon WER17: Turbic Cryosol (Calcaric, Oxyaquic, Skeletic) age of glacial till — more than 70 years								
A	0–6	10Y 4/1	SL, MC — A	SB, MO	FR	MO	C	G
ABg	6–12	10Y 5/2	SL, MC — A	SB, MO	FR	MO	C	G
BCg	12–22	5GY 5/2	L, MC — A	SB, MO	FR	MO	F	G
Cg1	22–40	5GY 5/1	L, MC — A	AB, MO	FR	MO	VF	G
Cg2	40–125	5GY 4/1	L, MC — A	AB/PL, MO	FR	MO	N	A, W
I	125+	Permafrost — “ice nucleus” of the moraine						

Textural classes: S — sand, SL — sandy loam, L — loam, SiL — silt loam; dominant size of rock fragments: FI — fine (2–6 mm), ME — medium (6–20 mm), MC — medium and coarse gravel (6–60 mm); abundance of rock fragments: V — very few (0–2%), F — few (2–5%), C — common (5–15%), M — many (15–40%), A — abundant (40–80%), D — dominant (>80%); types of soil structure: SG — single grain, MA — massive, PM — porous massive, AB — angular blocky, AS — angular and subangular blocky, SB — subangular blocky, PL platy; grade of development: WE — weak, MO — moderate, ST — strong; consistence (moist): LO — loose, FR — friable, FI — firm; carbonates content: SL — slightly calcareous (<2%), MO — moderately calcareous (2–10%), abundance of roots: N — none, V — very few, F — few, C — common; horizon boundary (distinctness, topography): A — abrupt, C — clear, G — gradual, S — smooth, W — wavy.

Table 3
Particle-size distribution and chemical characteristics of soils in the foreland of the Werenskiøld Glacier.

Horizon	Depth (cm)	Particle size distribution, % (particle diameters in mm)				pH_{H2O}	$CaCO_3$ (%)	Organic carbon	Total N	Fe_o	Fe_d	Fe_o/Fe_d
		>2[a]	2.0–0.05[b]	0.05–0.002[b]	<0.002[b]			g kg^{-1}				
Site WER4												
Cg1	0–3	50	40	45	15	8.35	3.0	4.5	0.2	4.7	10.4	0.46
Cg2	3–10	42	40	44	16	8.36	3.3	4.6	0.2	4.6	10.3	0.45
Cg2	10–18	38	40	45	15	8.38	3.6	4.2	n.d.	4.6	10.3	0.45
Site WER6												
Cg1	0–3	45	42	45	13	8.12	2.7	6.4	0.3	5.1	10.5	0.49
Cg2	3–12	38	42	43	15	8.35	3.5	5.2	0.2	4.9	10.2	0.48
Cg3	12–30	35	41	43	16	8.32	3.8	4.5	n.d.	4.8	10.1	0.48
Cg4	30–45	49	49	39	12	8.33	3.8	3.9	n.d.	4.4	9.5	0.46
Site WER7												
A	0–3	47	54	35	11	7.95	2.9	10.4	0.8	5.6	10.8	0.52
AC	3–6	38	49	39	12	8.13	4.6	5.0	0.3	5.3	10.7	0.50
Cg1	6–15	39	43	45	12	8.29	4.5	4.3	n.d.	5.0	9.5	0.52
Cg2	15–30	33	42	45	12	8.34	4.6	5.1	n.d.	4.2	8.9	0.47
Cg3	30–45	47	41	45	14	8.29	4.9	4.8	n.d.	4.6	9.7	0.47
Site WER10												
A	0–3	42	45	45	10	7.82	3.2	11.0	1.1	6.9	12.4	0.56
AC	3–8	38	41	45	14	8.20	4.1	5.6	0.4	6.7	12.7	0.53
Cg1	8–20	37	44	42	14	8.41	4.6	5.5	n.d.	5.9	11.9	0.50
2C	20–33	72	91	7	2	8.36	1.6	1.6	n.d.	1.7	13.5	0.13
2C	50–60	82	89	8	3	8.33	1.5	1.4	n.d.	1.5	8.7	0.17
2C	70–80	60	90	8	2	8.35	1.8	1.2	n.d.	0.8	11.2	0.07
Site WER16												
A	0–3	55	44	47	9	7.73	0.9	13.3	1.1	7.0	13.1	0.53
AC	3–10	50	40	48	12	8.20	2.8	5.7	0.5	6.8	13.2	0.52
Cg	10–14	6	33	62	5	8.25	3.6	4.5	0.4	6.3	13.5	0.47
2Ab	14–20	37	69	27	4	7.31	1.7	14.6	1.7	8.7	14.8	0.59
2ABb	20–26	37	62	24	4	7.22	1.6	13.1	1.2	7.3	14.3	0.51
2Bwb	26–40	36	68	29	3	7.08	0.8	4.3	n.d.	3.9	15.5	0.25
3C	40–70	80	67	31	2	7.09	0.6	3.3	n.d.	2.5	14.0	0.18
4Cx	70–90	81	68	30	2	7.21	1.8	1.6	n.d.	2.3	11.0	0.21
Site WER17												
A	0–6	55	58	35	7	7.74	2.3	19.2	1.7	7.2	12.3	0.59
ABg	6–12	58	58	36	6	7.92	2.6	6.7	0.6	6.3	11.1	0.57
BCg	12–22	60	68	25	7	8.16	3.3	4.4	0.4	4.7	9.2	0.51
Cg1	22–40	55	51	40	9	8.23	3.3	4.0	n.d.	4.7	10.2	0.46
Cg2	40–70	60	49	41	10	8.23	3.2	4.4	n.d.	4.3	9.5	0.45
Cg2	70–90	60	48	41	11	8.24	4.1	4.9	n.d.	4.4	10.9	0.40

[a] Percentage of bulk soil sample.
[b] Percentage of fine earth fractions (<2 mm).

older outwash gravels, found in the lower part of pedons WER10 and WER16, contained less $CaCO_3$, namely 0.6–1.8% (Table 3).

