

Controls on carbon accumulation and storage in the mineral subsoil beneath peat in Lakkasuo mire, central Finland

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Summary

What processes control the accumulation and storage of carbon (C) in the mineral subsoil beneath peat? To find out we investigated four podzolic mineral subsoil profiles from forest and beneath peat in Lakkasuo mire in central boreal Finland. The amount of C in the mineral subsoil ranged from 3.9 to 8.1 kg m⁻² over a thickness of 70 cm and that in the organic horizons ranged from 1.8 to 144 kg m⁻². Rates of increase of subsoil C were initially large (14 g m⁻² year⁻¹) as the upland forest soil was paludified, but decreased to < 2 g m⁻² year⁻¹ from 150 to 3000 years. The subsoils retained extractable aluminium (Al) but lost iron (Fe) as the surrounding forest podzols were paludified beneath the peat. A stepwise, ordinary least-squares regression indicated a strong relation ($R^2=0.91$) between organic C concentration of 26 podzolic subsoil samples and dithionite–citrate–bicarbonate-extractable Fe (negative), ammonium oxalate-extractable Al (positive) and null-point concentration of dissolved organic C (DOC_{np}) (positive). We examined the ability of the subsoil samples to sorb dissolved organic C from a solution derived from peat. Null-point concentration of dissolved C (DOC_{np}) ranged from 35 to 83 mg l⁻¹, and generally decreased from the upper to the lower parts of the profiles (average E, B and C horizon DOC_{np} concentrations of 64, 47 and 42 mg l⁻¹). The DOC_{np} was positively correlated with percentage of soil C and silt and clay content. The concentration of dissolved organic C in pore water in the peat ranged from 12 to 60 mg l⁻¹ (average 33 mg l⁻¹), suggesting that the sorptive capacity of the subsoil horizons for C had been exhausted. We suggest that the increase of C contents in the subsoil beneath mires is related to adsorption of dissolved organic C and slow mineralization under anaerobic conditions.

Introduction

The accumulation of peat involves an interaction between plant production and carbon (C) losses by decay, mire fires, leaching and deposition of C into the mineral soil beneath the peat. Information concerning the influence of these factors on the average long-term rate of C accumulation is necessary when estimating the C cycle of mires and its relationship to climate change, which can be expected to have an impact on mire C fluxes. Studies on C dynamics to date have examined mainly the C cycle in deep peat deposits and have ignored C in the mineral subsoil of mires. Since about half of the mires in Finland, and perhaps an even greater percentage of mires in other boreal areas, are formed by paludification (formation of mire systems over previously forested land, grassland or even bare rock, due to climatic or autogenic processes), it is also

important to understand the significance of paludification to the accumulation and storage of C in mineral subsoils beneath peat (Turunen *et al.*, 1999).

The soil of boreal forests is estimated to contain 15% of the global C stored in the soil, excluding wetlands (Post *et al.*, 1982). The mineral subsoil under mires is an additional C sink that has been overlooked and could account for some 5% of the unaccounted C in the global C budget (Turunen *et al.*, 1999). Turunen *et al.* (1999) studied the C content (kg m⁻²) and long-term apparent rate of carbon accumulation (LORCA, g m⁻² year⁻¹) in the mineral subsoil of boreal mires in Finland formed by paludification by comparing them with podzolic soils at adjacent forest sites. Mean C content in the mineral subsoil of mire sites was 1.5-fold greater than that in adjacent forest profiles, and the average C input from peat into the mineral subsoil was 13.6 g m⁻² year⁻¹, depending on the time since paludification. There was no increase in LORCA with increasing age of the basal peat, but

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the trend was towards an increasing C content in the soil. Also, they found considerable variation both in the C content and in the LORCA in individual podzolic profiles below the peat.

Our knowledge of the formation and controls of C concentration in mineral soils is fragmentary, and laboratory experiments, especially on the dynamics of dissolved organic matter, have often contradicted field observations, which underlines the need for field studies (Kalbitz *et al.*, 2000). We have investigated the processes controlling the accumulation and storage of C in the mineral subsoil beneath peat. Two processes contributing to the accumulation of C in the subsoil mineral horizons beneath mires are the death and decomposition of plant roots and the leaching of pore water rich in dissolved organic C causing adsorption of C in the mineral horizons. The anaerobic conditions in the mineral subsoil would tend to preserve the C, once it has entered the system. Ratios of aerobic to anaerobic decomposition of organic matter in peat soils range from 5:1 to 40:1 (Scanlon & Moore, 2000). For four mineral soil profiles in the Lakkasuo mire complex, we examined the relationship between soil C concentration and extractable Fe and Al, and particle-size distribution. We conducted a sorption experiment to examine the ability of the subsoil samples to sorb C from a solution derived from peat. We established differences in the null-point concentration of dissolved organic C (DOC_{np}), at which there is neither gain nor loss of C in the solution, within and among profiles, and the relationships between DOC_{np} and sorption parameters and soil properties.

