DIVISION S-6—SOIL & WATER MANAGEMENT & CONSERVATION

Changes in Soil Organic Matter Fractions under Subtropical No-Till Cropping Systems

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ABSTRACT

Conservation management systems increase soil C and N pools. However, their effects on particulate (>53 μm) and mineral-associated (<53 μm) soil organic matter (SOM) fractions are less understood under subtropical climatic conditions. This study evaluated the long-term (12-yr) effects of three no-till cropping systems {bare soil, BS; oat (*Avena strigosa* Schreb.) + vetch (*Vicia sativa* L.)/maize (*Zea mays* L.) + cowpea (*Vigna unguiculata* L. Walp.), O + V/M + C; and maize + *Cajanus* [*Cajanus cajan* (L.) Millsp.], M + C} on C and N pools in particulate and mineral-associated SOM. The study was performed in southern Brazil, on a sandy clay loam Acrisol. Cropping systems that included cover crops increased C and N pools in both particulate and mineral-associated SOM when compared with BS. Mineral-associated SOM had five to nine times more C and 13 to 26 times more N than particulate organic matter and was responsible for 69 to 80% of total atmospheric CO_2 sequestred by soil in O + V/M + C (38 Mg ha^{-1}) and M + C (51 Mg ha^{-1}). The higher C and N pools were associated with greater recalcitrance of mineral-associated SOM to biological decomposition, resulting from its interaction with variable charge minerals. The negative relationship between decay rates of SOM and the concentrations of Fe oxides and kaolinite demonstrated the physical stability of SOM caused by interaction with variable charge minerals. Power saturation curves of electron spin resonance (ESR) spectroscopy in the 20- to 53-, 2- to 20-, and <2-μm granulometric fractions also reinforced this hypothesis. The SOM interaction with variable charge minerals plays an important role in preserving SOM storage, enhancing the potential of tropical and subtropical soils to act as an atmospheric CO_2 sink.

S OIL ORGANIC MATTER decomposition rates are nearly five times greater in warm, wet tropical and subtropical regions than under temperate conditions (Sanchez and Logan, 1992). Decomposition rate differences are higher in agricultural soils where soil aggregation and SOM physical stability decrease with increased soil disturbance (i.e., tillage) (Carter et al., 1998). Consequently, SOM losses equivalent to one half of the natural pools have been observed within a 10-yr period in the tropics because of biological oxidation and erosion (Tiessen et al., 1994). The rapid SOM shortage results in decline of soil quality and crop productivity, bringing negative social and environmental implications.

The study and implementation of soil conservation management systems in southern Brazil began in the 1980s to reduce widespread soil degradation and erosion. Tillage intensity reduction and the use of cropping systems that maximize residue addition to soil surface have been efficient agricultural practices to maintain or increase SOM (Bayer et al., 2000a,b,c). The rate of SOM accumulation depends on inherent soil and regional climatic characteristics, such as texture, mineralogy, and temperature (Bayer, 1996; Alvarez and Lavado, 1998). SOM accumulation rates as high as 1 Mg C ha^{-1} yr^{-1} have been reported in the warm, wet subtropical regions of Brazil under no-tillage and with cropping systems using cover crops (Bayer, 1996).

There are ~2.2 million hectares under the no-tillage system in Rio Grande do Sul State, southern Brazil (Denardin et al., 1997), representing about 28% of the State's total cultivated area. Nearly 6 million hectares were under the no-tillage system in the whole country during the 1996–1997 growing season (Saturnino and Landers, 1997). The use of no-tillage system has increased mainly within the last decade, and a significant contribution towards reducing soil CO_2 emissions is anticipated (Bayer et al., 2000c).

Soil conservation management systems probably increase both labile (young, light, coarse and particulate) and humified SOM fractions at different rates. Studies developed under cold or semiarid climatic conditions have determined that SOM changes occur primarily in the labile fractions (Janzen et al., 1998; Chan, 1997). In these regions, particulate SOM represents ~50% of total SOM (Franzluebbers and Arshad, 1997; Chan, 1997).