Despite the relatively short time of transformation, the change in soil pH was evident ($p<0.05$). Initial pH of the recent moraine till was about 8.3–8.4 throughout the till layer. The value of pH in the uppermost 0–3 cm soil layer falls below 8.0 already in 12-year old sediments and below 7.8 in 30-year-old sediments. The pH value decreased logarithmically with time (Fig. 3b); however, it tended to stabilize at pH = 7.7 in sediments older than 40 years due to the presence of carbonates. The acidification of a deeper layer (3–6/8 cm) was very weak: although the pH value decreased below 8.2 in 12-year-old sediments, the level of 8.0 was reached only in soils older than 70 years (Table 3). No significant acidification was observed in moraine till below the depth of 10 cm within the observed time span.

The overall sum of exchangeable base cations (Ca^{2+}, Mg^{2+}, K^+, Na^+) in the soils under investigation is influenced primarily by the large amounts of calcium released easily from $CaCO_3$, which explains the differences between its amount when extracted at pH = 7 (with ammonium oxalate) and at pH = 8.2 (with ammonium chloride). Ammonium oxalate at pH = 7 extracted in such soils ca. 5-fold higher amounts of calcium, twofold more magnesium, 10-fold more potassium and twofold more sodium when compared to extraction at pH = 8.2 (Table 4). The calcium is therefore "extractable" or "soluble" rather than "exchangeable" in these soils. Decrease in the total sum of base cations observed in the uppermost layers of soil in the chronosequence is correlated with the dissolution of carbonates and leaching of calcium ions. Change in the vertical distribution of "exchangeable" Ca in particular soil profiles was more evident than differences in carbonate content or pH. While in the youngest till (pedon WER4) Ca content increased towards the soil surface, in sediments 6-year old and older the calcium content significantly decreased towards the surface (Table 4, extraction with ammonium chloride). The loss of "exchangeable" Ca in the layer 0–3 cm reached 23% of the initial Ca level after 6 years of leaching, 45% after 12 years and 57% after 70 years of cation leaching, whereas the loss of $CaCO_3$ reached maximally 30% (in the same horizons). This time-dependent trend was however not significant statistically, as some medium-age soils (pedons WER 10 and 16) revealed a smaller loss of Ca than expected on the basis of their age.

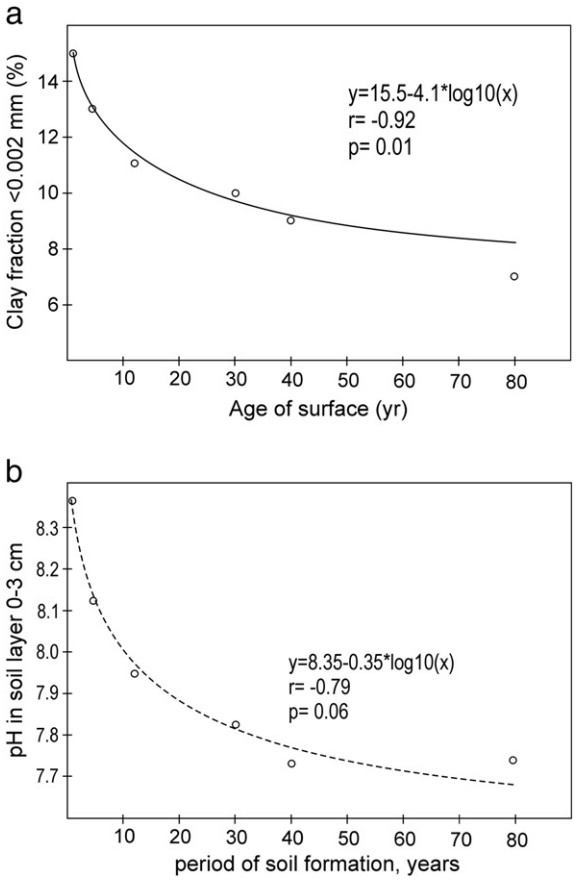

Fig. 3. Change in clay content (a) and soil pH (b) in the uppermost soil layer (0–3 cm) with time of soil development.

Enhanced Ca leaching narrowed the proportion of "exchangeable" Ca to Mg, which in recent till had an initial value 4.5 or more, and decreased with age to 2.5. A similar change was found in the relationship of "exchangeable" Ca to K, which decreased from the level of 38–40 to 15–20 (in the soil layer 0–3 cm). The most spectacular change in the proportion of exchangeable cations was however found between Ca and Na, which decreased from 42–45 in recent till to about 10 in 70-year old sediments. It resulted both from Ca loss and from significant increase in Na content in the layer 0–3 cm.

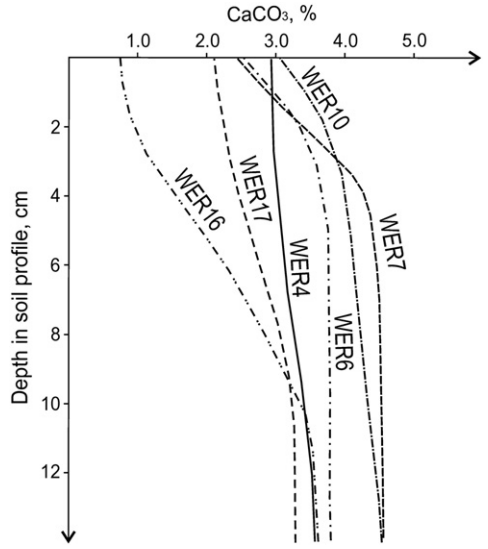

Fig. 4. Calcium carbonate distribution in the upper soil layers.

