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Contemporary carbon stocks of mineral forest soils in the Swiss Alps

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Abstract. Soil organic carbon (SOC) has been identif ed as the main global terrestrial carbon reservoir, but considerable uncertainty remains as to regional SOC variability and the distribution of C between vegetation and soil. We used gridded forest soil data (8-km × 8-km) representative of Swiss forests in terms of climate and forest type distribution to analyse spatial patterns of mineral SOC stocks along gradients in the European Alps for the year 1993. At stand level, mean SOC stocks of 98 t C ha^{-1} ($N = 168$, coeff cient of variation: 70%) were obtained for the entire mineral soil prof le, 76 t C ha^{-1} ($N = 137$, CV: 50%) in 0–30 cm topsoil, and 62 t C ha^{-1} ($N = 156$, CV: 46%) in 0–20 cm topsoil. Extrapolating to national scale, we calculated contemporary SOC stocks of 110 Tg C (entire mineral soil, standard error: 6 Tg C), 87 Tg C (0–30 cm topsoil, standard error: 3.5 Tg C) and 70 Tg C (0–20 cm topsoil, standard error: 2.5 Tg C) for mineral soils of accessible Swiss forests (1.1399 Mha). According to our estimate, the 0–20 cm layers of mineral forest soils in Switzerland store about half of the C sequestered by forest trees (136 Tg C) and more than f ve times more than organic horizons (13.2 Tg C).

At stand level, regression analyses on the entire data set yielded no strong climatic or topographic signature for forest SOC stocks in top (0–20 cm) and entire mineral soils across the Alps, despite the wide range of values of site parameters. Similarly, geostatistical analyses revealed no clear spatial trends for SOC in Switzerland at the scale of sampling. Using subsets, biotic, abiotic controls and categorial variables (forest type, region) explained nearly 60% of the SOC variability in topsoil mineral layers (0–20 cm) for broadleaf stands ($N = 56$), but only little of the variability in needleleaf stands ($N = 91$, $R^2 = 0.23$ for topsoil layers).

Considerable uncertainties remain in assessments of SOC stocks, due to unquantif ed errors in soil density and rock fraction, lack of data on within-site SOC variability and missing or poorly quantif ed environmental control parameters. Considering further spatial SOC variability, replicate pointwise soil sampling at 8-km × 8-km resolution without organic horizons will thus hardly allow to detect changes in SOC stocks in strongly heterogeneous mountain landscapes.

Introduction

At the local scale, soils play a central role in terrestrial ecosystems serving both as habitat for plants, soil animals and microorganisms and as water and nutrient reservoir for plants. More recently, soils have gained attention within the global change debate as largest terrestrial carbon (C) pool. According to Batjes (1996) organic and mineral soil horizons store about 1505 Pg C surpassing, thus, the C pools of vegetation (610 Pg C) and atmosphere (750 Pg C) by more than a factor of two (Schimel et al. 1995). Moreover soil organic carbon (SOC) is a signif cant dynamic component of the global C cycle: First, a substantial fraction of C in upland soils has characteristic turnover times of a few decades only (Harrison et al. 1993; Trumbore et al. 1996; Perruchoud et al. 1999a). Second, there is increasing evidence for heat and moisture controlling heterotrophic respiration (Raich & Potter 1995), organic matter decomposition (Kirschbaum 1995; Moore et al. 1999), SOC stocks (Homann et al. 1995; Gärdenäs 1998) and age distribution of SOC (Tate et al. 1995; Townsend et al. 1995). Third, land-use change can within a few decades lead to substantial SOC depletion (Davidson & Ackerman 1993). Thus, since CO_2 emissions from organic matter decomposition in terrestrial ecosystems amount to about 50–60 Pg C yr^{-1} (McGuire et al. 1995; Post et al. 1997), i.e. roughly one order of magnitude more than the anthropogenic CO_2 emissions, even small changes in the functioning of terrestrial ecosystems can perturb the global C cycle substantially. Forest soils, in particular, are a huge C pool (787 Pg C (Dixon et al. 1994)) and considerable changes of SOC pools and f uxes are to be expected in managed forests in response to afforestation (Harrison et al. 1995; Smith et al. 1997), timber harvesting (Pennock & Van Kessel 1995) and cultivation of previously untilled soils (Balesdent et al. 1988; Davidson & Ackerman 1993).

Sophisticated methods have now become available to assess changes in forest C stocks by integrative measures (GACGC 1998): Micrometeorological techniques based on CO_2 gradient measurements represent the net CO_2 exchange between atmosphere and biosphere over spatial scales of one km^2, larger spatial scales (50-km $\times$ 50-km areas) can be assessed with the convective boundary layer method. These methods can be used to estimate C budgets of major ecosystem types (Wofsy et al. 1993), but serious problems remain when spatially extrapolating to (sub-)continental scales from a few tower measurements (Martin et al. 1998). Moreover, sudden disturbance cannot be analysed with this approach. Thus, the classical approach of periodically inventorying forest C pools (i.e. overstory and understory vegetation, forest f oor and soil) at stand level remains an important source of information and combination of all three approaches the best way to achieve a better understanding of the state and dynamics of C sequestration in forests.

Despite its importance SOC has until now often been neglected in national forest inventories. Most European assessments have used one average SOC density for forest soils derived from a number of soil prof les (Burschel et al. 1993; Murillo 1994), or combined soil and vegetation type maps with published soil prof le data assumed as representative (Körner et al. 1993; Nabuurs & Mohren 1993; Paulsen 1995; Howard et al. 1995; Milne & Brown 1997). Finland (Kauppi et al. 1997), and Germany (R. Barritz, BFH *pers. comm.*) are among the few European countries where soil was sampled systematically within the national forest inventory to estimate SOC.

Alternatively, statistical relationships between SOC and environmental quantities controlling soil organic matter (SOM) decomposition such as temperature, precipitation and actual evapotranspiration (Burke et al. 1989; Grigal & Ohmann 1992; Homann et al. 1995; Liski & Westman 1996) are established to arrive at regional estimates of SOC pools. Edaphic factors such as clay mineralogy (Parf tt et al. 1996; Torn et al. 1997) or soil texture (Burke et al. 1989; Davidson 1995; Arrouyas et al. 1995) have also been successfully related to SOC sequestration, besides site characteristics such as productivity (Liski & Westman 1996), soil pH (Hanawalt & Whittaker 1976), soil moisture drainage (Davidson 1995; Homann et al. 1995) and topographical parameters (aspect, slope and elevation) (Haber 1985; Datta et al. 1989; Arrouays et al. 1998). Such studies complement process-oriented studies and allow to identify ecosystem types, or regions of high sensitivity with respect to natural and anthropogenic changes in the environment.