The particulate SOM pool is smaller in warm, wet climates than under temperate conditions because of restrictions imposed by greater biological activity. We hypothesize that, in tropical and subtropical soils, interaction with variably charged minerals can result in high recalcitrance of mineral-associated SOM to biological decomposition, with this fraction being higher under conservation management systems than the particulate SOM fraction. To test this hypothesis, we conducted a study to (i) evaluate the long-term (12-yr) effects of three no-till cropping systems on C and N pools in particulate and mineral-associated SOM in a sandy clay loam Acrisol under southern Brazilian subtropical conditions; and (ii) identify the relationship of Fe oxides and kaolinite concentrations with mineral-associated SOM stability.

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Abbreviations: BS, bare soil; ESR, electron spin resonance; O + V/M + C, oat + vetch/maize + cowpea; M + C, maize + Cajanus (Cajanus cajan); SOM, soil organic matter; TOC, total organic C; TN, total N.

Table 1. Cropping systems and annual amounts of C and N added by aboveground parts of winter cover crops and the maize residues plus the aboveground parts of summer intercrops (adapted from Bayer et al., 2000a).

Cropping systems	Winter	Summer	Annual inputs	
			C	N
			Mg ha^{-1}	kg ha^{-1}
BS†	Fallow (bare soil)	Maize (*Zea mays* L.) (1988 and 1993)	0.47	11
O+V/M+C	Oat (*Avena strigosa* Schreb) + Vicia (*Vicia sativa* L.)	Maize + cowpea [*Vigna unguiculata* (L.) Walp]	5.61	212
M+C	Cajanus [*Cajanus cajan* (L.) Milsp.]	Maize + Cajanus	6.27	286

† Shallow plots cultivated for weed control in 1983–1987; after that no-till, same as all other treatments since beginning the experiment.

MATERIAL AND METHODS

Site Description

The experimental site is located at the Experimental Station of the Federal University of Rio Grande do Sul (30°50′52″ S, 51°38′08″ W). The soil is a sandy clay loam Acrisol (Typic Paleudult by US and Dark Red Podzolic by Brazilian taxonomies) with 540, 240, and 220 g kg^{-1} of sand, silt, and clay, respectively, on a dry soil basis. The main minerals in the clay fraction are Fe oxides (109 g Fe$_2$O$_3$ kg^{-1}) and kaolinite (720 g kg^{-1}). The regional climate is subtropical, with an annual average temperature of 19.4°C ranging from 13.9 to 24.9°C. Annual mean rainfall is 1440 mm, varying monthly from 95.7 to 168 mm.

Experimental Design and Treatments

This study was started in 1983. The experiment was conducted using a randomized complete block design with three field replications for each treatment. Three no-till cropping systems were evaluated (O + V/M + C, M + C, and BS). Each plot had a 20-m^2 experimental area (4 by 5 m).

At the beginning of the experiment (1983), the soil was degraded and compacted because of 13-yr period of intensive conventional tillage cultivation. To mitigate soil compaction, the whole experimental area was deep plowed before starting this study.

Crop residue addition of each system was monitored during the whole experimental period (12 yr) (Bayer et al., 2000a), presenting a wide variation of C (0.47–6.27 Mg ha^{-1} yr^{-1}) and N (11–286 kg ha^{-1} yr^{-1}) inputs to the soil (Table 1). Additional information on historic and experimental procedures can be obtained in Burle et al. (1997) and Bayer et al. (2000a). Soil chemical, physical, and biological characteristics are presented in Table 2.

Soil Sampling

Soil samples were collected in September 1994 from 0- to 25-, 25- to 75-, and 75- to 175-mm depths. Samples were taken from a random area of 0.05 m^2. All soil was removed manually, homogenized, and subsampled to ~1.0 kg of soil mass. Subsampled soil was then air dried for ~1 wk at 20 ± 5°C, manually crushed with a rolling pin, and sieved to pass through a 2-mm screen.