Table 4
Comparison of exchangeable ions extracted with NH_4OAc pH 7.0 and NH_4Cl pH 8.2.

Horizon	Depth (cm)	Exchangeable ions extracted with NH_4OAc pH 7.0 (cmol[+] kg^{-1})				Sum	Exchangeable ions extracted with NH_4Cl pH 8.2 (cmol[+] kg^{-1})				Sum
		Ca	Mg	K	Na		Ca	Mg	K	Na	
Site WER4											
Cg1	0–3	30.0	3.83	0.70	0.20	34.7	6.64	1.39	0.16	0.16	8.4
Cg2	3–10	34.4	4.05	0.81	0.31	39.6	6.42	1.45	0.18	0.12	8.2
Cg2	10–18	30.4	4.02	0.73	0.30	35.5	6.00	1.60	0.21	0.12	7.9
Site WER6											
Cg1	0–3	24.0	0.68	3.02	0.27	28.0	4.74	0.99	0.13	0.14	6.0
Cg2	3–12	26.4	0.73	2.54	0.28	29.9	6.16	1.19	0.15	0.14	7.6
Cg3	12–30	28.4	0.67	2.72	0.34	32.1	6.20	1.38	0.19	0.16	7.9
Cg4	30–45	26.8	0.63	2.36	0.24	30.0	5.98	1.02	0.11	0.10	7.2
Site WER7											
A	0–3	12.9	5.62	0.46	0.10	19.1	3.30	0.95	0.06	0.28	4.6
AC	3–6	18.7	2.78	0.55	0.13	22.2	4.48	1.36	0.08	0.12	6.0
Cg1	6–15	26.8	3.96	0.62	0.18	31.6	5.26	1.15	0.11	0.16	6.7
Cg2	15–30	25.2	3.83	0.57	0.17	29.8	5.50	1.13	0.12	0.16	6.9
Cg3	30–45	25.6	3.26	0.82	0.26	29.9	5.94	1.19	0.13	0.14	7.4
Site WER10											
A	0–3	27.2	0.91	1.64	0.06	29.8	5.38	1.84	0.12	0.26	7.6
AC	3–8	34.8	0.75	1.53	0.05	37.1	6.32	1.57	0.14	0.19	8.2
Cg1	8–20	37.6	0.62	1.40	0.05	39.7	6.62	1.66	0.16	0.23	8.7
2C	20–33	35.6	0.76	1.35	0.05	37.8	5.56	1.41	0.03	0.17	7.2
2C	50–60	32.8	0.83	1.86	0.03	35.5	4.80	1.08	0.02	0.16	6.1
2C	70–80	26.8	0.65	1.35	0.05	28.8	5.78	1.64	0.05	0.33	7.8
Site WER16											
A	0–3	18.7	1.58	0.46	0.24	21.0	4.94	1.35	0.14	0.14	6.5
AC	3–10	21.2	2.68	0.68	0.21	24.8	5.72	1.53	0.15	0.10	7.5
Cg	10–14	29.2	5.10	0.67	0.16	35.1	5.00	0.91	0.12	0.07	6.1
2Ab	14–20	36.8	6.97	0.82	0.04	44.6	1.82	0.89	0.04	0.02	2.8
2ABb	20–26	32.4	5.10	0.75	0.09	38.3	0.88	0.65	0.06	0.05	1.6
2Bwb	26–40	37.6	6.05	1.02	0.13	44.8	0.72	0.48	0.07	0.04	1.3
3C	40–70	39.6	4.71	0.92	0.11	45.3	0.38	0.34	0.05	0.03	0.8
4Cx	70–90	37.6	10.10	0.98	0.23	48.9	1.32	1.02	0.03	0.02	2.4
Site WER17											
A	0–6	27.2	2.55	0.52	0.26	30.5	2.62	1.03	0.17	0.24	4.1
ABg	6–12	31.2	1.75	0.59	0.16	33.7	3.92	0.80	0.13	0.09	4.9
BCg	12–22	33.2	1.82	0.44	0.18	35.6	4.54	0.79	0.13	0.10	5.6
Cg1	22–40	34.3	2.56	0.38	0.17	37.4	4.60	1.26	0.16	0.14	6.2
Cg2	40–70	38.6	2.50	0.27	0.18	41.5	6.68	1.07	0.16	0.12	8.0
Cg2	70–90	37.8	3.42	0.15	0.16	41.5	5.88	1.16	0.16	0.10	7.3

3.4. Soil organic carbon and soil nitrogen

Accumulation of organic carbon in the uppermost 3 cm of soil showed a polynomial increase with time over the observed period of time (Fig. 5a). Its concentration in fine earth fractions ranged from 4.5 g kg^{-1} in the newly deglaciated till (throughout till layer), to 19.2 g kg^{-1} in the oldest soil. Increase in organic C concentration was observed mainly within the uppermost 3 cm of the soil, and a weak increase was initiated at greater depths (3–8/10 cm) starting from the fourth decade of soil development (Table 3). Organic C stocks, calculated for the upper 15 cm of soils, increased gradually with soil development, but rates of accumulation declined logarithmically with soil age (Fig. 5b). In the 80-year-old soil, organic C stock increased to 0.76 kg m^{-2} with an average accumulation rate of 4.4 g m^{-2} a^{-1}, however, the annual rate decreased from ca. 20 g m^{-2} a^{-1} in the first decade after deglaciation to 1 g m^{-2} a^{-1} in the fifth and following decades. It is noteworthy that the content of organic C in the buried soil (pedon WER16) was relatively high, up to 14.6 g kg^{-1} throughout 12 cm deep A and AB layers. Carbon stock calculated for the upper 15 cm of buried soil was 1.52 kg m^{-2}, that is at least twice as much as in 80-year-old modern soil.