Our study had two objectives: First to give a national estimate for the C in mineral soils of upland forests in Switzerland for the mid 1990s, thus complementing already existing national estimates of forest tree and organic horizon C. Here, we additionally emphasised the topsoil layers (0–20 cm and 0–30 cm soil depth) with their high potential of mobilising substantial amounts of C under climate or land-use changes. Second, we used the wealth of available site-specif c biotic and abiotic parameters to reveal the existence of signif cant relationships between SOC stocks and specif c environmental factors in mountain forest soils of Central Europe.

Materials and methods

Study area

Forests cover about 29% or 12,000 km^2 of Switzerland and extend from 200 to 2,300 meters above sea level (SAEFL 1997; Brassel & Brändli 1999). They strongly vary with respect to productivity, stocking density and soil characteristics in response to differences in geology, topography, climate, soil formation and land-use. Site conditions in Swiss forests are representative of many Central and Northern European forest ecosystems. The spectrum of Swiss forests ranges from broadleaf forests on moderately drained, periodically dry, warm soils at low elevations, to strongly drained, warm soils in the Jurassic Alps and strongly drained, cold soils under needleleaf forests in the Alps. Spruce, beech and f r are the dominant species in Switzerland with about 80% of the total timber volume reserves (Stierlin & Ulmer 1999). Lower elevation stands have a complex vegetation history due to intensive management, so that spruce contributes substantially in this naturally beech dominated region. With increasing elevation more natural forests prevail, and broadleaf trees are gradually replaced by needleleaf vegetation. Half of the forest area is found at altitudes above 1,000 meters with about 75% located on intermediate slopes where hydrological inputs and outputs are roughly at equilibrium (EAFV 1988). Climate is characterised by the strong altitudinal gradient in Switzerland and ranges from intra-alpine and continental (annual precipitation sum < 500 mm) to insubrian (July temperature > 20 °C, annual precipitation sum > 1600 mm), or from temperate (Plateau) to cold climate (Alps) (Brzeziecki et al. 1993).

Soil data

Forest soils had been sampled in 1993 in the context of the Swiss forest damage inventory (Lüscher et al. 1994; Vanmechelen et al. 1997) at a spatial resolution of 8-km $\times$ 8-km (Figure 1). Not all of the organic horizon was sampled, so that C stocks here refer to mineral soil horizons. Soil prof les were dug down to the C horizon. Sampling of mineral soil in the top 80 cm was done using the soil layers specif ed by UN/ECE (1994), below 80 cm depth soil sampling was done by soil horizons. For the top 20 cm, soil was sampled continuously at 0–5 cm, 5–10 cm and 10–20 cm depth. The sampled soils were described as Cambisols ($N = 41$), Fluvisols (1), Gleysols (11), Lithosols (1), Luvisols (20), Podsols (28), Rankers (8), Regosols (20), and Rendzinas (38) using a simplif ed FAO classif cation. Histosols which only account for about 0.1% of the forested area in Switzerland and typically have thicknesses of the order of 10 cm were not encountered on our sampling

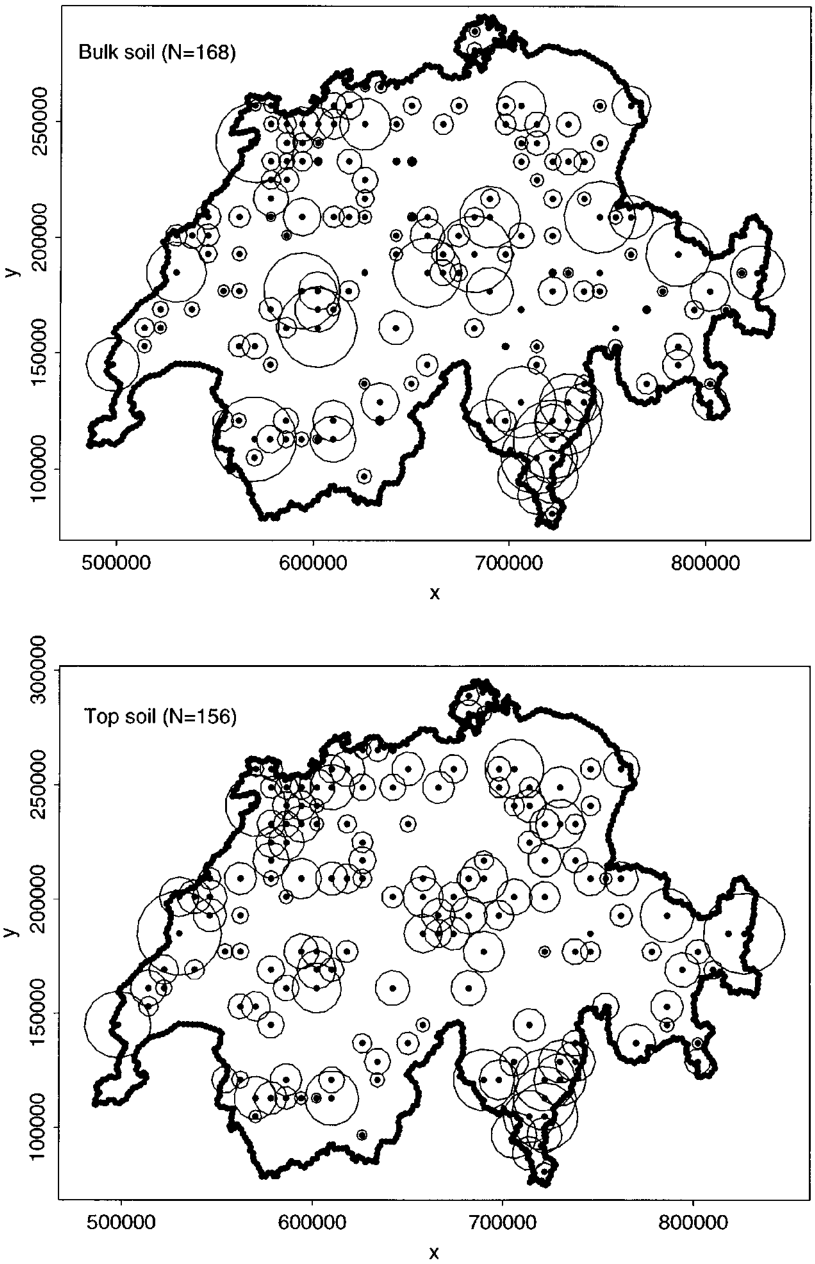

Figure 1. Spatial distribution of SOC in total (top, $N = 168$) and 0–20 cm mineral soil layers (bottom, $N = 156$) of upland forest soils at 8-km × 8-km resolution. Sampling locations are indicated by dots centred on the circles whose radii are proportional to the size of SOC stocks. Spatial coordinates ('x', 'y') are in units of meters and refer to Swiss National Coordinates.

grid. Field capacity was determined *in situ* according to AAB (1994). Additionally, qualitative information about shallowness, rock fraction and water holding capacity of the soils was available at national scale from Frei et al. (1980).