Materials and methods

Study area

The Lakkasuo mire complex (61°47'N, 24°18'E, and 150 m above sea level) is in the southern boreal zone representing the eccentric raised bog region of Finland (Ruuhijärvi, 1983). The bedrock consists mainly of granite, granodiorite, and mica gneiss. The mean annual temperature is +2.9°C, the mean

annual precipitation is 709 mm (Finnish Meteorological Institute, 1991), and the average long-term runoff from the catchment is 315 mm (Iivanainen *et al.*, 1999). The podzolized soils developed on glaciofluvial, well-sorted medium sand. The area emerged from the Baltic Sea around 10 500 radiocarbon years BP. Turunen *et al.* (1999) have described the area in detail.

For the present study we chose four representative profiles of the 63 studied by Turunen *et al.* (1999) from a transect in the southern margin of the area (Figure 1, Table 1) where C accumulation was linked to continuous paludification. We included one mire subsoil profile (site 4) with exceptionally small C content to examine the controls of C concentration within the whole range of C content (Figure 2, Turunen *et al.*, 1999). A spruce–pine swamp lies from the margin of the mire to site number 4, and from sites 4 to 8 there was a cotton grass–sedge–pine fen (Laine & Vasander, 1990). Most of the peat was moderately decomposed (H4–6) in von Post's (1922) 10-grade scale, and only some 10–20 cm of the bottom-most peat is highly decomposed (H8–9). The margin of the mire was paludified after the most recent fire in 1845 (Alm *et al.*, 1992).

Soil sampling and laboratory analyses

Samples were taken manually with a box sampler (8.5 cm × 8.5 cm × 100 cm), and a Russian pattern peat sampler (5 cm × 50 cm). The podzolic soils showed clear, strongly bleached eluvial horizons (E) up to 15 cm thick and illuvial horizons containing dark brown humus and sesquioxides (B) up to 55 cm thick. The cores were divided into A, E, B and C horizons.

Mineral subsoil samples were taken to a depth of 70 cm from the bottom of the A horizon. We compared only E, B and C horizons to avoid errors in interpretation, because identification of the A horizon at the peat–mineral soil transition was uncertain in many cases. We subsampled the core so as to avoid contamination. Core samples were sealed in plastic bags and stored at +5°C until dried at 70°C to a constant

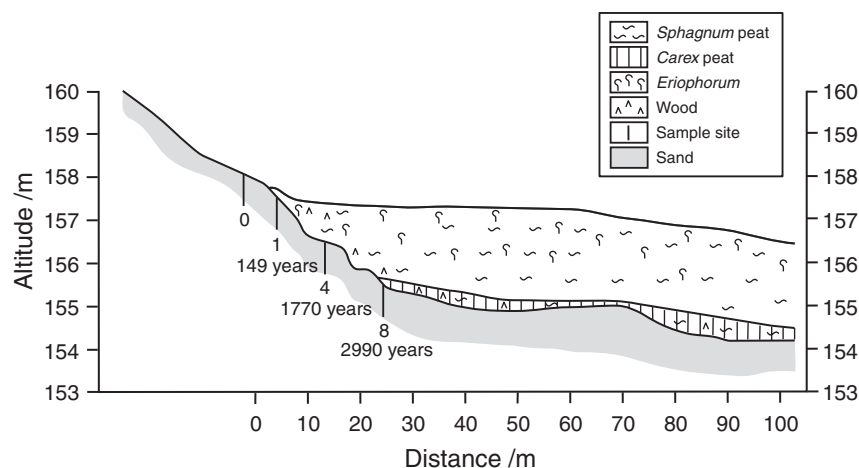


Figure 1 The transect profile in Lakkasuo mire, showing the peat types and locations of the sampling sites, and sample codes (Table 1). The mire type from the margin of the mire through site 4 is spruce–pine swamp and from site 8 on is cotton grass–sedge–pine fen. Age of the basal peat refers to the calibrated calendar years (Stuiver & Reimer, 1993) from the conventional ^{14}C datings. Age of the basal peat at site 1 (149 years) is based on dendrochronological dating (Alm *et al.*, 1992).

Table 1 Sample description and ^{14}C dates of basal peat from Lakkasuo mire

Sample code	Peat depth /cm	Peat material ^a	^{14}C date /year BP	Most probable date ^b /calendar year BP
1	12–13	LS, H9	–	149 ^c
4	85–91	LCS, H7–8	1820 ± 90	1770
8	182–189	LS, H8–9	2840 ± 90	2990

^aPeat constituents are: C, *Carex*; L, wood; S, *Sphagnum*. H is degree of decomposition in von Post's (1922) 10-grade scale.