Soil Organic Matter Fractionating and Chemical Analysis

Two 20-g subsamples of soil were used in this study. The first was ground in a mortar to pass through a 0.2-mm sieve and analyzed for total organic C (TOC) and total N (TN) using the Walkley Black and Kjeldhal digestion procedures, respectively (Page, 1982). The second 20-g subsample was used to obtain particulate SOM by a procedure adapted from Cambardella and Elliot (1992). These subsamples were dispersed in 70 mL of distilled water with two glass beads and

shaken for 15 h in a reciprocal shaker. The soil suspension was passed through a 53-μm sieve to separate particulate SOM.

The soil fraction >53-μm was dried in a forced-air oven at 50°C, weighed to assess soil mass, ground with mortar and pestle, and analyzed for TOC and TN (Page, 1982). The C and N pools in particulate SOM fraction were calculated considering the mass of soil of this granulometric fraction plus TOC and TN concentrations. Compared with the procedure of Cambardella and Elliot (1992), this method showed a 6% particulate SOM loss. The C and N pools in the mineral-associated SOM were calculated considering the difference of their corresponding contents in total and particulate SOM. In all SOM fractions, the C and N pools were calculated based on an equivalent soil mass (Bayer et al., 2000a) from the soil bulk density data (Table 2).

To identify the relationship between Fe oxides, kaolinite concentration and mineral-associated SOM stability, the fraction <53-μm from 0- to 25-mm layer was further divided in three subfractions (20–53, 2–20, and <2 μm). Seventy milliliters of distilled water were added to a 100-mL snap-cap tube containing 20 g of <53-μm granulometric fraction obtained from the mixture of the three treatment replications. This suspension was sonicated at 240 W for 6 min (1114 J mL^{-1}) to cause soil dispersion. The 20- to 53-, 2- to 20-, and <2-μm granulometric fractions were separated through sedimentation in polietilene tubes, assuming a 2.65 g cm^{-3} mean particle density (Bayer et al., 2000a). The mass of granulometric fractions was quantified, and their TOC (Page, 1982), Fe oxides (Mehra and Jackson, 1960), and kaolinite concentrations analyzed. Kaolinite was semiquantified by integration of exothermic peaks in thermic differential analysis. Quantification was performed by comparison with a kaolinite standard (Bayer, 1996, p. 241).

The SOM stability was associated with relative SOM decay rates in granulometric fractions. Decay rates were calculated as the ratio of annual C input (Ak)/soil C pool (Cs) in the granulometric fractions, assuming first-order kinetics and C addition and loss equilibrium (Huggins et al., 1998). We also assumed that the C added was evenly distributed among all soil granulometric fractions considered in the study (>53, 20–53, 2–20, and <2 μm). Therefore, the annual C input (A) for each granulometric fraction was calculated as equivalent to 25% of the total C added (Table 1), considering an 0.2 humification coefficient (k) (Buyanovsky et al., 1997). The SOM decay rates were estimated taking into account only the soil C pool in the 0- to 25-mm layer. Therefore, the values reported herein have only a relative meaning among granulometric fractions. They are not comparable with decay rates commonly reported in the literature, which are calculated using the C pool in the whole plow layer (0–200 mm).

Power Saturation Curve Determination by Electron Spin Resonance Spectroscopy

The semiquinone free radical signal in ESR spectroscopy has been linked to the degree of SOM humification (Schnitzer

Table 2. Chemical, physical, and biological soil characteristics of the 0- to 175-mm layer of the sandy clay loam Acrisol of southern Brazil.