The national C reservoir of Swiss forest soils was assessed on the basis of 168 soil profiles. SOC was integrated across depth for layers with detectable C concentration or until bedrock was met (this is referred to as maximum soil depth hereafter) assuming homogeneity across pedogenic horizons and linear changes in C concentration across unsampled layers. To calculate SOC density in mass per area units, we used *in situ* estimates for layer thickness and the volumetric coarse soil fragment fraction $\geq$ 2 mm. Soil C concentration (in [g C/(100 g fine earth)]) and bulk density of fine earth were determined in the laboratory (see below).

Topographical data

Sample site elevations were based on a digital elevation model with a spatial resolution of 25-m $\times$ 25-m (Bundesamt für Landestopographie 1994). The same digital elevation model was used to derive slope, aspect, total curvature and bioclimatic parameters (see below).

Bioclimatic data

Mean annual temperature, annual precipitation sum, and annual actual evapotranspiration were available for every sample site. Climatic data were derived from mean monthly climate stations records (1961–1990) using regression analysis and spatial extrapolation of residuals, i.e. spline for precipitation (Kienast 1998) and kriging for temperature (L. Zgraggen ETH Zürich, *pers. comm.*). Evapotranspiration was obtained by a simplified version of ForClim–E (Bugmann & Cramer 1998), assuming spatially uniform climate variabilities and a constant field capacity of 15 cm (H. Lischke, WSL Birmensdorf, *pers. comm.*). A characterisation of the distribution of soil samples and sampling units of the Second Swiss Forest Inventory (SNFI 2, 1993–1995) in the temperature-precipitation parameter space is given in Figure 2.

Vegetation data

Forest vegetation data were available from the SNFI 2 at 1.4-km $\times$ 1.4-km resolution (Brassel & Brändli 1999). This allowed for an estimate of tree carbon including foliage, twigs, branches, stem, coarse and fine roots

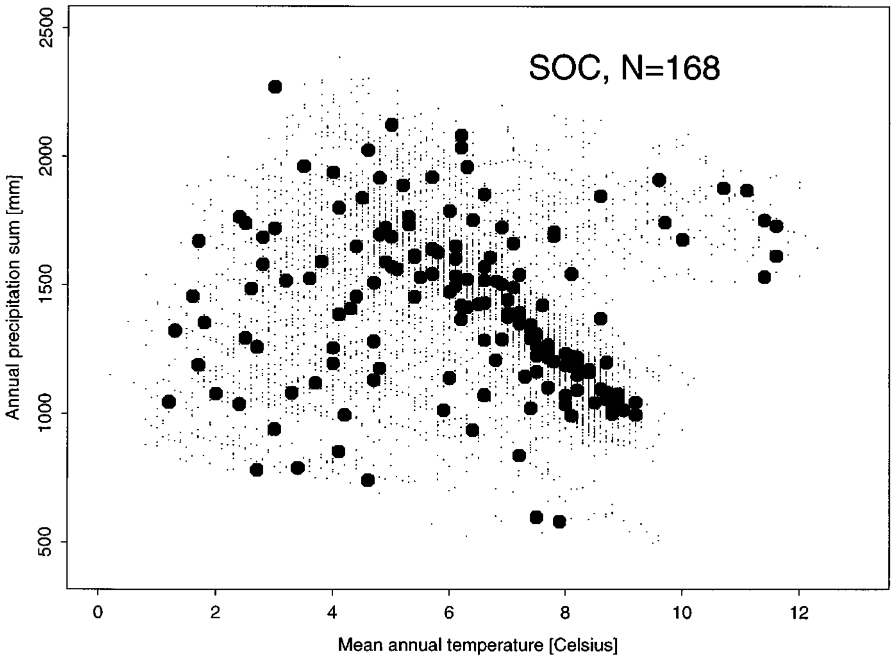

Figure 2. Distribution of soil samples (fille circles, $N = 168$) and forest inventory sampling units (f ne points, $N = 5298$) in temperature-precipitation space. Filled circles represent soil sites sampled at 8-km × 8-km resolution, f ne points were taken from the second National Forest Inventory SNFI 2 (Brassel & Brändli 1999) at a spatial resolution of 1.4-km × 1.4-km.

(Perruchoud et al. 1999b). SNFI 2 also specif es tree species with the largest basal area (referred to as dominant hereafter) and thus provided a simplif ed forest type classif cation by species. Additionally, the potential vegetation classif cation (Ellenberg & Klötzli 1972) was available for all soil sampling sites (P. Kull, WSL Birmensdorf, pers. comm.).

Laboratory analyses

Soil samples were dried to constant weight at 60 °C, passed through a 2-mm sieve and ground for chemical analyses. Total C concentration of the f ne earth fraction < 2 mm was determined with a Carlo Erba NA-1500 analyser using two replicates of each sample. For acidic soils (pH < 6) the concentration of inorganic C was assumed to be zero; for nonacidic soils, organic C concentration was determined by difference from total and inorganic C concentration. Inorganic C was measured directly by reacting soil with H_2SO_4 and weighing the evolved CO_2 previously collected in NaOH. For the conversion of % CO_3 (in $CaCO_3$) to % C, a factor of 1/5 was used.

The standard deviation of replicate C concentration measurements was ±3% of the measured value. Analytical limits of detection were 0.23% and 0.20% for $CaCO_3$ and total C respectively. Additionally, soil volume (i.e. 500 cm^3) and mass were measured for a subset of topsoil samples ($N = 138$, 0–10 cm soil depth). Soil density of f ne earth (fraction < 2 mm) was determined assuming negligible rock fraction for these topsoil layers and related to organic C concentration (see below).

Calculations and statistical analyses

SOC density. We assumed homogeneity of C concentration within sampling layers and interpolated C concentrations linearly between soil layers without measurement. Information about the C concentration from multiple measurements within the same pedogenic horizon were included. C content of the fraction ≥ 2-mm ($\delta_{i,2\mathrm{mm}}$ in %) was neglected (Zinke et al. 1986; McNabb et al. 1986; Homann et al. 1995) and SOC pool size (in [t C ha^{-1}]) determined via

$$\mathrm{SOC}_{d_z} = \sum_{i}^{d_z} \rho(\mathrm{C}_i) \cdot \left(1 - \frac{\delta_{i,2\mathrm{mm}}}{100}\right) \cdot d_i \cdot \mathrm{C}_i, \qquad (1)$$

where d_i denotes the thickness of layer i in cm, $\rho = f(\mathrm{C}_i)$ f ne earth density in $g \cdot cm^{-3}$, depending on C_i carbon concentration in % and $\sum_i^{d_z}$ integrates between the mineral soil surface and soil depth d_z. SOC estimates were calculated to maximum soil depth with detectable C concentration ($\mathrm{SOC}_{\mathrm{prof\,le}}$, $N = 168$) and – respecting the continuous sampling within the 0–20 cm layer – for a soil depth d_z of 20 cm (SOC_{20}, $N = 156$). Additionally, stand level means for the 0–30 cm mineral soil layer (SOC_{30}, $N = 137$) were determined to comply with the guidelines of the Intergovernmental Panel on Climate Change (Houghton et al. 1997).

