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HOLOCENE RUSTY AND RUSTY PODZOLIC SOILS IN THE TUNDRA AND TAIGA OF MIDDLE SWEDEN

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INTRODUCTION

In spite of ample literature, according to Tavernier and Smith [1957], as well as Arnold et al. [1990] both the origin and properties of part of soils developed in cold climates, referred to as brown earth or podbur and their transient forms to podzolic soils have not been so far explicated univocally. In more recent comparative publications [Konecka-Betley, Janowska 1996, Melke 1997], it is stated that chemical indicators in horizons B and others, which differentiate brown earths as well as rusty and podzolic soils in postglacial areas, as some authors emphasized [Mokma 1983, Prusinkiewicz, Bednarek 1985, Bednarek 1981], exhibit geographical differentiation and they may be unreliable as a criterion for type and subtype of soils.

In this publication a characterisation is presented of some chemical and physico-chemical properties of the catena of mountain soils: rusty, rusty podzolic and podzolic in the tundra and the northern taiga of Middle Sweden. At present these soils, in the conditions of subarctic and boreal climate are subject to intensive processes of podzolizing [Kowalkowski 1995a, b, Konecka-Betley, Janowska 1996]. They can be a good comparative material for soils of similar yet relic structures of profiles which occur on the areas of the Central European Lowlands within the subboreal climate. In the set of soil-formation factors, the changing climate gives soils in time and space some very characteristic permanent properties dependent on soil localisation in landscape and on its parent rock [Kowalkowski 1993].

OBJECT AND METHODS

The field investigation was carried out in 1990 in the slope catena of soils on NE and E postglacial and periglacial slopes of granite-gneiss pedestal of caledonian orogenesis in Middle Sweden. The flattened peak of the pedestal, reaching

1420, 1 m a.s.l, with slopes rising to 1000 m above the surrounding, has been smoothed by the processes of glacier detersion. The smooth proximal northern and north-eastern slope of this muton is incised by holocene cryoplanation systems of terraces of varying width.

The recent stages of subarctic and boreal vegetation as well as the structures of the investigated soil profiles were described in another paper [Kowalkowski 1995b]. A survey physiographic documentation of the investigated soils is shown in the Table 1.

The continental glacier of the last Scandinavian glaciation Salpausselkä receded from the discussed area some 9 000 [Lundquist 1986] to 6 000 [Hinneri 1974] years BP.

The development of the investigated soils could have thus been initiated in eo- to mezoholocene, from the boreal period to the third phase of the Atlantic period AT3 according to Starkel's chronology [1977]. In that period complexes of young

TABLE 1. Physiographic characteristics of soils investigated on the NE and E slopes of the Åreskutan massif

Pit No	Eleva- tion [a.s.l. m]	Climatic-plant stage	Soil type and kind
1	1180–1190	nival, mountain lichen-mosses tundra	peaty gleypodzol, cryogenic, thixotropic of tundra, developed from holocene weathering
2	850–860	subnival, mountain shrubby tundra	rusty humus-iron podzolic soil developed from holocene periglacial slope covers, on 60 cm depth glacial cryogenic weathering waste
3	680	alpine-subalpine spruce-birch tundra-taiga	gley podzolic soil developed from holocene alluvial deposits
4	580	subalpine northern spruce taiga	rusty pseudogley podzolic soil developed from holocene periglacial slope covers from 50 cm underlying glacial covers
5	520	subalpine spruce northern taiga	rusty pseudogley podzolic soil developed from holocene periglacial slope covers on 40 cm underlying glacial slope covers
6	410	subalpine spruce middle taiga	rusty podzolic soil developed from holocene glacial morainic slope covers
7	420	subalpine spruce middle taiga	rusty pseudogley podzolic soil developed from holocene glacial slope covers overlying compact solifluction cover; thickness of the deluvial Bv-horizon to several hundred of cm

holocene podzolic soils, middle-old holocene rusty podzolic soils and old holocene rusty soils on the different parts of the slope were formed.

In the laboratory of the Department of Soil Geography and Environment Protection, High School of Education in Kielce, an analysis of grain structure of fraction > 0.1 mm was performed by the method of water sieving, fraction < 0.1 mm by Casagrande's areometric method modified by Prószyński. From the quotient of the sums of dust fractions 0.10–0.05 mm, and sand fractions 1.0–0.10 mm the SP index was calculated, assuming that in periglacial environ- ment *in situ* fractions < 0.10 mm are formed as a result of cryohydrothermic processes of desquamation and fractions 0.10–1.00 mm due to granular frost disintegration.

The exchangeable cation content Ca^{2+}, Mg^{2+}, K^{+}, Na^{+} is denoted in extract 1 M CH_3COONH_4 by pH 7. Exchangeable H^{+} is denoted by the Kappen' method, exchangeable Al^{3+} – by the Sokolov method, and exchangeable Fe^{3+} – by the dipiridil method. $C_{org.}$ is denoted by the Tiurin's method, and $N_{org.}$ – by the Kjeldahl's method. The acidity is measured by the potentiometric method using glass and calomel electrodes.

THE CURRENT STATE OF CHEMICAL AND PHYSICO-CHEMICAL PROPERTIES

An important role in shaping the soil cover in mountain areas is played by pedo- and lithostratigraphic equilibrum which consists in permanent interference dependent on the climatic and vegetation tiers of the slope pedomorphogenesis. Soils on slopes can be developed only when morphogenetic processes are not too active, so that a minimum degree of topostability is obtained [Tricart 1965, Kowalkowski 1988, Coutard 1989]. On NE slope of Åreskutan exists holocene sequence of slope glacial and glacial-periglacial covers transformed pedogenically in various degrees [Kowalkowski 1995b].

Parent rocks of soils

The morphology of clasts and small grained waste, their grain size and stratigraphy are indicators of the development conditions of parent rock of soils. The investigated soils were formed from stratified granite-gneiss glacial and periglacial waste of very differentiated grain size (Table 2), in the investigated profiles and on different elements of slope [Kowalkowski 1995b]. A similar differentiation of grain size was found by Hinneri [1974] in the holocene tundra soils in Finland.