Soil characteristic	Depth mm	Cropping system		
		Bare soil	O+V/M+C	M+C
Soil bulk density, g cm^{-3}†	0–25	1.42	1.54	1.39
	25–75	1.58	1.58	1.41
	75–175	1.60	1.66	1.52
Macroaggregates, %†	0–25	75.0 ($\pm$1.0)#	80.7 ($\pm$2.2)	82.6 ($\pm$1.4)
	25–75	73.8 ($\pm$0.6)	79.3 ($\pm$0.9)	81.4 ($\pm$0.5)
	75–175	81.8 ($\pm$2.2)	81.8 ($\pm$0.3)	81.7 ($\pm$0.7)
Biomass-C, mg kg^{-1}‡	0–50	217 ($\pm$37)	409 ($\pm$64)	441 ($\pm$64)
Soil respiration, mg C-CO$_2$ kg^{-1}‡¶	0–50	108 ($\pm$26)	233 ($\pm$28)	287 ($\pm$29)
pH water§	0–25	5.6 ($\pm$0.1)	5.3 ($\pm$0.0)	5.3 ($\pm$0.1)
	25–75	5.2 ($\pm$0.1)	5.0 ($\pm$0.1)	5.1 ($\pm$0.1)
	75–175	5.2 ($\pm$0.1)	5.0 ($\pm$0.1)	5.0 ($\pm$0.1)
Ca exchangeable, cmol$_c$ kg^{-1}§	0–25	2.2 ($\pm$0.2)	3.1 ($\pm$0.1)	4.6 ($\pm$0.4)
	25–75	1.7 ($\pm$0.2)	2.1 ($\pm$0.1)	3.0 ($\pm$0.1)
	75–175	1.9 ($\pm$0.2)	2.0 ($\pm$0.1)	2.2 ($\pm$0.2)
Mg exchangeable, cmol$_c$ kg^{-1}§	0–25	1.5 ($\pm$0.2)	1.3 ($\pm$0.1)	2.0 ($\pm$0.1)
	25–75	1.0 ($\pm$0.2)	0.9 ($\pm$0.1)	1.2 ($\pm$0.1)
	75–175	1.0 ($\pm$0.2)	0.8 ($\pm$0.1)	1.1 ($\pm$0.1)
Al exchangeable, cmol$_c$ kg^{-1}§	0–25	0.1 ($\pm$0.0)	0.2 ($\pm$0.0)	0.1 ($\pm$0.0)
	25–75	0.4 ($\pm$0.1)	0.5 ($\pm$0.1)	0.3 ($\pm$0.1)
	75–175	0.5 ($\pm$0.1)	0.6 ($\pm$0.2)	0.7 ($\pm$0.1)
K exchangeable, mg kg^{-1}§	0–25	153 ($\pm$11)	156 ($\pm$10)	314 ($\pm$20)
	25–75	108 ($\pm$14)	112 ($\pm$7)	193 ($\pm$20)
	75–175	67 ($\pm$15)	66 ($\pm$4)	102 ($\pm$16)
CEC at pH 7.0, cmol$_c$ kg^{-1}§	0–25	6.7 ($\pm$0.2)	8.7 ($\pm$0.2)	11.9 ($\pm$0.8)
	25–75	6.1 ($\pm$0.1)	7.4 ($\pm$0.1)	9.1 ($\pm$0.3)
	75–175	6.2 ($\pm$0.1)	6.6 ($\pm$0.1)	7.3 ($\pm$0.1)
P, mg kg^{-1}§	0–25	21 ($\pm$3)	59 ($\pm$11)	90 ($\pm$7)
	25–75	12 ($\pm$3)	33 ($\pm$7)	50 ($\pm$5)
	75–175	5 ($\pm$1)	11 ($\pm$3)	15 ($\pm$3)

† Paladini and Mielniczuk (1991).
‡ Cattelan and Vidor (1990).
§ Burle et al. (1997).
¶ During 20 d.
# Values in parentheses represent the mean standard deviation.

and Levesque, 1977; Martin-Neto et al., 1998). In this study we hypothesized that the interaction between SOM and mineral-soil components affected the relaxation process of the semiquinone signal by ESR, which we checked by saturation power (from 0.05 to 50 mW) experiments. This power region was experimentally determined in granulometric fractions, where the presence of paramagnetic ions (mainly Fe^{3+}) limited detection of semiquinone ESR signal at higher values of microwave power.