Bulk density. For 138 sites, organic C concentration (in [%]) and soil density (ρ in [$g \cdot cm^{-3}$]) were simultaneously available from the topsoil (0–10 cm depth). The C concentration was averaged within this layer and related to ρ by non-linear regression with procedure NLIN (method Marquardt) (SAS 1990). We tested published relationships between ρ and organic matter (Curtis & Post 1964; Grigal et al. 1989) or C (Alexander 1980; Harrison & Bocock 1981; Huntington et al. 1989) for forest soils and selected the regression equation with the lowest residual sum of squares.

Regression analyses. Statistical analyses were carried out for total mineral soil SOC ($\mathrm{SOC}_{\mathrm{prof\,le}}$) and for 0–20 cm SOC ($\mathrm{SOC}_{20}$) with the S-Plus software package (S-Plus 1996). The 0–20 cm mineral soil layer was considered in particular, given the continuous soil sampling in this layer (see above), and because topsoils are thought to be affected by environmental changes most directly (Parton et al. 1993; Schimel et al. 1994; McGuire et al. 1995; Melillo et al. 1995). Simple linear regressions were f tted using continuous climatic, topographical, and vegetation variables as predictors. In addition, categorial variables for forest productivity region, forest type, soil unit, as well as soil shallowness, rock fraction and water holding capacity were used.

Moreover, multiple linear regression models were f tted involving different sets of the above continuous predictor variables and combinations of continuous and categorial variables. Predictors were analysed with respect to correlations and variables x, y not included simultaneously if $|cor(x, y)| > 0.85$. Model selection was done on the basis of multiple R-square and p-values for the f tted regression models. Normality of the residual distribution was checked via T3 plots (Ghosh 1996) and equality of variances by means of Tukey-Anscombe plots. Exploratory analysis showed natural logarithm to be the appropriate transformation on all continuous variables in the regression models. Description and summary statistics of the variables used in the statistical analyses are given in Table 1 and Table 2.

Geostatistical analyses. Spatial dependencies of $SOC_{prof le}$ and SOC_{20} were analysed with the S-Plus software package S+SPATIALSTATS (function "variogram") (Kaluzny et al. 1996) using directional and omnidirectional, empirical variograms, i.e.,

$$\gamma(\mathbf{h}_{kl}) = \frac{1}{2|N(\mathbf{h}_{kl})|} \sum_{i,j \in N(\mathbf{h}_{kl})} [\mathrm{SOC}_{d_z}(\mathbf{x}_i) - \mathrm{SOC}_{d_z}(\mathbf{x}_j)]^2,$$

where $|N(\mathbf{h}_{kl})|$ is the number of pairs (i, j) whose distance $\mathbf{x}_i - \mathbf{x}_j$ lies within lag class $\mathbf{h}_{kl}$, characterised by a ring segment of radius $2dh$ and angle $2d\phi$. Additionally, semivariograms were used to study the spatial dependence of the residuals obtained from the above regression analyses.

National SOC extrapolation. For extrapolation of SOC to national scale we used an area of 1.1399 Mha corresponding to accessible forests without brushwood in Switzerland and accounting for 92% of the total forest area in Switzerland (Brassel & Brändli 1999). National SOC estimates were calculated by combining area estimates of forest type classes (Stierlin & Ulmer 1999) with SOC densities calculated after Eq. (1) and using Cochran (1977)'s formulae for a stratif ed population's total, mean and variance, i.e.,

Table 1. List of variables used for statistical analysis.

	Continuous variables	
$SOC_{profile}$	Soil organic carbon in the entire mineral soil	t C ha^{-1}
SOC_{20}	Soil organic carbon in topsoil	t C ha^{-1}
toc	Total tree carbon	t C ha^{-1}
elev	Elevation	m (a.s.l.)
slope	Slope	° ($\in[0°, 90°]$)
aspect	Aspect	° ($\in[0°, 180°]$)
mat	Mean annual temperature	°C
map	Annual precipitation sum	mm
aet	Actual evapotranspiration	mm
fcap	Field capacity	mm
	Categorial variables	
eku	Vegetation units	
forest	Forest type	12 classes
region	Productivity region	5 classes
fao_{soil}	FAO soil units	9 classes
shallow	Soil shallowness	5 classes
rock	Rock fraction	5 classes
whc	Water holding capacity	6 classes

Vegetation units refer to the classif cation system of Ellenberg and Klötzli (1972) and were provided by P. Kull (WSL Birmensdorf, pers. comm.). Forest types were def ned by dominant tree species and derived from the Swiss national forest inventory SNFI 2 (Brassel & Brändli 1999).

$$\widehat{Y_{st}} = \sum_{l=1}^{L} N_l \bar{y}_l, \qquad \text{and}$$

$$\mathrm{var}\langle\widehat{Y_{st}}\rangle = \sum_{l=1}^{L} \frac{N_l}{n_l}(N_l - n_l)\left(\frac{1}{n_l - 1}\sum_{k=1}^{n_l}(y_{lk} - \bar{y}_l)^2\right),$$

where

$$\bar{y}_l = \frac{1}{n_l}\sum_{k=1}^{n_l} y_{lk}$$

is the sample mean in stratum l, L is the number of strata, N_l is the total number of 1-ha area units and n_l the respective number of sampled area

Table 2. Summary statistics of soil, vegetation, and site properties.

Parameter	Symbol	Minimum	Maximum	Mean	CV [%]	*N*
Soil organic C [t C ha^{-1}]	$SOC_{profile}$	6.2	319.0	98.2	70	168
Soil organic C (0–20 cm) [t C ha^{-1}]	SOC_{20}	6.2	174.6	62.0	46	156
Solum thickness [cm]	d_z	5	190	61	58	168
Total tree C [t C ha^{-1}]	toc	3.7	506.8	127.7	62	160
Elevation [m a.s.l.]	elev	322	2207	1032	44	168
Slope [°]	slope	1	60	23	54	165
Aspect [°]	aspect	0.5	178.9	78.6	65	165
Temperature [°C]	mat	1.2	12.0	7.1	31	168
Precipitation [mm]	map	581	2271	1399	23	168
Evapotranspiration [mm]	aet	398	700	531	12	168
Field capacity [mm]	fcap	0	355	94	64	164

Solum thickness refers to the deepest soil layer with detectable amounts of carbon. Total tree C includes foliage, twigs, branches, stem, coarse and fine roots and was estimated from forest inventory data (Kaufmann & Brassel 1999; Perruchoud et al. 1999b).

units in stratum l. This approach was applied for total mineral and top soil ($y_{lk} = SOC_{profile,lk}$, $SOC_{20,lk}$, and $SOC_{30,lk}$) in Swiss forests using statistically significant strata.