The stratification and morphology of waste within the profile of the investigated soils points to a relatively young age of their formation. In the lower parts of the soils numbers 3 and 4, from the depth of 50 cm, in the pit number 5 from 30 cm, and in the pits numbers 6 and 7 from the surface the waste contain clasts of different size with rounded edges, with longitudinal axes laid parallelly to their stratum roof. These are indicators of the glacial environment of their formation, and congelifluctional transport on slope. Only in pit No 2, beginning with the depth of 60 cm laminar stratified deposits of frost granular disintegration occur in which over 70% are sharp-edged sand grains, mainly fine sand of grain size 0.25–0.10 mm (Table 2). In pits 2, 4, 5 younger series of congelifluctional periglacial holocene wastes

TABLE 2. Distribution of grain of different size in the soil of catena

Soil pit No	Horizon	Depth of sampling [cm]	Percentage of skeletal parts [mm]			of earth particles [mm]							SP index
			>10	10–2	2–1	1–0.5	0.5–0.25	0.25–0.10	0.10–0.05	0.05–0.02	0.02–0.005	<0.005	
1	Egg	20–30	2.5	13.3	5.4	9.7	10.1	22.8	13.9	11.0	5.5	5.5	0.71
	Bhfegg	35–40	8.4	14.9	5.5	10.0	10.0	21.9	9.6	10.0	6.5	3.8	0.62
2	E	10–20	11.2	11.6	4.4	12.6	8.1	21.4	10.3	7.3	5.8	7.2	0.56
	Bhfe	20–25	4.3	10.7	6.0	21.2	12.9	21.0	15.1	3.4	0	5.5	0.34
	BvBfe	45–50	12.5	17.0	10.0	15.1	13.9	18.0	9.8	3.6	0	0	0.29
	DBv	75–80	0	1.1	8.0	25.2	20.8	31.4	7.2	5.5	0.9	0	0.18
3	Eg	10–15	1.9	15.9	6.1	13.7	6.7	18.2	12.9	13.0	7.6	4.6	0.67
	Bhfeg	20–25	15.6	13.8	5.5	13.8	6.1	18.9	11.0	8.4	5.2	1.3	0.63
	Dgg	50–55	23.8	10.3	2.7	7.9	6.1	17.3	10.3	8.4	7.6	5.6	0.84
4	E	4–9	5.8	9.8	2.2	4.0	6.6	27.9	23.0	9.4	8.3	3.2	1.06
	Bhfe	9–11	4.3	9.1	6.0	6.0	6.4	26.9	22.3	9.9	5.0	6.7	0.95
	Bv	15–20	11.8	9.1	3.0	8.5	7.9	25.0	15.5	9.9	7.7	1.4	0.80
	CBv	45–50	11.4	14.1	4.2	8.3	6.9	25.8	9.6	11.2	7.0	1.5	0.68
	sol Dg	70–75	1.5	8.0	3.9	16.5	14.9	30.0	11.5	9.5	3.5	0.8	0.40
5	E	10–12	1.8	8.2	2.8	5.7	4.7	23.9	22.3	9.6	10.5	10.4	1.24
	Bhfe	15–20	9.9	6.2	2.5	13.8	9.5	26.6	12.5	10.6	7.3	1.6	0.61
	Bv	25–30	0.8	9.8	3.3	10.4	6.2	23.1	12.0	12.9	14.7	16.8	1.00
	sol Bvg	70–75	6.6	9.6	2.6	8.0	5.8	19.7	8.6	13.0	13.8	12.1	1.06
6	AE	5–10	2.8	0.2	0.1	9.2	11.5	32.2	23.6	9.7	4.8	5.8	0.72
	BvBhfe	12–20	7.2	1.7	0.1	10.9	13.1	35.5	21.5	5.5	2.7	1.8	0.50
	Bv	40–50	0.1	0	0	13.2	14.0	39.5	22.3	8.0	3.0	0	0.50
	C1	10–20	1.8	3.5	3.9	8.9	8.4	32.2	20.4	10.9	8.1	1.8	0.79
7	AhBv	0–5	3.6	11.4	6.5	9.5	8.1	18.3	16.7	10.2	10.2	5.5	1.03
	Bv	5–10	30.6	6.2	1.8	9.5	7.1	15.2	8.1	6.1	6.7	8.6	0.65
	Bv	20–30	6.5	9.8	1.6	11.8	11.1	25.9	9.5	6.6	7.4	9.9	0.48
	Cg	40–50	3.7	8.0	3.6	8.4	6.8	16.6	13.1	9.4	22.9	7.6	1.43

occur on the surface in which clasts of diameter > 10 mm have sharp edges. Holocene fluvial sediments with sporadic stones of rounded edges occur in pit 3 situated on a cryoplanation terrace. Furthermore, a thin layer of sharp-edges waste on the base of solid rocks in pit 1 is of young holocene origin.

The silt fraction content of sizes from 0.10 to 0.005 mm, originated in the process of cryogenic scalling of sand grain surface in glacial and periglacial environment [Kowalkowski, Kocoń 1998], is as a rule the highest in upper horizons Ah and E, as well as in parent rock Cg and in the subsoil Dg. It ranges from 23.4 to 45.4%, on average it is 35% (Table 2). The content of these fractions in cryogenic horizons Bv is also very high, ranging from 20.9 to 39.6%, on average 35%. The lowest content of silt fraction was found in illuvial horizons Bhfe and BvBhfe, under the elluvial ones. The content of these fractions is from 13.4 to 30.4%, on average 27.7%, which can be the result of cementing of part of the silt grains into coarser aggregates by deposited Si-Al-Fe compounds with humus.

Therefore, it is not possible to state profile postsedimentional pedogenic differentiation of silt fractions in their soil section, although the content of these fractions goes towards the slope foot.