^bThe most probable date refers to the calibrated calendar years (Stuiver & Reimer, 1993) from conventional ^{14}C dating.

^cAge of the basal peat in site 1 is based on dendrochronological dating (Alm *et al.*, 1992).

mass, passed through a 2-mm sieve, and weighed to obtain dry bulk density. Carbon was determined on a LECO CHN-600 analyser. A replicate for C concentration analysis was systematically taken at every fifth sample. We calculated the mass of C per unit of surface area (g m^{-2}) in a given soil layer, by multiplying the C concentration in the unit volume of soil by the thickness of the layer. The C content of the sampled 70-cm-deep mineral soil layer was then calculated from the sum of all C contents of the sublayers. Turunen *et al.* (1999) describe the mineral subsoil sampling, ^{14}C dating and laboratory methods in detail. The complete peat profiles of study sites 1, 4 and 8 were sampled and analysed according to Turunen *et al.* (2002).

Ammonium oxalate- and dithionite-citrate-bicarbonate-extractable Fe (Fe_o and Fe_d) and ammonium oxalate-extractable Al (Al_o) of the <2-mm fraction were determined as described by McKeague (1978, pp. 98–104). The concentration of Fe was determined on a Hitachi Z-600 atomic absorption spectrophotometer, and Al on a Varian Spectra 400 atomic absorption spectrophotometer. Particle-size analysis was done by sieves of 2, 0.63 and 0.2 mm, and each fraction was weighed. A 0.5-g sample of the <0.2-mm fraction of oven-dry soil was weighed into a glass bottle, 5 ml of 0.1 M sodium pyrophosphate was added, and the suspension was stirred and allowed to stand for 24 hours. The <0.2-mm fraction was analysed on a HELOS 12LA SYMPATEC-diffraction spectrometer analyser with helium–neon laser (Sympatec, 1985) and evaluated together with the >0.2 mm fraction with the SYMPATEC QX-software package.

The methods in Moore *et al.* (1992) were used to examine the ability of the soil samples to sorb dissolved C from a solution derived from peat. We produced a solution rich in dissolved C by soaking a mixture of surface and subsurface peat from the Mer Bleue Bog, Ontario, in water for three days and filtering. The resulting solution was then diluted with distilled water to produce a series of six solutions ranging in dissolved C from 0.7 to 61 mg l^{-1} , similar to the range of concentrations in forest and peat soils reported by Kalbitz *et al.* (2000). Ten millilitres of the solution was shaken gently with 1 g of the soil sample, left overnight, and then filtered through Gelman A/E 1 μm paper. The concentration of dissolved C was determined on a Shimadzu TOC 5000 analyser. We determined sorption characteristics by regressing the change in dissolved C concentration in the supernatant against the original dissolved C

concentration, determining the coefficient of determination R^2 , slope m , and intercept b . Functionally, m and b can be viewed as measures of the soil's ability to sorb dissolved organic C (Nodvin *et al.*, 1986). The regression was inverted to calculate the null-point concentration of dissolved C (DOC_{np}), the original concentration at which there is neither adsorption nor desorption of C by the soil. To compare the differences in the soil characteristics between study sites, we did an analysis of variance (ANOVA). The normal distribution, linearity and homoscedasticity of residuals and the homogeneity of variances for linear regression analysis were tested by the Kolmogorov–Smirnov, Levene statistical tests and partial regression plots. Stepwise, ordinary least-squares regression was used to examine which soil physical properties could predict the C concentrations in the subsoil. Pearson's correlation coefficient r was used to express correlation.