Alternatively, y_{lk} was predicted pointwise by regression using site-specific continuous (x) and categorial variables (z). Since in particular SOC was transformed using the natural logarithm function, the model estimate $\mu_i = \ln(SOC_i) = \beta_0 + \sum_j \beta_j \ln(x_{ji}) + \sum_k \beta_k \ln(z_{ki})$ was used to determine SOC stocks for every sample site i via $\mathcal{E}\langle SOC_i \rangle = e^{\mu_i + \sigma^2/2}$ (Finney 1941) in the topsoil layer (0–20 cm) and the entire mineral soil. The same parameter σ^2 was assumed for all sites and estimated by the residual mean square error of the ln(SOC)-model. For this approach no variance of the population total was calculated.

Results

Data analysis: Stand level

The best fit between organic C and topsoil density ρ on the Swiss forest soil samples was obtained using the function proposed by Grigal et al. (1989) (Table 3, Figure 3) with a residual sum of squares of 2.95, i.e.,

$$\widehat{\rho}(C) = 0.44(\pm 0.08) + 0.65(\pm 0.07) \cdot e^{-0.14(\pm 0.05)\cdot C}, \tag{2}$$

Table 3. Model equations and coeff cients for determining soil density (in [g·cm^{-3}]) from organic C concentration (in [g C/(100 g soil)]).

Regression equation	*RSS*	Parameters			Reference
		b_0	b_1	b_2	
$\rho = b_0 + b_1 \cdot \sqrt{C}$	3.07	1.11	−0.15	×	Alexander (1980)
$\rho = b_0 + b_1 \cdot log_{10}(C)$	3.19	1.03	−0.40	×	Harrison and Bocock (1981)
$\rho = b_0 + b_1 \cdot e^{b_2 \cdot C}$	2.95	0.44	0.65	−0.14	Grigal et al. (1989)
$\rho = e^{b_o + b_1 \cdot ln(C) + b_2 \cdot ln^2(C)}$	2.97	0.04	−0.11	−0.05	Huntington et al. (1989)

Soil density is denoted by ρ, organic C concentration by C, and *RSS* refers to the residual sum of squares.

(parameters in parentheses refer to the upper and lower 95% conf dence bands). Eq. (2) was applied at all soil depths and combined with Eq. (1) to determine layer-specif c SOC densities. Thus, a mean SOC stock of 98.2(±68.8) t C ha^{-1} (± standard deviation) was obtained for the entire mineral soil prof le (Table 2, Figure 4, top). Sample locations and magnitude of $SOC_{prof\,le}$ are shown in Figure 1 (top).

We compared C stocks sequestered in mineral soil and trees including above- and belowground woody and f ne plant parts. In 38% of the 160 forest stands for which SOC and tree C data were simultaneously available, SOC was the larger pool. For 29% of the stands, tree C stocks exceeded SOC by less than a factor of 2. In summary, 44% of total forest C (excluding herbaceous vegetation, litter and organic horizons) was stored in the mineral soil. An analysis of variance revealed no signif cant differences of the tree-soil C distribution among productivity regions, soil types or vegetation units sensu Ellenberg and Klötzli (1972). However, forest types differed strongly in $SOC_{prof\,le}$ owing to the presence of larch stands ($N = 11$). For these high-elevation stands (> 1500 m a.s.l.), mean SOC was substantially higher than tree C stocks, but with a high coeff cient of variation (CV > 100%).

Stratif cation of $SOC_{prof\,le}$ by FAO soil taxa revealed substantial, but insignif cant variations ($\alpha = 0.05$) in contrast to classif cation by forest type, Ellenberg and Klötzli (1972)'s vegetation units and productivity region. For forest types, this effect was caused by the high SOC density in chestnut forests ($N = 6$). Substantially higher SOC stocks could also be assigned to Southern Alpine forest soils. The coeff cients of variations ranged between 41% and 71% in these classes (Table 4).

Linear regression was used to identify the effect of site parameters on C sequestration in forest soils. Fitting models of ln($SOC_{prof\,le}$) against log-transformed continuous predictors showed that only annual precipitation sum ($R^2 = 0.07$), and slope ($R^2 = 0.06$) were signif cant at the 1% level.

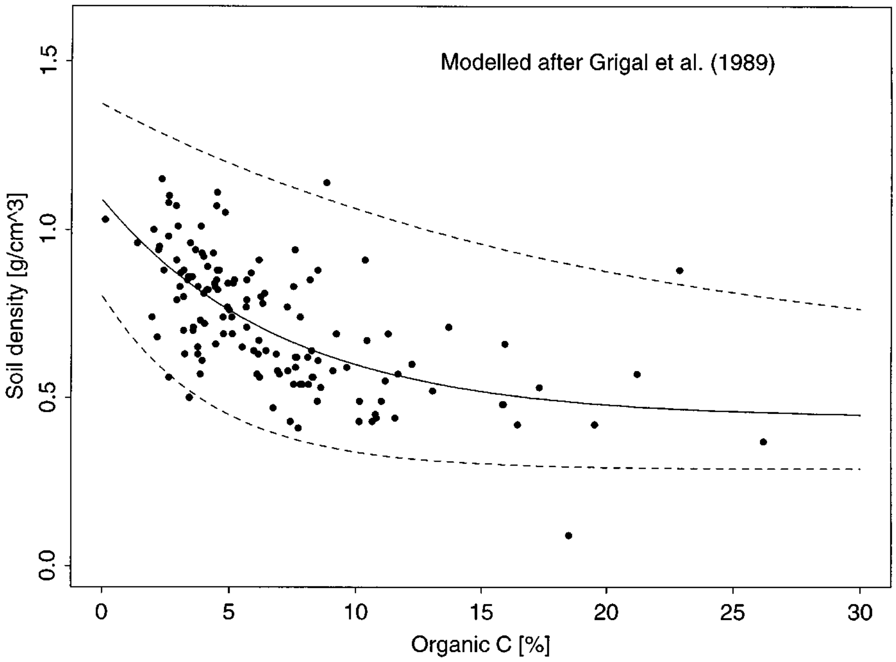

Figure 3. Organic C concentration vs. soil density for 138 forest mineral topsoil layers (0–10 cm). C concentrations were measured as [g C/(100 g soil)] and soil density as [g·cm^{-3}] and simultaneously available for a subset of the forest damage inventory prof les. The solid line represents the best model f t Eq. (2), dashed lines refer to the 95% conf dence bands.