An important component of parent rock is sharp-edged sand of diameter from 1.0 to 0.10 mm. Its content in the investigated slope deposits is from 31.3 to 61.4%, on average 41.8%. There is a distinct domination of fine sand, from 15.2 to 39.5%, of grain diameters 0.25–0.10 mm. The content of sand fraction due to frost granular disintegration of solid rocks and their splinters generally decreases with the profile depth and also decreases towards the slope foot, i.e. in older slope deposits and wastes. The weathering indicator SP of rock material is as a rule higher than 0.50, and it reaches 1.43 (Table 2), which can lead to a conclusion that there are constant hydrothermic conditions of frost granular weathering and scalling of waste in the area under discussion, hence domination of a more wet and cold climate.

The content of fraction <0.005 mm is very differentiated but low and it does not reveal any distinct dependence with the horizontal structure of the investigated soils. Only in horizons A usually the content of these fractions is higher than in horizons Bhfe and BvBhfe, which also found by Targuljan [1971] in podzolic soils of Northern Eurasia. This fact and also the general content of silt and sand fractions cause the soils formed from the discussed parent rocks to be easily permeable, which was also demonstrated by Hinneri et al. [1975] in the soils of Scandinavian subalpine ecosystems.

The excess of amounts rainfall and melting waters of perennial snow patches in shaped relief bends in the tundra stage [Kowalkowski 1995b] can be accumulated for a long time on the solid plates of subsoil rocks within active permafrost in the thixotropic horizon. Periodically, these waters are amassed over the illuvial horizons of podzolic soils from fragmentarily formed and hardly permeable fragipans. According to Langohr and Vermeire [1982], a periodically emerging water level over the fragipan ceiling may participate in gley bleaching of the lower part of horizon Eg.

Textural discontinuities in the investigated soils, characteristic of Arctic soils [Stoner et al. 1983] are factors that regulate water migration in soil profile.

ORGANIC MATTER, C:N RATIOS AND ACIDITY

The organic matter content in all horizons of the investigated soils, including horizons C and D is relatively high. In the ceilings of Ofh horizons it is 47.6–67.8%, and in floors it ranges from 32.6 to 39.2%. The organic matter content in elluvial horizons ranges widely from 1.3 to 6.7%, depending on humidity and water permeability of soils, as well as on the climatic and vegetational stages. In those horizons, the highest concentrations of organic matter are found in wet tundra. Its mass decreases together with the decreasing humidity of the investigated soils, alongside slope lowering [Kowalkowski 1995b], to the stage of middle taiga.

The pedogenic migration of organic matter in the investigated soils is testified by the second maximum of its accumulation in the illuvial horizons Bhfe and in illuvia superimposed on the ceiling part of horizon Bv which are from 11.4 to 2.4% depending on soil humidity, and by its smaller amounts in rusty horizons Bv and CBv, from 4.6 to 1.0%. Also in horizons C, Cg and Dg, the organic matter content is high, from 0.35 to 4.30%, which points to its intensive, deep and spatially differentiated horizontal and vertical migration with rainfall and melted snow waters.

There are known reports about the high humus content along the full depth of profile in cryogenic podzolic soils in Central and Northern Scandinavia [Schlichting 1963, Hinneri et al. 1975], in the podzolic soils of the subantarctic forest, subantarctic tundra, and Antarctic polar desert [Smith 1990, Blume et al. 1996], in podzolic soils of cold and humid areas of Eurasia and Northern America [Targuljan 1971], in gley-podzolic soils of the mountain tundra of Ireland [Wilson, Sellier 1995] or the high-mountain continental tundra in Changai Mountain [Kowalkowski, Starkel 1984]. In the above-mentioned and other cold areas there occurs accumulative distribution of organic matter in the profiles of podzolic and rusty podzolic soils not only due to its migration with water solutions, but also thanks to cryogenic homogenisation in the form of little changed organic residuals of overground and underground parts of plants with cell structure frequently well preserved.

This is confirmed in the investigated soils by the C : N ratios which ranges from 15.0 to 32.5 : 1; it is wider in more humid horizons and narrower in drier horizons (Table 3), as well as soil acidity pH_{KCl} which ranges from 3.00 to 5.33 (Table 5). Soil acidity in the investigated catena decreases with decreasing soil humidity from pH_{KCl} 3–4 in gley-podzol of the tundra to pH_{KCl} 4–5 in rusty podzolic soil and rusty soils of the middle taiga.

SORPTION PROPERTIES AND EXCHANGEABLE CATION CONTENT

The sorption capacity along the profiles in the investigated soils is relatively high (Table 4, Figs 1 and 2). Its highest values, on average 32.53 me/100 g (from 26.10 to 51.83 me/100 g) were found in organic material of horizons Ofh, although it points to a low degree of humification of organic deposits in this horizon and to a small content of organic colloids in it. In horizons E and Egg the sorption capacity

TABLE 3. Some chemical and physico-chemical properties in genetic horizons of the soil catena