Results

Soil characteristics

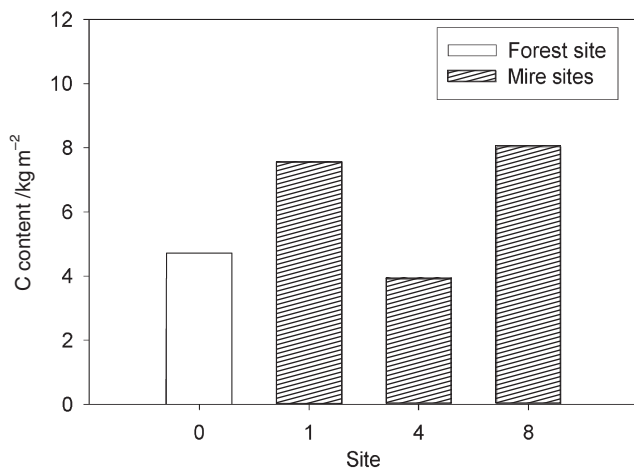
The subsoil samples were well-sorted medium sands (Table 2, average grain size 268–355 μm). The C content in the mineral subsoil of sites 1 and 8 was significantly larger (7.6 and 8.1 kg m^{-2}) than in the adjacent forest profile (4.7 kg m^{-2}), but mire site 4 had a smaller C content (3.9 kg m^{-2} , Figure 2). In the three paludified sites, on average 17% of the C was in the E horizon, 67% was in the B horizon, and 16% was in the parent material. In the upland forest profile, the corresponding C values were 18%, 70% and 12%, respectively (Table 2). The C contents in the organic horizons of forest floor and peat at sites 0, 1, 4 and 8 were 1.8, 5.7, 69 and 144 kg m^{-2} , respectively. The average long-term apparent rate of carbon accumulation (LORCA) of peat deposits at sites 1, 4 and 8 was 19.2, 19.5 and 23.7 $\text{g m}^{-2} \text{year}^{-1}$, respectively.

There was little variation in dry bulk density, average grain size, and clay, silt and sand content among the four soil profiles (Table 2). Extractable Fe (Fe_o and Fe_d) and Al_o concentrations ranged from 0.1 to 16.2 and 0.5 to 14.5 g kg^{-1} , respectively. The dithionite and oxalate extractions of Fe were strongly correlated ($r=0.99$) but Fe and Al concentrations were relatively independent ($r=0.30$ – 0.34 , Table 3). There was an increase in extractable Fe and Al from the forest podzol to site 1 and then a decrease in extractable Fe

Table 2 Properties of mineral soil samples from the four profiles. The final four columns present the results of the DOC sorption study, in terms of the regression coefficient, m , intercept, b , the coefficient of determination, R^2 , and null-point concentration of dissolved C, DOC_{np}

Site	Horizon	Depth /cm	Dry bulk density / kg m^{-3}	Clay	Silt	Sand	C / g kg^{-1}	Fe_o	Al_o	Fe_d	m	$-b$ / mg g^{-1}	R^2	DOC_{np} / mg l^{-1}
0	E	2–15	1266	19	82	900	5.1	0.8	0.5	1.8	0.41	0.30	0.83	53.8
	B	15–40	1077	18	66	916	9.3	3.3	4.7	5.2	0.33	0.18	0.71	74.3
	B	40–55	1275	14	60	926	4.0	2.5	4.3	4.2	0.33	0.17	0.94	50.5
	C	55–65	1166	12	49	939	2.8	1.8	2.5	3.2	0.43	0.17	0.97	39.2
	C	65–70	1398	18	64	918	3.3	1.1	1.9	2.2	0.31	0.16	0.87	52.3
1	E	1–13	1113	13	78	908	16.6	0.2	1.2	0.3	0.56	0.47	0.86	83.4
	B	13–37	1055	14	65	922	18.2	2.3	10.0	5.2	0.46	0.24	0.93	52.5
	B	37–54	1270	8	31	961	2.7	7.3	6.8	14.3	0.39	0.16	0.81	42.0
	C	54–70	1292	3	11	986	1.0	6.9	2.8	16.2	0.23	0.10	0.80	41.6
4	E	2–9	616	8	47	928	4.4	0.1	0.7	0.3	0.20	0.12	0.66	58.7
	E	9–14	810	10	50	927	7.4	0.6	3.1	1.2	0.35	0.15	0.93	41.5
	B	14–24	971	7	36	945	10.7	0.8	5.1	1.5	0.21	0.09	0.76	43.8
	B	24–34	1090	7	34	956	5.7	0.6	4.3	1.4	0.33	0.12	0.97	34.6
	B	34–39	1057	7	30	959	7.0	0.7	5.0	1.4	0.25	0.11	0.92	45.9
	B	39–44	1001	9	36	955	5.3	0.7	4.3	1.6	0.25	0.09	0.89	35.9
	C	44–51	1107	8	28	964	4.7	0.9	4.3	1.8	0.36	0.13	0.84	36.1
	C	51–70	1300	11	27	962	3.2	0.4	2.0	1.2	0.21	0.10	0.87	44.7
8	E	2–6	811	10	62	904	26.4	0.1	3.4	0.3	0.47	0.27	0.90	55.2
	B	6–11	864	13	70	895	30.9	0.3	8.3	0.5	0.49	0.26	0.99	53.2
	B	11–16	902	22	117	849	49.1	0.4	11.6	0.9	0.50	0.27	0.98	54.3
	B	16–21	956	17	93	874	31.7	0.8	14.5	1.0	0.49	0.23	0.97	46.9
	B	21–26	1001	11	60	917	14.9	0.5	6.2	1.0	0.41	0.19	0.96	46.1
	C	26–31	912	6	33	952	8.5	0.5	4.3	0.9	0.43	0.17	0.99	40.8
	C	31–36	852	7	36	953	5.6	0.4	2.5	0.8	0.34	0.15	0.96	43.4
	C	36–43	920	9	41	950	2.5	0.3	1.9	0.5	0.46	0.20	0.91	43.1
C	43–70	1141	12	39	943	2.0	0.2	1.1	0.4	0.40	0.14	0.94	35.3	