The saturation curve relates square root of microwave power (abscissa) to the intensity of the semiquinone free radical signal (Sposito et al., 1996). Curves where the linearity between ESR signal intensity and square root of microwave power disappear at low values suggest systems with long relaxation time (Sposito et al., 1996; Czoch and Francik, 1989) and weak interaction of semiquinone free radicals with neighboring species. Curves where the linearity vanishes only at high levels of microwave power show a short relaxation time of the ESR signal, indicating strong interaction of semiquinone free radicals with neighbors. In organo-mineral aggregates, neighboring semiquinone free radicals are the SOM macromolecules themselves and associated soil mineral components.

Measurements of ESR were carried out in 20- to 53-, 2- to 20-, and <2-µm soil granulometric fractions from the M + C cropping system. A Varian ESR spectrometer, line E-109 Century (9 GHz) (Varian, Palo Alto, CA) was used under the following experimental conditions: central magnetic field (H_0) equals 340 mT; amplitude of modulation (AM) equals 0.2 mT; frequency of microwave equals 9.4 GHz; and room temperature.

Statistical Analysis

An analysis of variance was used to evaluate no-till cropping system effects on C and N pools in particulate and mineral-associated SOM fractions. Tukey's test was used to assess differences between treatment means at the 0.05 significance level.

RESULTS AND DISCUSSION

When the experiment started in 1983, the degraded soil contained 32.6 Mg C ha^{-1} and 2720 kg N ha^{-1} in the 0- to 175-mm layer (Bayer, 1996). Soil C and N pools did not change significantly throughout the 12-yr period under the BS system which can be verified by adding C and N pools in particulate and mineral-associated SOM in the 0- to 175-mm depth interval (Tables 3 and 4). The stable soil C and N pools of the BS treatment provided an opportunity to use this treatment as a reference when interpreting the effects of high residue addition cropping systems (O + V/M + C and M + C) on soil C and N pools of SOM fractions.

The C and N pools in particulate SOM showed a higher relative increase than in mineral-associated organic matter, constituting a more sensitive SOM fraction to the effect of soil management changes. Compared with the BS system, the C pool in particulate SOM increased 72 (2.1 Mg ha^{-1}) and 148% (4.3 Mg ha^{-1}) under O + V/M + C and M + C, respectively (Table 3). The corresponding increases in the N pool were 85 (78 kg ha^{-1}) and 154% (142 kg ha^{-1}) (Table 4). There were higher C and N pool increases in soil surface layers (0–25 and 25–75 mm) than in deeper layers. Particulate SOM has been associated with mineralizable N (Franzluebbers and Arshad, 1997; Chan, 1997). The increase of this SOM fraction may be important to improve sev-

Table 3. The C pool in particulate (POM) and mineral-associated organic matter and ratio POM-C/total soil organic C (SOC) in a sandy clay loam Acrisol under three no-till cropping systems.

Cropping systems	POM-C			Mineral-associated C			POM-C/SOC
	Mg ha^{-1}	Δ%	Δ Mg ha^{-1}	Mg ha^{-1}	Δ%	Δ Mg ha^{-1}	
				0–25 mm			
Bare soil	1.1b†			4.4b			0.21
O+V/M+C‡	2.2ab	100	1.1	6.9ab	57	2.5	0.24
M+Cajanus	3.8a	245	2.7	7.4a	68	3.0	0.34
CV, %§	23			19			
				25–75 mm			
Bare soil	0.8b			8.3b			0.09
O+V/M+C	1.4ab	75	0.6	11.2a	35	2.9	0.11
M+Cajanus	2.1a	162	1.3	11.1a	33	2.8	0.16
CV, %	38			13			
				75–175 mm			
Bare soil	1.0b			14.3b			0.06
O+V/M+C	1.5a	50	0.5	17.1ab	20	2.8	0.08
M+Cajanus	1.3ab	30	0.3	18.1a	27	3.8	0.07
CV, %	21			10			
				0–175 mm			
Bare soil	2.9b			27.0b			0.10
O+V/M+C	5.0ab	72	2.1	35.2ab	30	8.2	0.13
M+Cajanus	7.2a	148	4.3	36.5a	35	9.5	0.16
CV, %	23			12			