Soil Pit No	Horizon	Depth of sampling [cm]	Organic matter C 1.724	$N_{org.}$ [%]	C:N	Exchangeable cations [me/100 g] Ca+Mg +K+Na	Exchangeable cations [me/100 g] H+ Al+Fe	Alcalic e.c / acidic e.c.	V [%]	$\frac{Ca^{2+}+Mg^{2+}}{K^{+}+Na^{+}}$	$\frac{Ca^{2+}}{Mg^{2+}}$	$\frac{Ca^{2+}}{K^{+}}$
1	Tgg	0–10	46.03	1.05	25.4	6.81	29.36	0.28	18.8	0.94	0.20	0.42
	Egg	20–30	6.70	0.20	19.5	2.85	24.69	0.12	10.3	2.63	0.09	0.41
	Bhfegg	35–40	5.00	0.13	22.3	3.69	10.78	0.34	25.5	4.77	1.10	4.44
2	Ofh	0–5	62.91	1.25	29.2	17.01	13.19	1.29	56.3	2.35	0.06	0.16
	AhE	5–10	32.62	0.67	28.2	5.08	21.05	0.24	19.4	2.19	0.05	0.15
	E	10–20	4.68	0.09	30.2	1.68	19.08	0.09	8.1	1.85	0.42	0.74
	Bhfe	20–25	11.36	0.22	30.0	2.62	35.86	0.07	6.8	3.34	0.81	2.57
	BvBfe	45–50	3.64	0.07	30.1	2.59	27.46	0.09	8.6	7.69	5.19	6.43
	DBv	75–80	0.97	0.02	28.0	1.72	24.63	0.07	6.5	12.23	4.13	11.64
3	Ofh	0–5	61.43	1.15	31.0	12.54	13.56	0.92	48.0	2.24	0.04	0.11
	Ofh	5–10	66.63	1.69	22.9	7.56	27.01	0.27	21.9	1.87	0.15	0.33
	Eg	10–15	1.31	0.03	25.3	1.78	20.64	0.09	7.9	2.49	0.65	1.32
	Bfegg	20–25	7.21	0.09	46.4	2.53	24.07	0.11	9.5	6.03	1.36	4.31
	Dgg	50–55	4.31	0.06	41.7	1.12	24.90	0.04	4.3	4.33	0.64	2.38
4	Ofh	0–3	47.65	0.88	31.4	14.84	25.49	0.58	36.8	4.82	0.06	0.27
	E	4–9	1.60	0.08	11.6	5.22	19.13	0.27	21.4	5.69	0.39	2.45
	Bhfe	9–11	3.69	0.07	30.6	3.27	28.42	0.12	10.3	5.54	0.29	1.68
	Bv	15–20	3.25	0.09	22.6	2.98	27.43	0.11	9.8	5.33	1.68	4.21
	CBv	45–50	1.12	0.04	16.2	2.53	24.18	0.10	9.5	13.05	1.30	8.87
	sol Dg	70–75	0.35	0.01	20.0	6.33	27.39	0.23	18.9	30.65	3.06	33.00
5	Ofh	0–5	67.82	1.21	32.5	17.48	34.35	0.49	33.7	5.64	0.95	2.99
	OfhAh	5–10	39.17	1.20	18.9	14.97	30.74	0.49	32.8	5.87	0.24	1.52
	E	10–12	2.09	0.06	20.7	3.15	16.91	0.19	15.7	10.74	0.51	11.88
	Bhfe	15–20	5.74	0.12	27.8	1.77	35.86	0.05	4.7	11.64	0.65	7.11
	Bv	25–30	2.69	0.08	19.5	1.67	29.75	0.06	5.3	14,18	1.36	11.25
	solBvg	70–75	0.67	0.02	19.5	1.18	21.36	0.05	5.2	12.11	0.85	16.67
6	AhE	5–10	4.72	0.18	26.8	12.65	13.11*	.	.	8.03	2.59	9.22
	BvBhfe	12–20	2.41	0.12	20.4	11.86	5.12*	.		33.56	1.48	34.59
	Bv	40–50	0.98	0.06	17.2	1.31	4.69*	.		17.71	2.43	37.00
	C	10–20	0.80	0.04	19.5	3.37	3.11*	.	.	13.04	2.77	10.95
7	AhBv	0–5	9.19	0.42	21.7	6.36	17.70*	.	.	47.92	1.99	37.72
	Bv	5–10	4.57	0.28	16.2	2.59	8.40*	.		24.90	0.49	10.25
	Bv	20–30	2.36	0.13	18.5	0.85	1.97*	.		11.14	0.56	5.60
	Cg	40–50	0.34	0.01	24.0	1.00	0.66*	.		8.09	0.78	4.87

* – $H^{+}+Al^{3+}$

is considerably lower, on average it is 23.6 me/100 g and varies a little from 20.60 to 27.54 me/100 g. In horizons Bhfe and BvBhfe it grows to an average 32.89 me/100 g with fluctuation from 26.60 me/100 g in the dry rusty podzolic soil in the taiga to 38.48 me/100 g in the wet tundra rusty podzolic soil. In horizons Bv and CBv the sorption capacity is relatively high, on average 27.54 me/100 g and

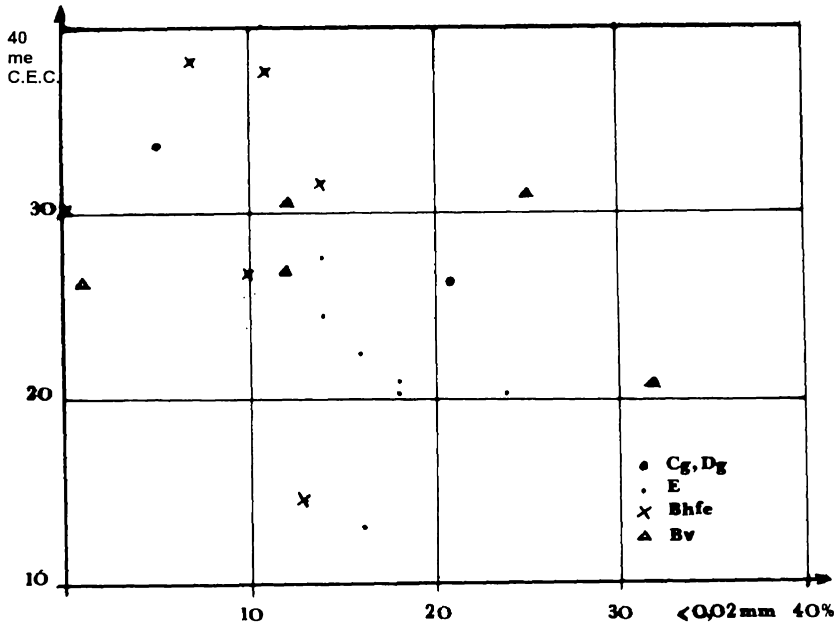

FIGURE 1. Relationship between floatable parts (< 0.02 mm) and the exchangeable cation content in the soil horizons

it increases on average to 29.89 me/100 g in the materials of subsoil and parent rock of glacigen origin.