concentration in profiles 4 and 8, presumably related to the reduction and mobilization of Fe under anaerobic conditions beneath the peat. Excluding site 4, the trend was towards

**Figure 2** (a) Mean C content in the uppermost 70 cm of the four mineral subsoils (horizons E through C) at sites shown in Figure 1.

increasing C and Al concentrations and decreasing Fe concentration in the subsoil with increasing age of the basal peat.

Carbon concentration of the samples ranged from 1 to 49 g kg^{-1} , the largest values being in the upper B horizons of profile 8. If variables were not normally distributed, \log_{10} transformation was used to correct for heteroscedasticity, non-linearity, and outliers. The \log_{10} -transformed C concentrations of the samples showed a strong correlation with soil dry bulk density (negative), silt and clay content (positive), sand content (negative) and extractable Al_o concentration (positive), and a weakly negative one with Fe_d concentration (Table 3). A stepwise, ordinary least-squares regression indicated a fairly strong relation ($R^2 = 0.63$) between organic C and Fe_d (negative) and Al_o (positive):

$$\log_{10}(\text{orgC}) = -0.230 - 0.577 \log_{10}(\text{Fe}_d) + 0.867 \log_{10}(\text{Al}_o). \quad (1)$$

The sorption study revealed DOC_{np} concentrations ranging from 35 to 83 mg l^{-1} , with a general decrease with depth within the profiles. Averages in the strongly bleached eluvial E, illuvial B and in the parent material C horizons were 64, 47

Table 3 Correlation matrix for properties of Lakkasuo mineral soil samples ($n = 26$)

	Dry bulk density	$\log_{10}(\text{clay})$	$\log_{10}(\text{silt})$	$\log_{10}(\text{sand})$	$\log_{10}(\text{C})$	$\log_{10}(\text{Fe}_o)$	$\log_{10}(\text{Al}_o)$	$\log_{10}(\text{Fe}_d)$	m	b	$\log_{10}(\text{DOC}_{np})$
Dry bulk density	1.000										
$\log_{10}(\text{clay})$	0.097	1.000									
$\log_{10}(\text{silt})$	-0.229	0.898	1.000								
$\log_{10}(\text{sand})$	0.172	-0.736	-0.832	1.000							
$\log_{10}(\text{C})$	-0.499	0.494	0.723	-0.610	1.000						
$\log_{10}(\text{Fe}_o)$	0.617	-0.134	-0.291	0.194	-0.350	1.000					
$\log_{10}(\text{Al}_o)$	-0.129	0.062	0.141	-0.124	0.537	0.338	1.000				
$\log_{10}(\text{Fe}_d)$	0.628	-0.176	-0.351	0.232	-0.397	0.987	0.303	1.000			
m	-0.147	0.470	0.630	-0.472	0.555	-0.211	0.226	-0.285	1.000		
b	0.065	-0.532	-0.687	0.551	-0.575	0.264	0.061	0.329	-0.826	1.000	
$\log_{10}(\text{DOC}_{np})$	-0.032	0.516	0.613	-0.530	0.433	-0.191	-0.282	-0.226	0.341	-0.777	1.000

and 42 mg l^{-1} , respectively (Table 2). The slope of the sorption regression, m , ranged from 0.21 to 0.56, with the smallest values in site 4. The intercepts, b , ranged from 0.09 to 0.47 mg g^{-1} , generally declining with depth in the four profiles. The use of dried samples in the sorption study means that the magnitude of DOC_{np} and b should be treated with caution, as Kaiser *et al.* (2001) have shown that the method of storage and extraction conditions affect the intercept, though the slope remained unaffected by the methods employed.

The slope of the sorption regression, m , was positively correlated with silt and organic C and negatively with intercept, b . The intercept, b , was correlated with sand (positive) and silt, clay and organic C (negative) (Table 3). The null-point concentration of dissolved C (DOC_{np}) was positively correlated with clay, silt and organic C, and negatively with sand. There was no significant correlation between either m , b or DOC_{np} and extractable Fe or Al, though there were correlations among the three sorption parameters (Table 3). The relations between sorption parameters and organic C are illustrated in Figure 3.