† Values in column followed by the same letter are not significantly different at $P \leq 0.05$ according to Tukey's test.
‡ Oat + vetch/maize + cowpea.
§ Coefficient of variance.

eral soil characteristics such as macroaggregate stability (Paladini and Mielniczuk, 1991), biomass C, and microbial activity (Cattelan and Vidor, 1990). Specific studies to evaluate the effects of particulate SOM on such soil characteristics have not been performed under the current experimental conditions.

These results were similar to those of other studies which determined that the more active SOM fractions, such as biomass C and N (McCarty et al., 1998), and light fractions (Bremer et al., 1995), have also been more sensitive to soil use and management system changes. Increases of 24% and 71% in total soil organic C and in

the C pool of particulate SOM, respectively, have been reported for a sandy-loam Typic Paleudalf from southern Brazil under fallow and *Stizolobium cinereum* + maize cropping system compared with BS (Amado et al., 2000).

The C and N pool ratios of particulate SOM/total SOM ranged from 0.10 to 0.16 and from 0.04 to 0.07 in the 0- to 175-mm layer, respectively (Table 3 and Table 4). These ratios were lower than those reported by Franzluebbers and Arshad (1997) and Chan (1977) in soils evaluated under cold subtropical semiarid climatic conditions. The lower C and N pools in particulate SOM were probably caused by more favorable environ-

Table 4. The N pool in particulate (POM) and mineral-associated organic matter and ratio POM-N/total nitrogen (TN) in a sandy clay loam Acrisol under three no-till cropping systems.

Cropping systems	POM-N			Mineral-associated N			POM-N/TN
	Kg ha^{-1}	Δ%	Δ Kg ha^{-1}	Kg ha^{-1}	Δ%	Δ Kg ha^{-1}	
				0–25 mm			
Bare soil	33b†			398b			0.08
O+V/M+C‡	74ab	124	41	647a	63	249	0.10
M+Cajanus	127a	285	94	700a	76	302	0.15
CV, %§	40			17			
				25–75 mm			
Bare soil	26b			704b			0.04
O+V/M+C	52a	100	26	997ab	42	293	0.05
M+Cajanus	67a	158	41	1039a	48	335	0.06
CV, %	43			14			
				75–175 mm			
Bare soil	33a			1312a			0.02
O+V/M+C	44a	33	11	1495a	14	183	0.03
M+Cajanus	40a	21	7	1385a	6	73	0.03
CV, %	32			8			
				0–175 mm			
Bare soil	92b			2414b			0.04
O+V/M+C	170ab	85	78	3139a	30	725	0.05
M+Cajanus	234a	154	142	3124ab	29	710	0.07
CV, %	36			9			

† Values in column followed by the same letter are not significantly different at $P \leq 0.05$ according to Tukey's test.
‡ Oat + vetch/Maize + cowpea.
§ Coefficient of variance.

mental conditions for biological activity, such as higher soil temperatures and moisture, which influence the decomposition and humification of SOM (Vanlauwe et al., 1999).