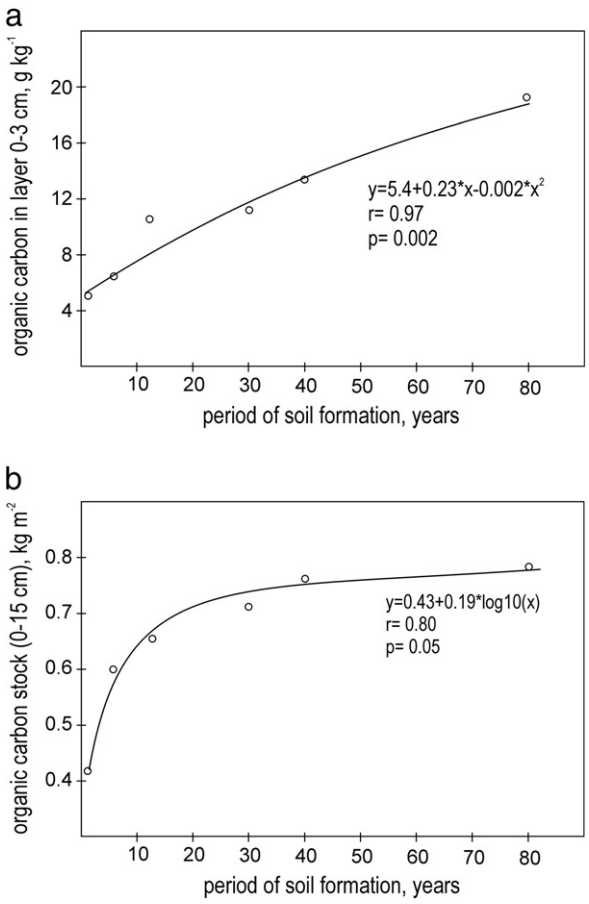

Fig. 5. Increase of organic carbon content in a soil layer 0–3 cm (a) and stocks of organic carbon in the upper 15 cm of soil (b).

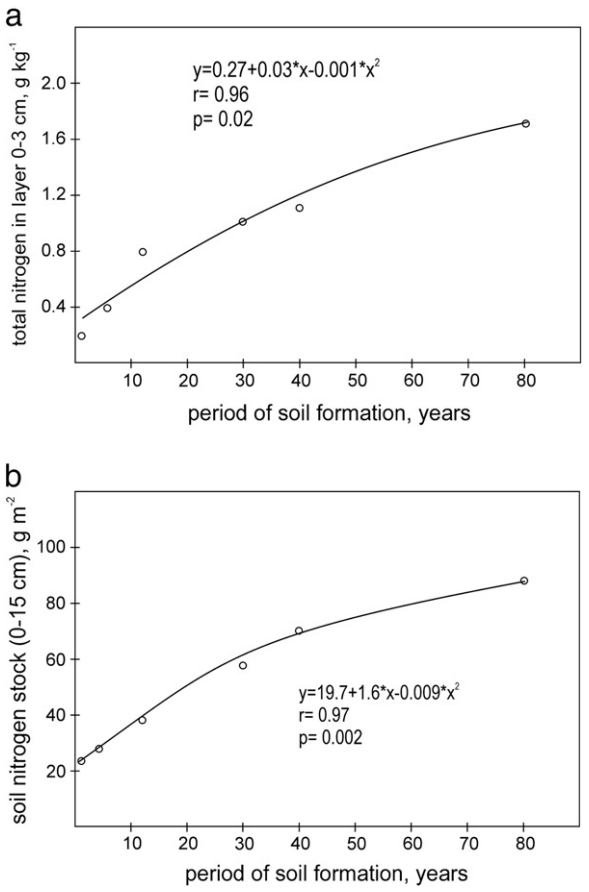

Fig. 6. Increase of total nitrogen content in a soil layer 0–3 cm (a) and stocks of nitrogen in the upper 15 cm of soil (b).

Distribution of total soil N within and among profiles was similar to that for organic C but differed in magnitude (Table 3). Accumulation of soil N in the uppermost 0–3 cm layer showed a rather polynomial function of time and its concentration ranged from nearly zero in the newly deglaciated till to 1.7 g kg^{-1} in the 80-year-old soil (Fig. 6a). Within 80 years, soil N in the upper 15 cm of soil had accumulated to about 85 g m^{-2}, at an average rate of 1.1 g m^{-2} a^{-1}. In contrast to the logarithm model of organic carbon stock, the polynomial model provided a better fit to the increasing N accumulation across the chronosequence (Fig. 6b). The annual rate of N accumulation decreased, but not so radically when compared to carbon, e.g. from ca. 1.1 g m^{-2} a^{-1} in the first four decades after deglaciation to 0.6 g m^{-2} a^{-1} in the following decades.

3.5. Pedogenically-derived Fe

"Free" iron in these soils, as indicated by dithionite citrate–extractable Fe (Fe_d) was relatively low, in the range 8.7–13.5 g kg^{-1}, where up to 10.4 g kg^{-1} of Fe_d appears to be inherited from the parent glacial material, and was weakly differentiated within and among pedons across the chronosequence (Table 3). More evident was a rise of acid oxalate–extractable Fe (Fe_o) that is a measure of the most reactive Fe forms, e.g. water-soluble, exchangeable, poorly crystalline or short-range ordered, and a fraction of organically-bound Fe (Loeppert and Inskeep, 1996). Starting from an initial 4.6 g kg^{-1} (inherited from the parent glacial material) throughout the till in pedon WER4, Fe_o increased up to 7.2 g kg^{-1} in the uppermost soil layer at site WER17. The annual increase of Fe_o was significantly higher during the first three decades of soil development when compared to the following decades, thus the logarithm model gives the best fit to Fe_o accumulation in the 0–3 cm layer (Fig. 7a). As early as the second decade after deglaciation a weak Fe_o increase was observed at the 6–8 cm depth and more, but no deeper than 12 cm below the soil surface even in the oldest soil (pedon WER17). Older glacio-fluvial materials had a significantly lower Fe_o content (0.8–2.5 g kg^{-1} in parent material), thus the pedogenically-induced increase of Fe_o in buried soil was evident in the AB horizon, where it was nearly 4-fold higher that in its parent material (Table 3).