In partial regression analysis, the strong relation between soil organic C and DOC_{np} concentration showed that C adsorption was diminished at large concentrations of organic C (Figure 4d). The overall DOC_{np} was large in nearly all samples, suggesting that the sorption capacity is small and that water percolating down from the upland forest floor or the peat profiles is probably near equilibrium with solution rich in organic C (Table 2).

Incorporation of DOC_{np} into the regression analysis of organic C increased R^2 from 0.63 to 0.91, Equation (2), with weak relationships between the three independent variables (Table 3). Also, the adjusted R^2 increased significantly from 0.60 to 0.90:

$$\log_{10}(\text{orgC}) = -4.329 - 0.479 \log_{10}(\text{Fe}_d) + 1.023 \log_{10}(\text{Al}_o) + 2.535 \log_{10}(\text{DOC}_{np}). \quad (2)$$

Figure 4(a) illustrates the weak correlation between the $\log_{10}\text{C}$ concentration and regression standardized residuals

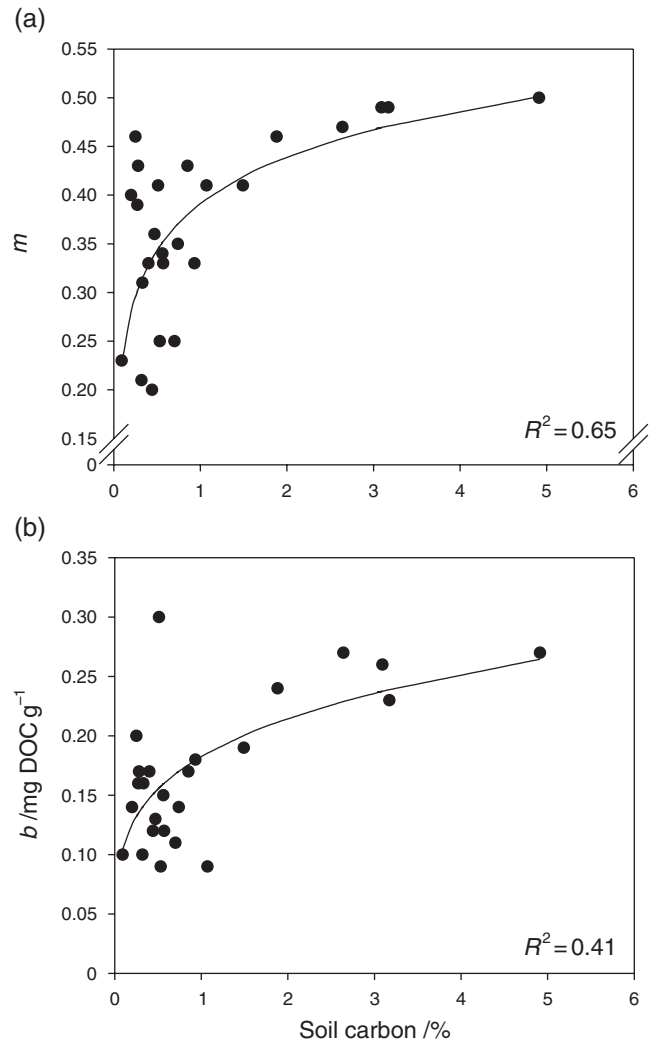


Figure 3 Sorption parameters slope (m) and intercept (b) plotted against soil C. The regression lines are: (a) $m = 0.391 + 0.07 \log_{10}(\text{soil C})$, and (b) $b = 0.182 + (\text{soil C})^{0.234}$.