The fraction content <0.02 mm, with the exception of Bv horizons, which are under the horizon Bhfe (Fig. 1) does not influence the value of the sorption capacity. However, there is a distinct dependence between sorption capacity and the organic matter content, which is particularly marked in horizons AhE that have the character of organic overlay and in the horizons Ofh (Table 2). The main factor of high and little differentiated values of sorption capacity is strong weathered surface of mineral grains [Kowalkowski, Kocoń 1998], due to which their ion exchange surface along the whole depth of soil profile, including horizons C and D, increases many times.

Cations Al^{3+} + Fe^{3+} + H^{+} of acidic character dominate over cations Ca^{2+} + + Mg^{2+} + K^{+} + Na^{+} of basic character. This prevalence, which is in horizon Ofh 66.5% (it ranges from 43.5 to 81.2%), increases in the successive horizons laid lower – E and Egg – to mean 87.3%, in Bhfe and BvBhfe to 89.1%, in Bv and CBv to 92.5% and in horizons Cg and Dg it is 90.6%. The expression of this situation are the ratios of both groups of exchangeable cations which decrease from 1.29 to 0.28% in upper horizons Ofh to 0.27–0.09% in horizons E and to 0.24–0.04 in horizons Cg and Dg at depths 50–80 cm (Table 3). With the decreasing absolute heights of slope surface in the investigated soils in all horizons, the prevalence of acidic cations increases several times in relation to basic cations.

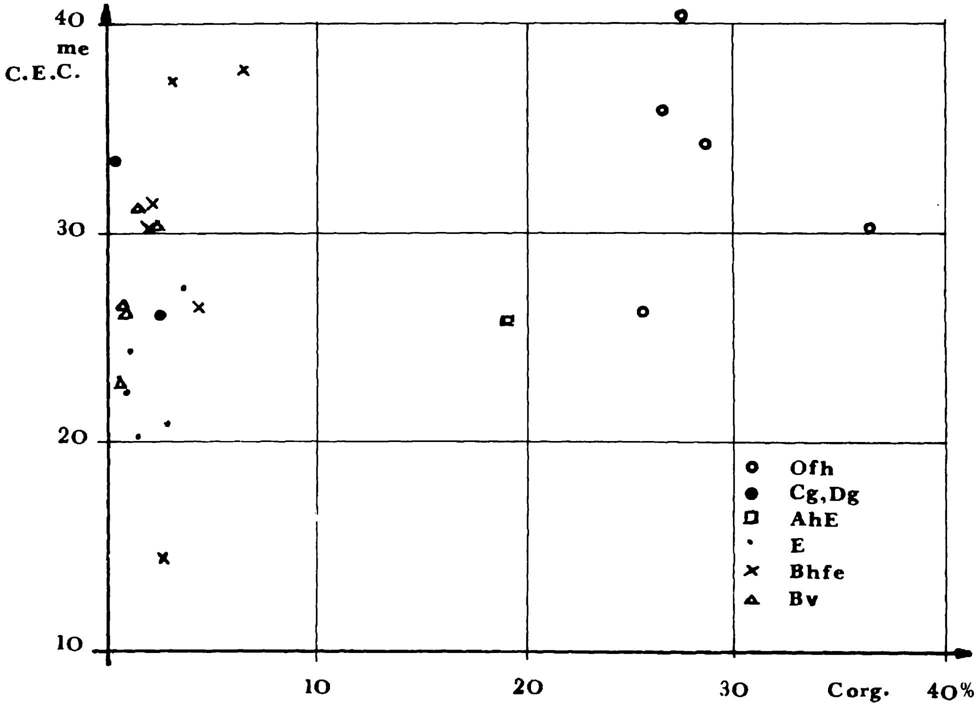

FIGURE 2. Relationship between the organic matter and the exchangeable cation contents in the soil horizons

One can observe low concentrations of exchangeable H^+ which reach 0.88-4.55 me/100 g in organic horizons. However, they decrease rapidly in mineral horizons from 0.53 to 0.09 me/100 g of soil material. In quantitative ratios of the investigated exchangeable cations H^+ occupies 3–5th place in horizons Ofh, in horizons E and Egg the 4–7th place, in horizons Bhfeg and BvBhfe the 6–7th place (Table 5). It follows from the quantitative arrangement of the investigated cations in Table 5 that in horizons Ofh, beside dominant Fe^{3+}, there occurs Mg^{2+} in the second place, and in horizons E and Egg exchangeable cations Mg^{2+} occupies the third place after dominant Fe^{3+} and Al^{3+}. In horizons Bhfeg and BvBhfe dominate cations Fe^{3+} and Al^{3+}, as a rule Ca^{2+} and Mg^{2+} occupy the third and fourth places. In horizons Bv and CBv dominate Fe^{3+} with Ca^{2+}, Al^{3+} and Mg^{3+} that occur alternatively in the 2–4th places.

Low concentrations of exchangeable H^+ in comparison with Fe^{3+} and Al^{3+}, also known from currently developing arctic and subarctic soils of Eurasia [Targuljan 1971], can be caused by use of protons H^+ during a relatively intensive cryohydrothermic weathering of silicate rocks and minerals. This is undoubtedly permitted by low pH, presence of aggressive humic acids in soil solutions, and an ever increasing ion exchangeable surface of cryogenic origin [Kowalkowski, Kocoń 1998].