When the cropping system effects on C and N pools were analyzed in absolute values, a different behavior was observed. Quantitative increases of C and N pools in mineral-associated SOM were two to nine times higher than those for particulate SOM in the 0- to 175-mm layer (Table 3 and Table 4). The total C and N pools in mineral-associated SOM of BS were 27.0 Mg ha^{-1} and 2414 kg ha^{-1}, respectively. The C pool in mineral-associated SOM increased 8.2 and 9.5 Mg ha^{-1} under O + V/M + C and M + C, respectively. Corresponding N increases were 725 and 710 kg ha^{-1}. The presence of higher C and N pools and greater increase of mineral-associated SOM compared with particulate SOM were associated with greater recalcitrance of the first organic fraction to biological decomposition, as suggested from the relative SOM decay rates.

Relative SOM decay rates in the 20- to 53-, 2- to 20-, and <2-μm granulometric fractions declined exponentially with increases in Fe oxides and kaolinite (Fig. 1). These results suggest that mineral components were significant in SOM stability for the sandy clay loam soil. Considering that humic substances have similar saturation curves (Sposito et al., 1996), the differences in saturation curves observed in our experiment (Fig. 2) can be linked to the interaction of SOM with soil mineral components. This interaction was particularly intense in the 2- to 20- and <2-μm granulometric fractions, where high Fe oxides and kaolinite concentrations were detected (Fig. 1). The strong interaction with SOM resulted in a short relaxation time. Consequently, saturation of semiquinone ESR signal was never achieved in the smaller granulometric fractions. Conversely, for the 20- to 53-μm granulometric fraction, where Fe oxides and kaolinite concentrations were low (Fig. 1), the saturation curve indicated higher relaxation time and weak interaction with mineral components.

The increase in the C pool of mineral-associated SOM emphasized the importance of the physical stability of SOM on both C sequestration and preserving or increasing soil quality in tropical and subtropical environments. Organic-mineral interactions in soils with variable charge determine the potential of these soils to act as an atmospheric CO_2 sink. The C pool increase in particulate SOM under O + V/M + C and M + C represented a CO_2 net removal of 7.7 and 15.8 Mg ha^{-1} from the atmosphere, respectively. The C pool increase in mineral-associated SOM represented a CO_2 net removal of 30.1 and 34.9 Mg ha^{-1}, respectively, compared with the BS system. Therefore, respectively 80 and 69% of the total atmospheric CO_2 net removal under O + V/M + C and M + C, compared with BS, was because of an increase in the C pool of mineral-associated SOM. The increase of more humified fractions promoted by cover crops is probably largely responsible for improving physicochemical soil characteristics (Burle et al., 1997), such as cation-exchange capacity (CEC) at pH 7 (Table 2), reinforcing the importance of this SOM fraction in improving soil quality.
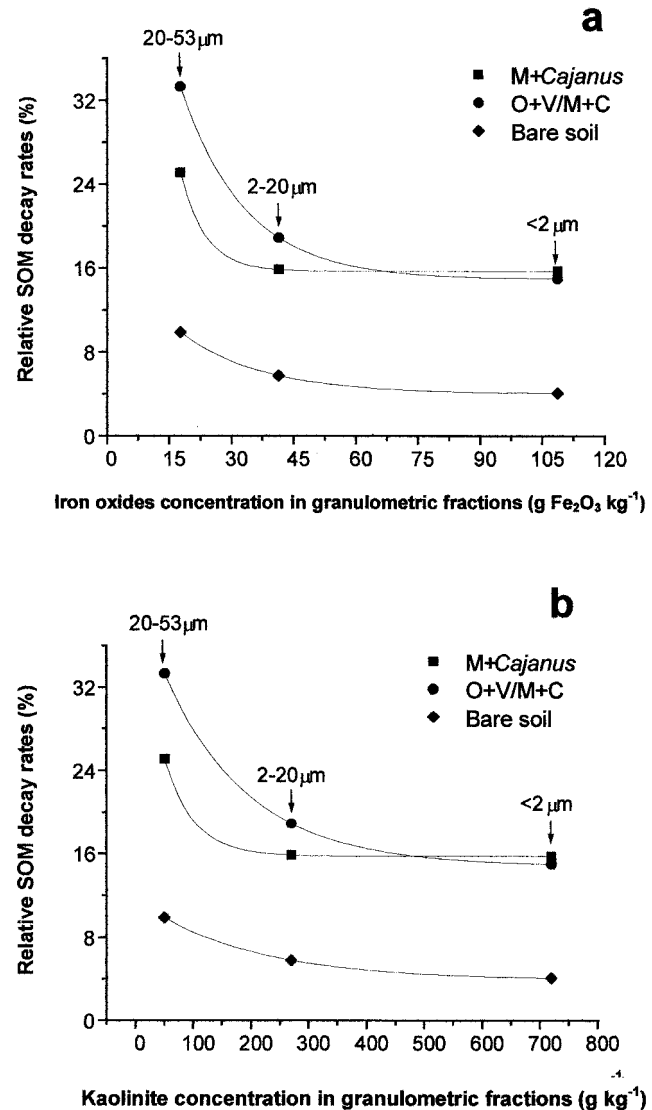