The ratio of Fe_o:Fe_d has been called an "activity ratio" and has been used as a relative measure of the crystallinity of free iron oxides (Loeppert and Inskeep, 1996). Similar to Fe_o, the ratio Fe_o:Fe_d increased logarithmically in the 0–3 cm soil layer across the chronosequence. Initially, the ratio was below 0.46, while in soils 30–50 years old it exceeded a level of 0.53–0.55 and maximally rose to 0.59 indicating a greater proportion of poorly crystalline Fe (relative to total pedogenic Fe) in older soils (Fig. 7b). Both pedogenically-derived Fe forms and the "activity ratio" increased towards the soil surface and no features of iron translocation were detected within pedons across the chronosequence (Table 3).

3.6. Clay fraction mineralogy

Numerous authors have reported the low intensity of mineral transformation and a lack of detectable differences in mineralogy within the profiles of recent arctic soils on Spitsbergen (Klimowicz et al., 2009; Skiba et al., 2002; Szerszen and Chodak, 1983). However, these conclusions are based on soil profiles situated on various parent materials, but never on those combined in chronosequences. It was therefore decided to compare the clay fraction (<0.002 mm) mineralogy of the uppermost layer of soils across the chronosequence in the forefield of the Werenskiold Glacier.

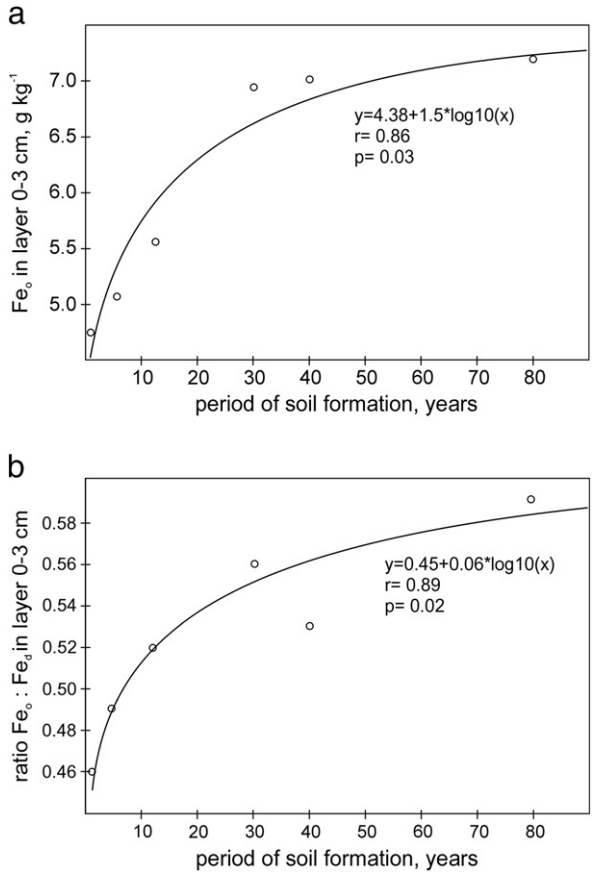

Fig. 7. Changes in (a) Fe_o concentration, and (b) Fe_o:Fe_d ratio in soil layer 0–3 cm as a function of time.

The clay fraction appears to be relatively uniform among pedons and in close relation to the mineral composition of rocks surrounding and underlying the glacier. XRD scanning of all clay samples after standard preparation (Mg^{2+} and K^+ saturation, glycol solvation, and heating at 350, and 550 °C), supported by DTA-DTG analyses, confirmed the occurrence of primary minerals and the lack or only traces of secondary (pedogenic) clay minerals. Primary minerals, inherited from mica-chlorite schists-trioctahedral chlorites and micas (presumably sericite) — dominated the clay fraction of all samples (Fig. 8). In a recent till (site WER4) they were accompanied by amphiboles, mainly Fe–hornblende, feldspars and dolomite. Quartz was absent in the clay fraction while in the coarser fractions it was regularly present. Both XRD and DTA-DTG methods did not indicate clearly the presence of kaolinite, and peaks 0.716 and 0.353–0.355 were often stronger after heating to 550 °C than at room temperature. However, kaolinite identification in the presence of chlorites in a clay fraction requires application of an infrared adsorption technique (Kodama and Oinuma, 1963) that was not conducted and needs to be completed in the course of further studies.

Some distinct changes in clay fraction mineralogy were found among soils across the chronosequence: disappearance of dolomite, decrease in amphibole content, and inversion of the relative ratio of chlorite to mica. The last ratio was calculated based on intensities of the strongest peaks: 1.021 nm for mica and 0.716 nm for chlorite. In a recent till the chlorite:mica ratio was nearly 3.5, but in the first three decades it decreased to ca. 1.0 and finally, in 80-year-old soil, to 0.74, indicating relatively rapid transformation of the clay-size primary chlorites. The changes in the chlorite:mica ratio fitted the logarithmic model (Fig. 9).

4. Discussion

The results of the present study indicate that soil development on moraines of the Werenskiold glacier is, in general, a function of time. However, the accumulation of substances and other changes in soil chemical properties rarely fit a simple linear model over the observed time span of 80 years. Even in such a short time the logarithm model provided a better fit for the studied phenomena. Similar trends were also observed on other glacial chronosequences, usually covering a much longer period of soil development (e.g. Burt and Alexander, 1996; Egli et al., 2001; He and Tang, 2008; Jacobson and Birks, 1980; Mavris et al., 2010).