TABLE 4. Profile distribution of exchangeable cations in the soil catena

Soil pit No	Soil horizon	Depth of sampling [cm]	Exchangeable cations [me/100 g]							T [me/ /100 g]
			Ca^{2+}	Mg^{2+}	K^{+}	Na^{+}	H^{+}	Al^{3+}	Fe^{3+}	
1	Tgg	0–10	0.65	3.29	1.34	1.53	0.88	11.37	17.11	36.17
	Egg	20–30	0.17	1.97	0.30	0.41	0.18	4.46	20.05	27.54
	Bhfegg	35–40	1.60	1.45	0.28	0.36	0.09	3.06	7.63	14.47
2	Ofh	0–5	0.65	11.28	1,09	3.99	4.20	4.20	4.79	30.20
	AhE	5–10	0.17	3.32	0.43	1.16	0.88	7.87	12.30	26.13
	E	10–20	0.32	0.77	0.16	0.43	0.70	7.70	10.68	20.76
	Bhfe	20–25	0.95	1.17	0.13	0.37	0.35	8.15	27.36	38.48
	BvBhfe	45–50	1.87	0.36	0.07	0.29	0.18	1.92	25.36	30.05
	DBv	75–80	1.28	0.31	0.02	0.11	0.18	0.87	23.58	26.35
3	Ofh	0–5	0.35	8.32	0.65	3.22	4.55	3.85	5.16	26.10
	Ofh	5–10	0.65	4.21	0.73	1.97	1.58	9.95	15.48	34.57
	Eg	10–15	0.50	0.77	0.13	0.38	0.18	1.05	19.41	22.42
	Bfegg	20–25	1.25	0.92	0.07	0.28	0.18	3.32	20.57	26.60
	Dgg	50–55	0.38	0.53	0.05	0.16	0.18	7.34	17.43	26.07
4	Ofh	0–3	0.65	11.64	0.17	2.38	1.75	2.10	21.64	40.33
	E	4–9	1.25	3.19	0.27	0.51	0.53	4.19	14.41	24.35
	Bhfe	9–11	0.62	2.15	0.13	0.37	0.35	3.24	24.83	31.69
	Bv	15–20	1.60	0.95	0.09	0.34	0.18	0.96	26.29	30.41
	CBv	45–50	1.33	1.02	0.03	0.15	0.09	0.17	23.92	26.71
	sol Dg	70–75	4.62	1.51	0.06	0.14	0.09	0.09	27.21	33.72
5	Ofh	0–5	1.15	12.14	0.35	3.84	4.55	1.75	28.05	51.83
	OAh	5–10	2.49	10.30	0.54	1.64	2.45	2.97	25.32	45.71
	E	10–12	0.95	1.97	0.15	0.08	0.26	7.62	9.03	20.06
	Bhfe	15–20	0.69	0.99	0.05	0.09	0.70	8.92	26.24	37.63
	Bv	25–30	0.90	0.66	0.03	0.08	0.18	3.09	26.48	31.42
	sol Bvg	70–75	0.50	0.59	0.06	0.03	0.09	0.52	21.02	22.81
6	AhE	5–10	8.12	3.13	0.88	0.52	2.19	.	.	.
	BvBhfe	12–20	0.74	0.50	0.02	0.05	3.28			
	Bv	40–50	7.61	3.13	0.22	0.10	2.85			
	C	10–20	2.30	0.83	0.21	0.03	2.19	.	.	.
7	AhBv	0–5	14.15	2.08	0.11	0.02	6.78	.	.	.
	Bv	5–10	0.82	1.67	0.08	0.02	6.56			
	Bv	20–30	0.28	0.50	0.05	0.02	1.97			
	Cg	40–50	0.39	0.50	0.08	0.03	0.66	.	.	.

The actual course of this process can be inferred on the basis of relatively high concentrations of exchangeable Mg^{2+} from 0.6 to 2.1 me/100 g in mineral soil horizons, particularly in horizons E. Also in horizons Ofh high concentration of exchangeable Mg^{2+} occurs from 3.3 to 12.6 me/100 g which comes from biogenic accumulation and the second or first place in the quantitative series of exchangeable cations before or after Fe^{3+} (Table 5). In horizons E and Egg, Mg^{2+} occupies the third place, in horizons Bhfeg and BvBhfe – the second or third places, in horizons Bv and CBv – the second to the fourth places, and in horizons Cg and Dg – the third place (Table 5).