Subsequently, the simultaneous control of continuous variables was checked by multiple regression including predictors with correlation below 0.85, i.e., map, mat, slope, aspect, and toc. Although highly signif cant models were obtained, neither these models nor additions of second order interaction terms explained more than 20% of the SOC variability in our data. The best model for $SOC_{\text{prof le}}$ was obtained by including additional effects of the categorial variables forest type and productivity region. The f nal model used to estimate $SOC_{\text{prof le}}$ at national scale

$$\begin{aligned}\ln(SOC_{\text{prof le}}) &= \beta_0 + \beta_1 \cdot \ln(\text{map}_i) + \beta_2 \cdot \ln(\text{slope}_i) + \\ &\quad \beta_3 \cdot \ln(\text{mat}_i) + \beta_4 \cdot \text{forest}_i + \beta_5 \cdot \text{region}_i + E_i, \qquad (3)\end{aligned}$$

was highly signif cant ($p_F < 0.01$) with $R^2 = 0.24$. The analysis of residuals showed that their expectation $\mathcal{E}\langle E_i \rangle$ was close to zero with constant variance var$\langle E_i \rangle$ and near normal distribution at the 95% acceptance level.

Mean SOC stocks in the top 0–20 cm soil layer amounted to 62.0 (± 28.5) t C ha^{-1} (Figure 1, Figure 4). Classif cation of SOC_{20} by forest type and productivity region were statistically signif cant at the 5% level due to the

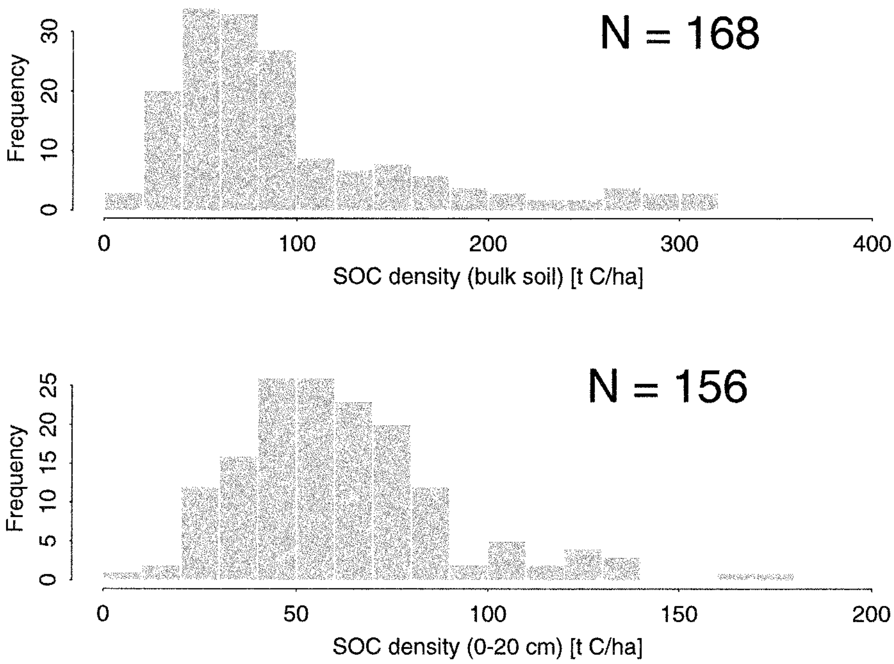

Figure 4. Histograms of SOC in total (top, $N = 168$) and 0–20 cm mineral soil layers (bottom, $N = 156$) for Swiss forest soils. Mean SOC density is 98.2 t C ha^{-l} (CV: 70%) for total mineral soil and 62.0 t C ha^{-l} (CV: 46%) for the topsoil. Note the different scales in the two fi ures.

presence of forest stands dominated by chestnut ($N = 6$) and samples in the Southern Alps ($N = 18$) (Table 4). Coeff cients of variation varied between 21% and 47% for SOC_{20}. For SOC_{20}, linear regression on double log-transformed data showed annual precipitation to be the only signif cant predictor ($R^2 = 0.14$, $p_F < 0.05$). Additionally, we found signif cance for slope at the 10% level and for mean annual temperature at the 15% level, but both with weak inf uence. Examination of T3 plots (Ghosh 1996) for the residuals showed, however, that neither neither simple, multiple regression nor interaction term models for SOC_{20} met the normality condition at the 95% conf dence level.

Thus, soil samples from broadleaf and needleleaf forest stands were analysed separately. For the 56 broadleaf stands, precipitation, slope and elevation were the only signif cant single continuous predictors ($\alpha = 0.10$). Forest type and productivity region improved the model substantially and

$$\begin{aligned} \ln(\mathrm{SOC}_{20,\mathrm{bl}}) &= \beta_0 + \beta_1 \cdot \ln(\mathrm{map}_i) + \beta_2 \cdot \ln(\mathrm{elev}_i) + \\ &\quad \beta_3 \cdot \ln(\mathrm{slope}_i) + \beta_4 \cdot \mathrm{forest}_i + \beta_5 \cdot \mathrm{region}_i + E_i \end{aligned} \tag{4}$$

Table 4. National estimate of SOC in total and top mineral soil layers of upland accessible forests in Switzerland.

Model	Forest area [Mha]	Stand level SOC [t C ha^{-1}]	Stratified mean SOC [Tg C]	SOC in Swiss forests Mean [Tg C]	SOC in Swiss forests Stand. error [Tg C]
		Total mineral soil, $SOC_{profile}$			
All	1.1399	98.2	–	112.0	6.0
Chestnut	0.0249	177.0	4.4	111.1	6.2
Other forest types	1.1150	95.7	106.7		
South Alps	0.1412	149.3	21.1	112.3	5.9
Other regions	0.9987	91.3	91.2		
Eq. (3)	1.1399	100.8	–	114.9	–
		Topsoil (0–20 cm), SOC_{20}			
All	1.1399	62.0	–	70.7	2.6
Chestnut	0.0249	98.9	2.5	69.7	2.5
Other forest types	1.1150	60.3	67.2		
South Alps	0.1412	85.2	12.0	70.9	2.4
Other regions	0.9987	59.0	58.9		
Broadleaf, Eq. (4)	0.3928	66.4	26.1	71.2	–
Needeleaf, Eq. (5)	0.7471	60.4	45.1		

National estimates refer to accessible forest without brushwood. Mean SOC was extrapolated from stand level SOC and estimates about the areal extent of strata (Stierlin & Ulmer 1999) assuming homogeneity within strata. The standard error of the population total $\widehat{Y_{st}}$ was calculated as $(\mathrm{var}\langle\widehat{Y_{st}}\rangle)^{1/2}$. A detailed description of the methods used for estimating SOC by linear regression (Eq. (3), Eq. (4), and Eq. (5)) is given in the text.

explained 59% of the data variability ($p_F \ll 0.01$, $df = 42$). The analysis of residuals showed an expectation value $\mathcal{E}\langle E_i \rangle$ close to zero with constant variance $\mathrm{var}\langle E_i \rangle$ and normal distribution. Plots of the residuals against predictors (not shown) revealed no need for additional parameter transformations.