Weathering processes and soil development are controlled by climate and succession of vegetation (Velde and Meunier, 2008). Rapid mineral weathering accompanied by high rates of organic matter accumulation in soil are common in cool, humid climates (Burt and Alexander, 1996). Thus, physical weathering (freeze-thaw weathering), in particular, is supposed to proceed quickly under an arctic climate (Skiba et al., 2002; Szerszen and Chodak, 1983). As

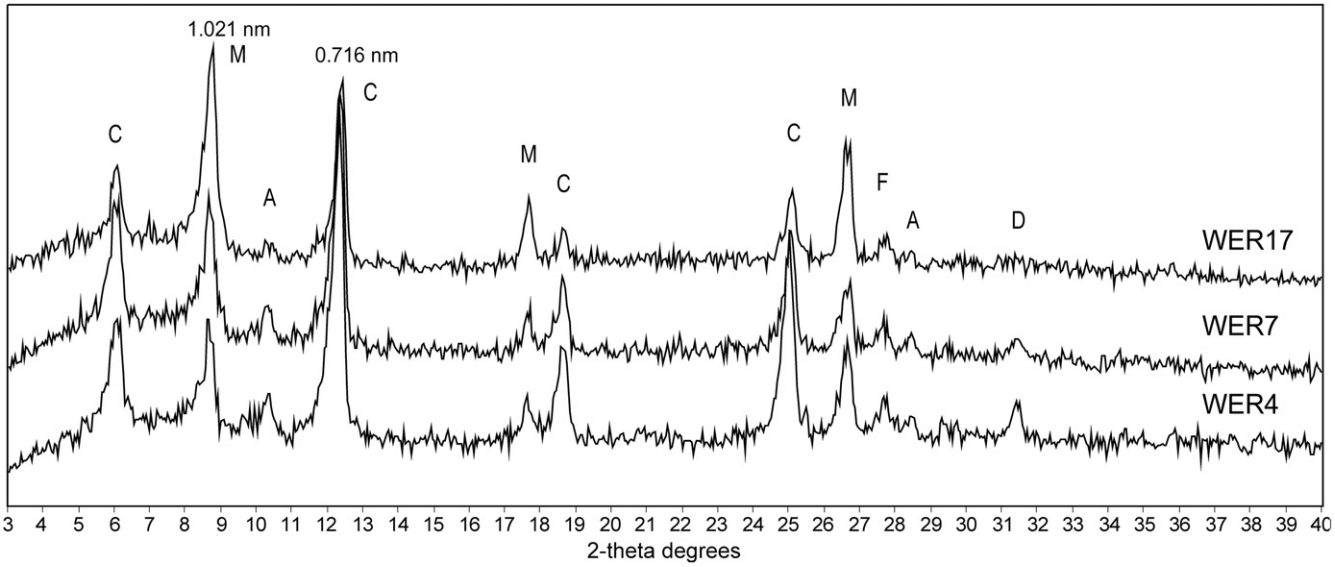

Fig. 8. The x-ray diffractograms of clay fraction (<0.002 mm), air-dried specimens. Diagnostic peaks of mica (1.021 nm) and chlorite (0.716 nm) for calculation of chlorite:mica ratio are indicated. A — amphibole, C — chlorite, D — dolomite, F — feldspars, M — mica.

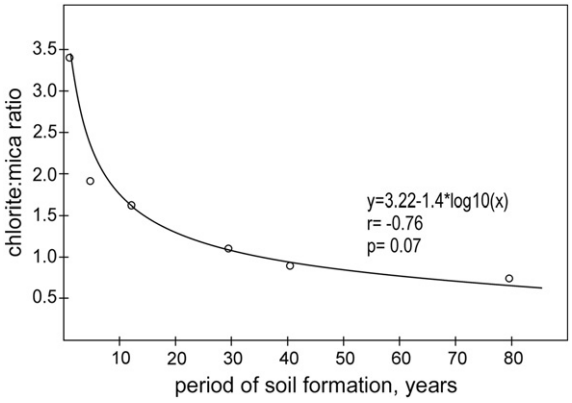

Fig. 9. Changes of chlorite:mica ratio in clay fraction of soils as a function of time.

indicated by the particle size distribution pattern, all of the studied soils have a relatively uniform texture. The stable content of the soil skeleton and decrease in the clay fraction in the uppermost soil layer contradict the observations made in other chronosequences on glacial moraines (He and Tang, 2008; Mavris et al., 2010). Righi et al. (1999) found that the increase in clay particle-size in soils derived on post-glacier moraines was time-dependent and relatively fast during the early stages of soil development. The loss of the clay fraction from the uppermost layer of soils across the chronosequence on the Werenskiold moraines may be therefore apparent and result from simple spatial (zonal) variability of glacial sediments. Moreover, the washing out of fine earth fractions from the soil surface is possible every spring, when melting water creates numerous seasonal micro-streams on the moraine. Washing out, most intense close to the glacier, may explain the larger clay loss on recent moraines and the relative stabilization on older ones (Fig. 3a). However, clay content increases with depth in some soil profiles. The possibility of clay translocation in arctic soils due to cryoturbation or formation of internal aggregate soil structure has been mentioned by Alexander and Burt (1996) and Jacobson and Birks (1980).