TABLE 5. Quantitative arrangement of exchangeable cations in soil horizons

Soil pit No	Soil horizon	Cation arrangement	pH	
			H_2O	KCl
1	Tgg	$Fe^{3+} > Al^{3+} > Mg^{2+} > K^{+} > Na^{+} > H^{+} > Ca^{2+}$	3,56	3,26
2	Ofh	$Mg^{2+} > Fe^{3+} > Al^{3+} = H^{+} > K^{+} > Na^{+} > Ca^{2+}$	4,08	3,03
3		$Mg^{2+} > Fe^{3+} > H^{+} > Al^{3+} > K^{+} > Na^{+} > Ca^{2+}$	3,99	3,00
3		$Fe^{3+} > Al^{3+} > Mg^{2+} > K^{+} > H^{+} > Na^{+} > Ca^{2+}$	4,29	3,50
4		$Fe^{3+} > Mg^{2+} > K^{+} > Al^{3+} > H^{+} > Ca^{2+} > Na^{+}$	4,47	3,71
5		$Fe^{3+} > Mg^{2+} > H^{+} > K^{+} > Al^{3+} > Ca^{2+} > Na^{+}$	6,08	5,26
5		$Fe^{3+} > Mg^{2+} > Al^{3+} > Ca^{2+} > H^{+} > K^{+} > Na^{+}$	4,04	3,10
1	Eg, E	$Fe^{3+} > Al^{3+} > Mg^{2+} > K^{+} > Na^{+} > Ca^{2+} > H^{+}$	3,88	3,04
2		$Fe^{3+} > Al^{3+} > Mg^{2+} > K^{+} > H^{+} > Na^{+} > Ca^{2+}$	3,58	2,73
2		$Fe^{3+} > Al^{3+} > Mg^{2+} > H^{+} > K^{+} > Ca^{2+} > Na^{+}$	3,92	3,50
3		$Fe^{3+} > Al^{3+} > Mg^{2+} > K^{+} > Ca^{2+} > H^{+} > Na^{+}$	4,65	4,12
4		$Fe^{3+} > Al^{3+} > Mg^{2+} > Ca^{2+} > H^{+} > K^{+} > Na^{+}$	4,51	3,60
5		$Fe^{3+} > Al^{3+} > Mg^{2+} > Ca^{2+} > H^{+} > Na^{+} > K^{+}$	4,27	3,19
1	Bhfeg, BvBhfe	$Fe^{3+} > Al^{3+} > Ca^{2+} > Mg^{2+} > K^{+} > Na^{+} > H^{+}$	4,14	3,78
2		$Fe^{3+} > Al^{3+} > Mg^{2+} > Ca^{2+} > K^{+} > H^{+} > Na^{+}$	4,58	4,23
2		$Fe^{3+} > Al^{3+} > Ca^{2+} > Mg^{2+} > K^{+} > H^{+} > Na^{+}$	4,94	4,70
3		$Fe^{3+} > Al^{3+} > Ca^{2+} > Mg^{2+} > K^{+} > H^{+} > Na^{+}$	4,83	4,54
4		$Fe^{3+} > Al^{3+} > Mg^{2+} > Ca^{2+} > K^{+} > H^{+} > Na^{+}$	4,54	4,17
5		$Fe^{3+} > Al^{3+} > Mg^{2+} > H^{+} > Ca^{2+} > K^{+} > Na^{+}$	4,83	4,50
2	Bv	$Fe^{3+} > Ca^{2+} > Al^{3+} > Mg^{2+} > H^{+} > K^{+} > Na^{+}$	5,81	5,33
4		$Fe^{3+} > Ca^{2+} > Al^{3+} > Mg^{2+} > K^{+} > H^{+} > Na^{+}$	5,01	4,70
5		$Fe^{3+} > Al^{3+} > Ca^{2+} > Mg^{2+} > H^{+} > K^{+} > Na^{+}$	5,96	4,92
4	CBv	$Fe^{3+} > Ca^{2+} > Mg^{2+} > Al^{3+} > K^{+} > K^{+} > Na^{+}$	5,64	4,78
5		$Fe^{3+} > Mg^{2+} > Al^{3+} > Ca^{2+} > H^{+} > Na^{+} > K^{+}$	5,56	4.45
3	Cg, Dg	$Fe^{3+} > Al^{3+} > Mg^{2+} > Ca^{2+} > K^{+} > H^{+} > Na^{+}$	4,30	4,04
4		$Fe^{3+} > Ca^{2+} > Mg^{2+} > K^{+} > Al^{3+} = H^{+} > Na^{+}$	6,13	5,23

Release of Mg as well as Al and Fe in the environment of pH_{KCl} 2.7–4.1 in the horizons E of the investigated soils from weathering waste is dependent not only on the influx of mobile organic acids of the fulvic acid type from horizons Ofh under the tundra-taiga and northern taiga vegetation cover. Processes of frost weathering of primary silica minerals are an important factor which stimulates this process in time not only in horizons E but also in deeper horizons. These processes are limited not only to the external surfaces of grains and rock splinters, but they are developed along the walls of the nets of cryohydrothermic microcracks and gaps, which reach their internal surfaces [Kowalkowski, Brogowski 1983; Kowalkowski, Starkel 1984; Kowalkowski, Kocoń 1991, 1998]. In equivalent utilization of H^+ protons, lamelles of tiny mineral chips of diameters < 0.02 mm originating during scalling process are chemically decomposed especially rapidly in those conditions. The lower pH values fall, the higher amounts of H^+ are used to release ions of Al^{3+} which easily migrates with waters in the soil profile.

The rate of cryohydrothermic weathering is limited, however, by deposition of thin aluminium-silica crusts on grain surfaces, which is marked distinctly in horizons Bhfeg and BvBhfe of the investigated soils [Kowalkowski, Kocoń 1998]. Low concentrations of exchangeable H^+ in all the investigated soils [Tables 4 and 5] under domination of exchangeable Al^{3+} and Fe^{3+} and considerable concentrations in aqueous solutions of organic acids, which are a protective colloid for stimulation of migration ability of those cations in solutes with pH_{KCl} 3.9–4.9, show that along the whole depth of the investigated soils the cryohydrothermic processes of weathering are intensive.

Certain buffering to those processes is manifested by horizon Bv developed in the earlier stage of incisions of the investigated soils in the conditions of a more continental climate with low rainfall. In recent conditions permitting the podzolizing processes in Central Sweden poorly developed Bhs horizons [Kowalkowski 1995a, b] have small but growing in time ability to neutralise leached substances, and acid compounds and the possibilities of originate of clay minerals are small (Table 2). In the phase of intensive podzolizing and related ion migration as well as the displacement of organic matter, increasing with depth, through all soil horizons, it is not possible to assume that there is a relative stability in any part of profiles of the investigated soils. What has formed there is association of the mezo- and microcompounds, which are in different concentration state, in time in unstable cryohydrochemical equilibria.

DISCUSSION

In the light of the presented data about currently developing podzolic and rusty podzolic soil in the Åreskutan massif and the cited literature, it is only possible to accept with reservation the statements repeatedly present in the literature that paleosoils defined as soils formed in old landscapes are a perfect record of the bygone conditions of their environment [Arnold et al. 1990]. The reservation concerns first of all the contemporary processes in such a currently functioning soil. In the past bioclimatic conditions, particularly in subarctic and boreal environment, the soil development had a different character in relation to the contemporary state. It is thus possible to assume that the old-holocene (recent) profiles of cryogenic rusty soil in the considered holocene soils has been preserved from the old bioclimatic conditions. It has become a kind of scaffolding-environment for young-holocene (recent) rebuilding processes of podzolizing, successing in consequence of the changes of climate and vegetation [Kowalkowski 1993].