For needleleaf forests ($N = 93$) precipitation, map, was the only significant, even though weak predictor ($R^2 = 0.06$, $p_F < 0.05$). Neither slope, elevation, temperature nor any of the other continuous and categorial variables reached a significance level of 10%. Various combinations of continuous and categorial variables were tested, but none met the normality

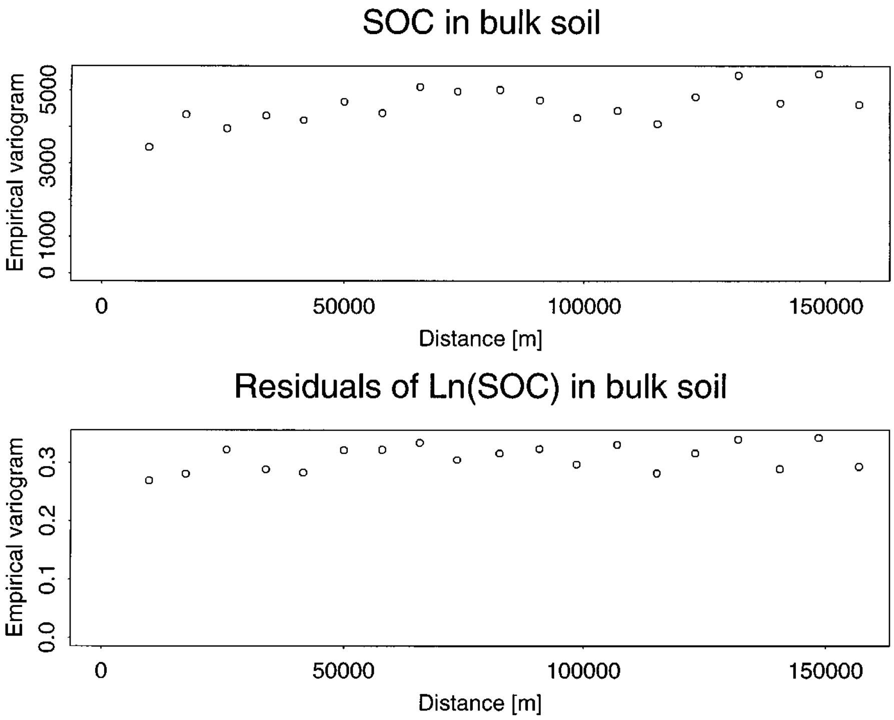

Figure 5. Empirical variograms for $\mathrm{SOC}_{\mathrm{prof\ le}}$ (top) and residuals of $\ln(\mathrm{SOC}_{\mathrm{prof\ le}})$ for regression model Equation (3) (bottom). See text for a defi ition of the empirical variogram. The sampling scale is 8 km for our data set. Both variograms show lack of clear spatial dependence at the scale of investigation.

T3 test for the residuals. However, removal of the needleleaf stands with the highest and lowest SOC stock improved model f ts and distribution of residuals considerably. The best model for needleleaf forest samples then was

$$\begin{aligned}\ln(\mathrm{SOC}_{20,\mathrm{nl}}) &= \beta_0 + \beta_1 \cdot \ln(\mathrm{map}_i) + \beta_2 \cdot \ln(\mathrm{mat}_i) + \\ &\quad \beta_3 \cdot \mathrm{forest}_i + \beta_4 \cdot \mathrm{region}_i + E_i, \end{aligned} \tag{5}$$

with $R^2 = 0.23$, $p_F < 0.05$ and 77 degrees of freedom. We subsequently derived a national estimate of SOC_{20} in needleleaf forests by means of Equation (5).

Geostatistical analyses revealed no clear spatial trends for SOC in Switzerland (see Figure 5). Similarly, at a sampling scale of 8 km, no spatial dependencies emerged for the residuals of any of the above regression models (Equations (3)–(5)).

Data analysis: National extrapolation

National SOC estimates in accessible Swiss forests (1.1399 Mha) ranged between 111 Tg C and 115 Tg C for mineral soil and were around 70 Tg C in the 0–20 cm topsoil mineral layer (Table 4, 1 Tg = 1 Mt = 10^{12} g). Based on a stand-level stock for SOC_{30} of 75.9 t C ha^{-1} we obtained a total of 87 Tg C in mineral soils at 0–30 cm depth. Accounting for the somewhat larger total area of Swiss forests, i.e. 1.2340 Mha (Strobel et al. 1999), we obtained a SOC estimate of 120–125 Tg C in the entire mineral soil and about 95 Tg C (0–30 cm soil depth) and 76 Tg C (0–20 cm) in mineral topsoil for the year 1993. Noticeable is the small difference among our national SOC estimates independent of the calculation approach and that stratif cation did not increase the precision of our estimate. Standard errors of the total were around 6 Tg C for total mineral soil, and 3.5 Tg C respectively 2.5 Tg C for the 0–30 cm and 0–20 cm topsoil layers. The proportionally lower standard errors of the estimated topsoil totals ref ect the smaller variation found at stand level for the 0–20 cm layer and the variability of soil depth on our sites (see Table 2).

Discussion and conclusions

The main objective of this study was to assess national SOC stocks in Switzerland for the year 1993. For accessible forests without brushwood in Switzerland (1.1399 Mha), our best estimate was around 110–115 Tg C for total mineral soil, respectively 87 Tg C and 70 Tg C for the 0–30 cm and 0–20 cm topsoil layer with very minor differences among the different approaches (Table 4). Paulsen (1995) has estimated the corresponding organic matter (OM) stock of foliar, nonfoliar litter (fruits and twigs with diameters below 2 cm), and organic horizons in Swiss forest soils with independent, random samples ($N = 236$) at 26.3 Tg OM or 13.2 Tg C (Vogt 1991) for the corresponding time period. Thus, 13% of the forest soil C reservoir (organic plus mineral horizons without coarse woody debris) is found on the forest f oor in Switzerland, in close agreement with an estimated 18% contribution in German forest soils (BML 1997).

Even without organic horizon C, mineral soils sequester a substantial fraction of the C in mountain forests: Based on 5298 forest inventory plots (Brassel & Brändli 1999), a mean tree C stock (above- plus belowground) of 119 t C ha^{-1} was obtained using the approach of Perruchoud et al. (1999b) corresponding to 136 Tg C for trees in accessible forests for Switzerland in 1993. Hence, 45% of the total forest C (excluding herbaceous vegetation, litter and organic horizons) is sequestered by mineral soils. This agrees with the C distribution found on our 160 soil sample plots and the range for plant biomass and soil organic matter (SOM) of “cold temperate needle-leaved evergreen” (34%) and “cold temperate broadleaf deciduous” (48%) forests (Vogt et al. 1995).