The 80-year-long period is too short to expect a spectacular transformation in chemical properties or formation of secondary clay minerals (Velde and Meunier, 2008). However, the results of chemical and mineralogical analyses provide some evidence of chemical weathering in this environment. The primary feature of moraine till at the Werenskiold glacier is the relatively high content of carbonates, inherited from bedrock or newly precipitated from stagnating glacial water (Bukowska-Jania, 2007). Progressive carbonate dissolution is viewed as being controlled by a combination of the influences of water supply and percolation within the soil, CO_2 production, temperature, surface properties of carbonates (Anderson et al., 2000; Mavris et al., 2010) and production of organic acids (Burt and Alexander, 1996). Many studies have shown that carbonate minerals disappear from parent materials in humid regions within a few hundred years of soil development (e.g. Alexander and Burt, 1996; Bain et al., 1993; He and Tang, 2008), and the leaching of the soil material may be initially rapid (Crocker and Dickson, 1957). Our results evidence significant but not total dissolution of carbonates even in the uppermost soil layer, where at least 0.8–2% of $CaCO_3$ was still found 80 years after deglaciation (Fig. 4). The front of carbonates dissolution descended maximally to 10–12 cm below the soil surface.

The only obvious primary mineral change in the clay fraction (excluding carbonates) within 80 years of soil development was amphibole and chlorite depletion (Fig. 8). Similar successive disappearance of hornblenda was reported by Mavris et al. (2010) and assumed its transformation into smectite. Munroe et al. (2007) documented a total transformation of chlorites inherited from bedrocks to hydrobiotite and vermiculite under the alpine conditions of Vermont, Canada. Other studies demonstrated the apparent instability of micas in glacial sediments, in particular that of biotite, and the relatively rapid transformation to hydrobiotite and illite, then to vermiculite, and — in a last step — to smectite (Egli et al., 2001; He and Tang, 2008; Mavris et al., 2010). Despite the successive disappearance of hornblenda and chlorite, no evidence of secondary illite, hydrobiotite, vermiculite or smectite were detected in the clay fraction of soils across the chronosequence. Similarly, Szerszen and Chodak (1983) reported primary mineral occurrence and a lack of secondary phyllosilicates in the clay fraction of moraine soils at the Werenskiold glacier. A possible explanation for mica stability in the soils under study is the predominance of sericite inherited from bedrock. Dioctahedral muscovites and sericites are more resistant to chemical weathering than trioctahedral biotites, particularly under alkaline conditions and an excess of Ca^{2+} and Mg^{2+} ions (Fanning et al., 1989). The beginnings of mica chemical weathering and the transformation of primary chlorite to vermiculite are assumed for these soils as evidenced at least by an increasing content of "pedogenic" iron (Table 3). Initial secondary products of these processes are however still "masked" by intense signals from prevailing primary micas, or leached from the uppermost soil layer (as evidenced by the decrease in clay content). Moreover, both the decrease of chlorite:mica (Fig. 9) and Fe_o:Fe_d (Fig. 7b) ratios fit the logarithm model, indicating the weakening of weathering intensity with time. The other explanation of the relative increase of primary mica in clay-size fractions is the successive fragmentation of sand- and silt-sized micas due to physical weathering (Szerszen and Chodak, 1983). Any time-dependent trend in these fractions was however not evident across the chronosequence. Further study is therefore necessary to understand the relationships between time and clay-sized particles, and secondary mineral formation in these soils.

Low weathering intensity, including weak carbonate leaching, is assumed to be at least in part related to the slow succession of vegetation. It is well documented that plants and microorganisms can significantly increase the rate of weathering of minerals, and through the exudation of protons, low molecular weight organic acids, or siderophores, leading to ligand-promoted mineral dissolution (Bernasconi and BigLink Consortium, 2008). The intensity of soil development is relatively more intense under continuous vegetation cover, and significantly accelerates as the forest becomes established (Burt and Alexander, 1996; He and Tang;, 2008; Mavris et al., 2010). Unfortunately, the harsh arctic climate on the Svalbard Islands eliminates forest vegetation and controls the succession of vascular plants in general (Piroznikow and Gorniak, 1992). We found that the first flowering plants (scattered individual examples of *Saxifraga oppositifolia* and *Saxifraga caespitosa*) invaded young deglaciated surfaces after about 5–6 years, and the density of plant cover (including a few species of flowering plants and lichens) increased during the first three decades after deglaciation. In the fourth decade, however, both the low number of plant species and the percentage of surface covered with vegetation stabilized at relatively low levels (Table 1). A similar limited plant succession was documented earlier by Piroznikow and Gorniak (1992) and more recently by Wojtun et al. (2007). Piroznikow and Gorniak (1992) concluded that the plant succession is controlled by low temperatures and unstable soil surface, but above all by climate dryness and the "low amount" of soil water available for plants on moraines. Plant succession in the foreland of the Werenskiold glacier, as evidenced by the percentage of the soil surface covered with vegetation, fits therefore to a logarithm model, where the "flattening" of the curve started at the same time as in the case of weathering evidence (Fe_o concentration, Fe_o:Fe_d ratio, chlorite:mica ratio etc.).

In addition to vascular plants, lichens and (in wet sites) mosses, and also cyanobacteria colonise the soil surface on recent moraines at the Werenskiold glacier, often being the first oxyphototrophic colonisers (Kastovska et al., 2005). Cyanobacteria and eukaryotic

microalgae play an indispensable role in soil stabilization against wind and water erosion, and accumulation of organic matter; however, their impact on mineral weathering and soil development in arctic environments is still not definitively established (Bernasconi and BigLink Consortium, 2008; Nakatsubo et al., 2005).

As plant succession proceeded, an increase of soil organic carbon and nitrogen was detected. It is worth noting that here the carbon accumulation process started not from zero content, but nearly 4–5 g kg^{-1}. Although the metamorphic rocks surrounding the Werenskiold glacier do not contain organic matter, it may be inherited from regoliths, marine sediments and other sediments enriched in the preglacial warmer period and then incorporated into the glacier body. Moreover, Kastovska et al. (2005) found that microbial communities can occur both in subglacial environments and on the glacier surface (in cryoconite sediments) and produce organic matter which is then added to glacial waters or sediments. Relic organic matter inherited from older deposits is common in profiles of soils developed from glacial materials in the Russian arctic and subarctic regions (Gubin and Veremeeva, 2010).