It seems that Stremme [1926] was correct when he stated that the „development of forest soils rusty coloured, frequently with podzolization features near the surface, with horizon B of illuvial features, depends rather on climate than flora". According to Ramann [1911] part of brown soils formed from loams and sands on the postglacial area is „the outcome of geological youth of glacial soils". He stated that those soils are connected with low annual temperatures, cold winters, and medium strong leaching, with dominant chemical weathering. These soils often have a little volume of bleached horizon which is the indicator of the rebuilding of the decaying brown soil from surface into young podzolic soils due to climatic change. Those soils with morphological features and structures originating in glacial and periglacial environment in profile [Kopp 1965, 1970; Kowalkowski 1988, Konecka-Betley 1982, 1991] are currently referred to as rusty and rusty podzolic soils. In the Systematics of Poland's soils [1989] they were incorrectly limited to sand substrates and included in the division of podzol-earths. Those soils are also formed from loamy substrates, and podzolizing of those soils in a younger successive process whose effects are superimposed on the original features of the older rusting process.

CONCLUSIONS

The previously cited and discussed factographic data allow to formulate the following conclusions:

1. The investigated peaty gleypodzol and rusty podzolic soils of the tundra and the northern taiga are at present in the phase of intensive processes to set going aggressive organic acids in horizons Ofh, which contribute in horizons E to the pedochemical decomposition of the fresh cryogenic surface of minerals and rock chips.
2. The investigated soils due to prevalence in the sorption complex of cations Al^{3+} and Fe^{3+} as well as elluvial and cryoturbation conditioned migration of organic matter can be referred to, after Targuljan [1971], as rusty podzolized Al-Fe-humous or peaty podzolic Al-Fe-humous soils.
3. The features of the recent process of podzolizing with eluviation and illuviation along the whole depth of profile are superimposed on stable features of pedocryogenic rusty horizons Bv, homogeneous with regard to colour and grain size.
4. Rusty horizons were shaped from water permeable cryogenic waste in the oxygen environment of active permafrost, in the introductory phase of extraglacial and periglacial cryopedogenesis.
5. The main controlling pedomorphogenetic factors in the successive phases of the old-holocene process of rusting and following the young holocene podzolising were changing in the time, climate, vegetation and water conditions.
6. The factors which determine the development of soil cover mosaics were parent rock and relief. Therefore the soils under consideration have in their profiles

features of polygenetic zonal soils of mountain stages, and their inclusion in authogenic soils is justified.

7. Rusty and rusty podzolic soils which occur outside the range of the contemporary tundra and taiga on considerable areas of Central Europe are relic pleistocene-old holocene soils which have preserved the profile of the past environment, and the ongoing recent young-holocene processes are controlled by the conditions of subboreal climate and the corresponding vegetation.

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HOLOCEŃSKIE GLEBY RDZAWE I RDZAWE BIELICOWANE W TUNDRZE I TAJDZE ŚRODKOWEJ SZWECJI

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STRESZCZENIE

W opracowaniu przedstawiono charakterystykę niektórych fizycznych i fizyko-chemicznych właściwości kateny gleb glejobielicowych, rdzawych bielicowanych i rdzawych na północno-wschodnim stoku góry Åreskutan, w tundrze mszystej i krzewinkowej oraz w świerkowej tajdze północnej od wysokości 1190 do 410 m n.p.m. w środkowej Szwecji (tab. 1). Gleby te powstały z holoceńskich zwietrzelin i osadów środowiska proglacjalnego i peryglacjalnego w okresie ostatnich 9000–6000 lat. Współcześnie znajdują się one w fazie intensywnych procesów uruchamiania w poziomach Ofh agresywnych kwasów organicznych, które w poziomach E i Egg powodują pedochemiczny rozkład świeżych kriogenicznych odłamków minerałów i skał. Produkty tego rozkładu w obecności kwasów organicznych w roli koloidów ochronnych migrują w głąb profilów i częściowo są osadzane w poziomach Bhfe i Bhfegg o niewielkich miąższościach (tab. 2 i 3). Ze względu na dominację w kompleksie sorpcyjnym kationów Al^{3+} i Fe^{3+} (tab. 4 i 5) oraz eluwialną i krioturbacyjną migrację materii organicznej, gleby te można określić jako Al-Fe-próchniczne [Targuljan 1971].

Cechy współczesnego humidowego procesu bielicowania z eluwiacją i iluwiacją nakładają się na całej głębokości profilu na reliktowe, stabilnie cechy pedokriogenicznych poziomów rdzawych Bv, homogenicznych pod względem barwy i uziarnienia (tab. 1 i 2). Te poziomy ukształtowały się z wodoprzepuszczalnych holoceńskich kriogenicznych i lodowcowych zwietrzelin i osadów w tlenowym semiaridowym środowisku, we wstępnej fazie kriopedogenezy ekstraglacjalnej i peryglacjalnej. Kolejne fazy – najpierw procesu rdzawienia w warunkach suchszego klimatu, a następnie procesu bielicowania po zwilgotnieniu klimatu – powstały pod wpływem zmieniających się w czasie czynników pedomorfogenetycznych klimatu, roślinności i gospodarki wodnej. Mozaikowatość pokrywy glebowej w tym terenie warunkowały sprzężenie skały macierzystej i reliefu. Są to gleby poligenetyczne, stosunkowo bogate w próchnicę na całej głębokości profilu z intensywną wymianą i migracją jonową, wskazującymi na niestabilność kriohydrochemicznej równowagi jonowej we wszystkich poziomach. Gleby rdzawe i rdzawe bielicowe, poza zasięgiem współczesnej tundry i tajgi północnej na znacznych obszarach Europy Środkowej, są glebami reliktowymi. Zachowany został w nich profil dawnych środowisk, a współczesne procesy pedogeniczne są sterowane przez zupełnie inne, nieporównywalne warunki klimatu subborealnego i odpowiadającą roślinność tajgi południowej oraz lasów liściastych.

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