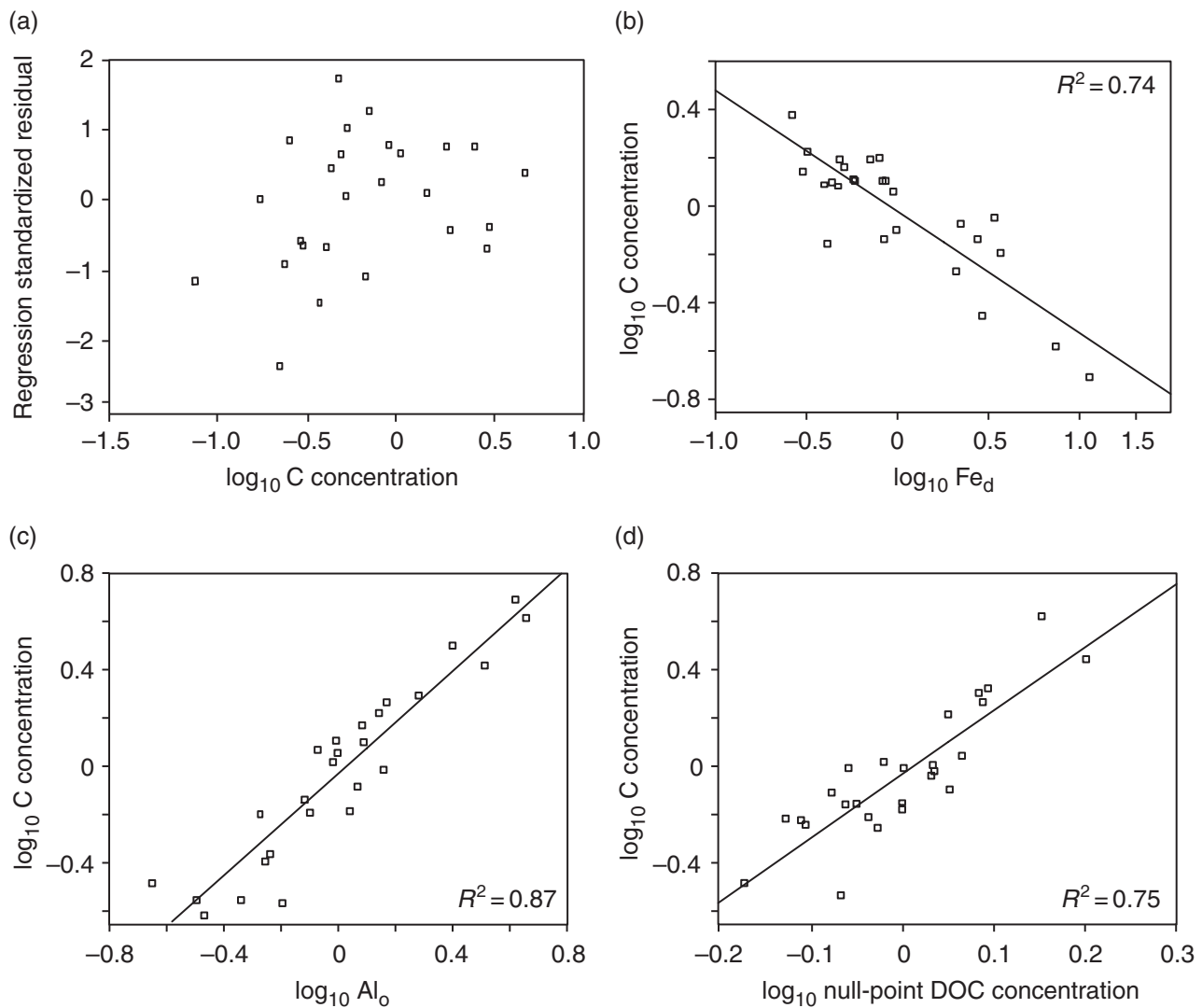


Figure 4 (a) Scatterplot illustrating the relation between the \log_{10} (C concentration) and regression standardized residuals, and partial regression plots showing the influence of (b) \log_{10} (Fe_d), (c) \log_{10} (Al_o) and (d) \log_{10} (DOC_{np}) concentration on the \log_{10} (C concentration), with the influence of the other independent variables removed.

(observed – estimated) of the model. The influence of each independent variable (\log_{10} Fe_d, \log_{10} Al_o and \log_{10} DOC_{np} concentration) on \log_{10} C concentration, while removing the influence of the other independent variables, is shown in partial regression plots (Figure 4b–d). Carbon concentrations of the soils were strongly related to Fe_d (negative), Al_o (positive) and DOC_{np} (positive) concentrations.

Discussion

Our results indicate a strong relation between organic C and other chemical properties of the soil, such as Al and Fe, null-point concentration of dissolved organic C and the sorption regression coefficient, m , and the intercept, b . According to

Neff & Asner (2001), these soil properties are indicators of the reactive soil surfaces available for ionic and physical associations. There seems to be a clear correspondence between the C concentration and the parameters m and b (Figure 3). These parameter values of C sorption are similar to those reported for podzolized soil profiles by Moore *et al.* (1992). There was a general decreasing trend in b with soil depth, but no clear pattern with m . The intercept, b , and C relation reflects a rapid equilibration between dissolved and solid organic matter (McDowell & Wood, 1984; Neff & Asner, 2001). The relation between the sorption regression coefficient, m , and the soil C varies among studies (e.g. Moore *et al.*, 1992; Kaiser *et al.*, 1996), and here the relation between m and C concentration was positive. Parameters m and b were also strongly correlated with the silt content

of the soil, probably related to the surface area (Mayer, 1994) as the clay content of these soils is small (< 2%).