Already after 80 years of soil formation, the soil organic carbon and soil nitrogen stocks reached 0.76 kg m^{-2} and 0.085 kg m^{-2}, respectively, rather low when compared to other chronosequences in subarctic or alpine environments. He and Tang (2008) reported carbon accumulation reaching 3.5 kg m^{-2} after 180 years on moraines of the Hailogou Glacier (Sichuan, China), and Mavris et al. (2010) reported up to 5.5 kg C m^{-2} after 150 years in the Morteratsch proglacial area (Swiss Alps). Our findings are consistent only with other reports from the High Arctic, such as from the deglaciated area near Ny-Alesund on Spitsbergen (Nakatsubo et al., 2005). Increases in soil organic C and N concentration in the Werenskiold forefield showed a non-linear (polynomial) pattern over time in the uppermost layer 0–3 cm. Organic C and N stocks within the 15 cm soil layer fit similar polynomial or logarithmic models with the maximal rates of accumulation being in the first decades of soil development. It is consistent with a general opinion based on chronosequences of several thousands of years and ^{14}C dating that indicates the greatest rates of net carbon accumulation during the initial phase of soil formation (Schlesinger, 1990). The mean annual accumulation rates of organic C and N in the upper soil layer (0–15 cm) over the observed time span were 4.4 g m^{-2} a^{-1} and 1.1 g m^{-2} a^{-1}, respectively (Figs. 5b and 6b). These findings are lower than results from other short-term chronosequences. Mean annual rates of C and N accumulation at the Hailogou Glacier were 28 g m^{-2} a^{-1} and 3.5 g m^{-2} a^{-1}, respectively (He and Tang, 2008), and up to 36 g C m^{-2} a^{-1} at Morteratsch (Mavris et al., 2010). However, the rates of C and N accumulation in soils at the Werenskiold glacier are similar to those reported by Egli et al. (2001) from a soil chronosequence in the Swiss Alps that ranged from 6.7 to 9 g C m^{-2} a^{-1} and from 0.33 to 0.50 g N m^{-2} a^{-1}. This chronosequence covers a period of 400 years and involves the stage of soil "maturity" with a steady-state of organic carbon. Similarly, a quasi steady-state is observed in a chronosequence at the Werenskiold glacier, where the rate of carbon accumulation (in a layer 0–15 cm) decreased from 20 g m^{-2} a^{-1} (in the first decade after deglaciation) to only 1 g m^{-2} a^{-1} in the fifth and following decades (Fig. 5b). The varying length of a steady-state period may significantly influence the final result of calculations. It seems, therefore, that the accumulation ratios for the period of "increase" (better for comparison), and for the whole observed period of soil development (informative only) should be calculated separately. The increase over time in organic carbon stocks in soils of the Werenskiold chronosequence can be fitted to a logarithmic model, where the carbon accumulation is an ever-increasing function of time, however, where a quasi steady-state is approached after a certain period of time. Similar patterns are reported in other chronosequences (e.g. Egli et al., 2011; Lichter, 1998; Schlesinger, 1990), but He and Tang (2008) suggested that the logistic model often provides a better fit to observed phenomena. Based on such a model they estimated the time of soil organic carbon accumulation to a steady-state (under present climate conditions) to about 800–1000 years. Lichter (1998) calculated that at least 400 years may be required to attain steady-state on sand dunes. We may conclude, therefore, that despite the presently limited plant succession and reduced C stock rate the accumulation of organic carbon in soils at the Werenskiold glacier has not yet reached a steady-state. Available reference levels for the potential content of organic carbon in soils on recent moraines are offered by the buried soils on the glacier foreland or relic soils on uplifted marine terraces (Kabala and Zapart, 2009).

As plant cover extends on moraines, the organic acids from decaying plant residues cause a progressive decrease in the pH value of the soils. The rate of decrease, however, has slowed with time and finally stabilized. Similar trends for pH were also found in the Hailougou (He and Tang, 2008) and the Morteratsch (Mavris et al., 2010) chronosequences, but the steady-state was never reached as early as after four to seven decades. The pattern of limited decline in pH was closely related to limited accumulation of organic carbon, carbonate leaching and pedogenic Fe oxide formation indicating interdependences between climatically-controlled succession of vegetation, weathering intensity and intensity of soil development. Similarly, in the opinion of Jacobson and Birks (1980), low temperatures, a short growing season, and low mean annual precipitation limit plant growth and account for the delayed soil development on the Klutlan moraines in the Yukon Territory.

5. Conclusions

Soils in the proglacial area of the Werenskiold glacier have a very young age. However, within only 80 years of soil development transformation trends could be estimated. The development of soil properties shows mostly logarithmic or polynomial patterns with soil age over the observed time span. Immediately after deglaciation, transformation of chemical soil properties started at high rates; however, most of the observed indicators of weathering and soil development, including stocks of organic carbon and nitrogen, carbonate dissolution, pH value, chlorite:mica ratio in clay fraction, pedogenically-derived Fe etc. decreased with time. These results highlight the importance of the local climate and the intensity of plant succession on weathering processes and soil development. In contrast to the Low Arctic and alpine environments of Eurasia, the plant succession on proglacial areas in the High Arctic is inhibited by low precipitation and limited water availability for plants. The same factor limits the carbonate dissolution and leaching that stabilizes soil pH and inhibits primary mineral weathering. Based on observations of the initial stage of soil development in the foreland of the rapidly retreating Werenskiold glacier we would therefore conclude that climate warming, not accompanied by an increase in climate humidity, does not accelerate plant succession and soil development. Thus, the globally discussed potential for carbon and nitrogen fixation in arctic soils currently developing in the proglacial areas of retreating glaciers should not be overestimated in further models of carbon sequestration.

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