Our data can probably not account for the large landscape heterogeneity, but data density (168 soil samples for roughly 1 Mha) was high compared to previous assessments of SOC: Kauppi et al. (1997) based their SOC estimate for 14 Mha of upland forest soils on data from 377 soil prof les. For German forests (10 Mha), SOC stocks were determined from 103 forest soil samples (Burschel et al. 1993) and Siltanen et al. (1997) used 1462 soil prof les to estimate SOC of Canada’s forests (404 Mha (Kurz et al. 1995)). Our soils were representative for the climatic conditions of forests in Switzerland in terms of the distribution and frequency in the temperature-precipitation space (Figure 2). Moreover, sample sites including 34% broadleaf and 66% needleleaf stands agreed well with forest inventory data with respect to dominant species and regional distribution of soil sampling sites (Stierlin & Ulmer 1999). Uncertainty remained as to how representative our soil samples are with respect to the depth distribution of Swiss forest soils across the Alps.

National SOC totals were aff icted with considerable uncertainty due to intersite variability of SOC. For total mineral soil (without stratif cation) upper and lower conf dence limits of 124 Tg C and 100 Tg C were obtained. The respective conf dence limits for the 0–20 cm topsoil layer were 76 Tg C and 66 Tg C respectively. These conf dence intervals encompass the population total in 95% of the time in case of repeated sampling. How likely are changes in SOC induced by environmental changes to be detected given the precision of our SOC estimate? Perruchoud et al. (1999a) have given evidence that a 10% increase in topsoil SOC stocks of temperate forests require more than 10 years even if litter inputs are instantaneously enhanced by one third. Slower responses can be expected for subsoil layers and gradual or net changes in SOC stocks could even be masked if e.g. higher litter inputs were accompanied by enhanced microbial activity in a warmer climate. Land-use change would change topsoil SOC more rapidly (Davidson & Ackerman 1993; Harrison et al. 1995; Pennock & Van Kessel 1995), but the lack of C estimates for organic forest soil horizons and the change in soil mass neglected here (and in conventional soil inventories in general) render comparisons of SOC by pedogenetic horizons or f xed sampling depth incorrect (Ellert & Bettany 1995). Dynamic SOC models driven by high-resolution forest inventory data (Perruchoud et al. l999b) could present an alternative for quantifying SOC changes.

Although doubted by Post et al. (1982) due to high intrasite variability, soil sampling at higher spatial resolution could improve future, regional SOC assessments. This would, however, involve much higher efforts and costs:

McNabb et al. (1986) reported 20 samples per 0.25-ha plot to be insuff cient to determine SOC stocks in the 0–15 cm layer within ±10% at the 95% probability level in mountain forests. Huntington et al. (1988) found that on an area of 23 ha about 60 pedons were needed to detect future SOC changes of 10% in the mineral soil of a northern hardwood forest. Rahman et al. (1996) observed no relation between C concentration and terrain attributes and only weak spatial dependence for solum thickness and rock fraction in A and B horizons of Rocky Mountains forest soil despite of a soil monitoring at 200 m intervals. In our study which used a spatial resolution of 8-km × 8-km, analysis of the variograms revealed no spatial correlation or long-range processes.

Forest SOC stocks in the Swiss Alps revealed no strong climatic or topographic signature conf rming earlier observations of forest SOC in the Bavarian mountains (Germany) (Haber 1985). In grasslands mean annual temperature was the main control on SOC stocks and high temperature was related to enhanced microbial respiration and hence low SOC stocks (Burke et al. 1989). However, the opposite was observed for forest soils in the Western US and Scandinavia where larger SOC pools were found at higher temperatures (Homann et al. 1995; Kauppi et al. 1997). For Swiss forest soils no signif cant relations were obtained between mineral soil SOC and temperature (mat, mat^2) or actual evapotranspiration (aet, aet^2) despite the latter's wide range of values. Annual precipitation was a highly signif cant, though weak predictor and positively related to SOC in our data as previously stated for grassland and forest soils (Burke et al. 1989; Homann et al. 1995). High SOC stocks could be explained by reduced OM mineralisation in landscape positions with limited soil aeration or external surface drainage (Davidson 1995; Pennock & Van Kessel 1995). This was, however, not supported here, since soils with poor drainage (Gleysols) or higher f eld capacity did not reveal higher SOC accumulation.

Classif cation of SOC stocks by soil texture based on *in situ* estimates ('Fingerprobe') revealed no signif cant SOC variations in our data set. Thus, soil texture (Arrouyas et al. 1995; Davidson 1995) and mineralogy (Torn et al. 1997) determining the presence of cation bridges between clay plates, biopolymers and soil microorganisms, as well as occlusion of organic matter within aggregates and thus microbial SOM decomposition (Oades 1995) did not express itself at SOC stock level. However, particularly high SOC stocks were obtained for chestnut forests in the Southern Alps. Blaser et al. (1997) explained this by leaching of OM from litter with strong metal-binding properties, formation of organo-metallic complexes in Fe- and Al-rich parent material and a concomitant reduction of substrate decomposability. A stabilisation of SOM over limestone bedrock (Jura) associated with binding of OM in interlayers of clay minerals was not observed in our data.

We see several possible reasons for the limited success of our statistical modelling besides natural intrasite variability. First, although SOC stocks are determined by soil and vegetation processes (e.g. SOM decomposition vs. primary productivity and litterfall), their differential response across environmental gradients (Raich & Potter 1995; Matthews 1997) could not be estimated. Second, ecosystem disturbances and soil age largely affect SOC stocks and their variability (Schlesinger 1990; Torn et al. 1997; Liski et al. 1998). However, reliable data on stand history (i.e. soil and stand age, f re frequency, insect calamities, or timber harvesting) were missing to test such effects on SOC stocks at our sites. Third, C in organic surface horizons which characterises stand and environmental conditions in coniferous forests (Haber 1985; Gärdenäs 1998) and contributes substantially to SOC in those forests was not included in the Swiss forest damage inventory (Vanmechelen et al. 1997). A fourth reason are combined uncertainties in C concentration, soil density and rock volume. The latter two had not been monitored and, thus, rock volume was classif ed and soil density estimated by regression from a subset of topsoil samples (0–10 cm) as done by Arrouays et al. (1998), Homann et al. (1998) or Rapalee et al. (1999). Given the unrepresented higher density of deep soil layers, our SOC estimate at national scale thus is probably "reasonable, if conservative" Siltanen et al. (1997).

Quantif cation of carbon reservoirs in forest ecosystems and SOC in particular remains an important element in deciphering the C cycle puzzle. Little is known about the eco-processes inducing the complex patterns of C storage between organic and underlying mineral soil horizons (Haber 1985; Grigal & Ohmann 1992) or the formation of humif ed soil organic matter in forests (Aber et al. 1990). The key role of soil fauna is recognised, but quantif cation of the incorporation of forest f oor organic matter into the mineral soil remains highly speculative. Process-analysis is one way to improve our understanding of ecosystem functioning. Assessments of the ecosystem's state, such as estimating SOC at national scale, provides complementary information which is needed for calibration, verif cation and application of simulation models. To comply with these requirements, future assessments of C in forest soils should simultaneously sample organic or mineral soil horizons, and focus on an assessment of the intrasite variability of forest soil C stocks.

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