Fig. 1. Relationship between relative soil organic matter (SOM) decay rates and (a) Fe oxides and (b) kaolinite concentrations in the 20- to 53-, 2- to 20-, and <2-μm granulometric fractions of a sandy clay loam Acrisol. Mineral concentrations in soil granulometric fractions were analyzed in composite samples resulting from the mixture of three field replicates. Mean values showed <8% difference among cropping systems.

SUMMARY AND CONCLUSIONS

Particulate SOM represented a small fraction of total SOM in a sandy clay loam Acrisol under subtropical climatic conditions of southern Brazil. No-till cropping systems with high annual crop residue additions increased both particulate and mineral-associated SOM fractions in a 12-yr period. The C and N pools in particulate SOM varied by cropping system and were more sensitive to evaluate soil management quality than both total and mineral-associated SOM fractions. However, the C and N pools in mineral-associated SOM increased several times more than the particulate fraction caused by greater physical recalcitrance of mineral-associated SOM to biological decomposition. The relationship between relative decay rates of SOM and Fe oxides and kaolinite concentrations demonstrated the physical sta-
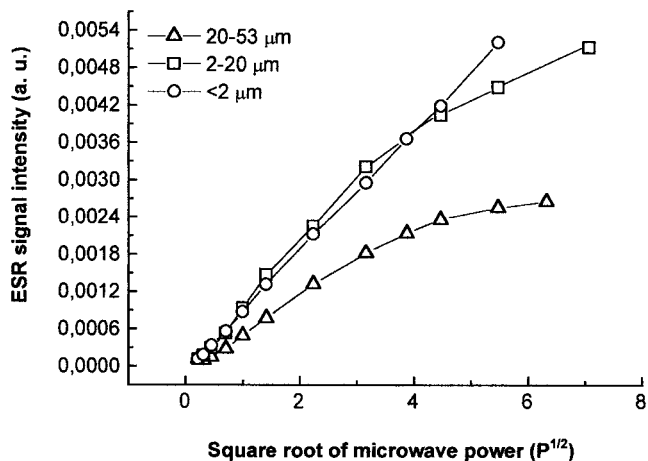
Fig. 2. Saturation curves of the semiquinone free radical signal by electron spin resonance (ESR) spectroscopy in the 20- to 53-, 2- to 20-, and <2-μm granulometric fractions of Maize + *Cajanus cajan* **cropping system. (a.u. = arbitrary units.)**

bility of SOM caused by interaction with variable charge minerals. Power saturation curves determined by ESR spectroscopy in 20- to 53-, 2- to 20-, and <2-μm granulometric fractions also reinforced this hypothesis. The physical stability of SOM in variable-charge soils is very important in maintaining or increasing soil quality and is determining the soil's potential to act as an atmospheric CO_2 sink in tropical and subtropical regions.

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