Jardine *et al.* (1989), Moore *et al.* (1992) and Kaiser & Zech (1998) have shown that Fe and Al oxides and hydroxides are the most important sorbents of dissolved organic C in soils. Mulder *et al.* (1992) showed that the dissolved organic C released from a podzol B horizon was much diminished by the addition of Al. Polyvalent cations such as Al^{3+} and Fe^{3+} can link negatively charged functional groups of organic molecules and diminish their solubility by flocculation or binding them to negatively charged exchange sites (Tipping & Woof, 1990; Kalbitz *et al.*, 2000). In our study, the negative relation between extractable Fe and organic C might be related to the development of strong reducing conditions beneath the peat, resulting in a loss of Fe, but retention of C.

Turunen *et al.* (1999) showed that the LORCA in the mineral subsoil was largest just after the mire began to form. There was no cumulative increase in C storage with increasing age of the basal peat, but there was a trend to increasing amounts of C in the mineral subsoil. Deep roots of trees, vascular plants and shrubs may add to the C input. Production in the soil constitutes the major source of organic matter in mires, through deep-penetrating fine roots of vascular plants (e.g. *Eriophorum*, *Carex*) into the catotelm (Wallén, 1986; Saarinen, 1996; Moore *et al.*, 2002). Individual species of mire plants have specific below-ground distributions, varying from a concentration in the uppermost few centimetres of the acrotelm to down into the catotelm. In fens, the fine roots of sedges can form a direct input of up to $20 \text{ g C m}^{-2} \text{ year}^{-1}$ below 30 cm (Saarinen, 1996). These values are of the same magnitude as the estimate of the average C input rate of $13.6 \text{ g m}^{-2} \text{ year}^{-1}$ from peat into the mineral subsoil by leaching (Turunen *et al.*, 1999).

The smaller LORCA rates in the mineral subsoil during development of the mire accord with the observation that organic matter already adsorbed to mineral surfaces diminishes the potential adsorption capacity of the sorbent (Jardine *et al.*, 1989; Moore *et al.*, 1992; Kaiser & Zech, 1998). We found fine roots in all soil profiles, but we did not determine their quantity and origin. Deposits of iron oxide commonly occur on roots growing in water-saturated soils and wetlands, related to the internal transport of O_2 to submerged roots and to O_2 leakage into the rhizosphere (Fisher & Stone, 1991). The increased Fe concentration was evident in site 1, where the peat was only 13 cm thick, but under thicker and older peat the concentrations of Fe were smaller than the mire margin or adjacent forest soil site, probably because of intense reduction and leaching (Table 2).

Fluxes of dissolved organic C are directly and indirectly controlled by the hydrology of the mire. The export of dissolved C from the Lakkasuo mire ranges from 8 to $17 \text{ g m}^{-2} \text{ year}^{-1}$ (Sallantausta, 1992), typical of temperate mires and catchments dominated by mires and similar to the rate of C accumulation in peat. Transport of C into and sorption by the subsoil may be the major source of the organic C in the

mineral soil of these mires. Concentrations of dissolved C increase following rewetting after drought, and anaerobic conditions may increase the release of dissolved organic matter from soil. In Lakkasuo mire, the average concentration of dissolved C in the pore water from peat is 33 mg l^{-1} , ranging from 12 to 60 mg l^{-1} (T. Sallantausta, Finnish Environment Centre, unpublished data). Percolating of dissolved C into the mineral horizons beneath the mire might be adsorbed, and the mineralization of C will be retarded by the anaerobic conditions compared with mineralization of sorbed C in adjacent well-drained soils. This sorption and retention of C is likely to be pronounced early in the development of the mire but be reduced as the sorption capacity is exhausted. In Finland, C content in the subsoil averaged 4.1 kg m^{-2} in upland forest soils and 6.2 kg m^{-2} beneath adjacent mires. Long-term accumulation of C (LORCA) averaged $14 \text{ g m}^{-2} \text{ year}^{-1}$, ranging from $19 \text{ g m}^{-2} \text{ year}^{-1}$ at sites < 500 years old to $1 \text{ g m}^{-2} \text{ year}^{-1}$ at sites older than 500 years (Turunen *et al.*, 1999). The LORCA of peat deposits at Lakkasuo ranged from 19 to $24 \text{ g m}^{-2} \text{ year}^{-1}$, similar to recent average estimates of 19, 17 and $20 \text{ g m}^{-2} \text{ year}^{-1}$ in Finland, Russia and Canada throughout the Holocene (Turunen *et al.*, 2001, 2002; Vitt *et al.*, 2000). Thus, increase in subsoil C in mires represents a significant proportion of the total accumulation of C in the initial stages, but becomes insignificant as the sorption capacity is exhausted.

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