

Monitoring Soil Organic Carbon Erosion with Isotopic Tracers: Two Case Studies on Cultivated Tropical Catchments with Steep Slopes (Laos, Venezuela)

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20.1 INTRODUCTION

With an estimated content of 1600 Pg carbon (C) in the first m, soil organic matter (SOM) is the largest terrestrial reservoir of organic C, exceeding the terrestrial biosphere (560 Pg C) and atmosphere (750 Pg C) storage capacities (Post et al., 1982; Eswaran et al., 1993; Sundquist, 1993). Tropical regions account for nearly 30% of the soil organic C (SOC) reservoir (Milliman et al., 1987; Ross, 1993; Dixon et al., 1994; Lal, 1995; Zech et al., 1997). SOM plays a key role in soil physical and chemical properties such as structural stability, porosity, nutrient availability, and ion-exchange capacity (Oades, 1986; Le Bissonnais, 1990; Tiessen et al., 1994; Lal et al., 1999; Roscoe et al., 2001; Puget et al., 2001).

Rising atmospheric CO₂ levels coupled with global warming may stimulate terrestrial photosynthesis (fertilization effect) and enhance SOC sequestration (e.g., Houghton et al., 1993; Kirschbaum, 2000; Schlesinger and Andrews, 2000). However, present-day deforestation and land-use change question the efficiency of C storage in soils because enhanced runoff and tillage erosion may induce opposite trends. Erosion leads to physical removal of SOC through soil particle redistribution along slopes and enhances mineralization of SOM, increasing as much the C flux to the atmosphere. Since the behavior of soils as a sink or a source for atmospheric CO₂ is still debated (Ciais et al., 1995; Houghton et al., 1998), monitoring SOC erosion in cultivated tropical environments may provide significant information on the C flux to the atmosphere and contribute to a better management of soil resources for sustainable development purposes (e.g., Lal, 1990; Houghton, 1991; Houghton et al., 1993).

The transfer by erosion of SOC to the hydrographic network may be addressed using relationships between total organic C concentrations in top soil horizons and in suspended loads of river discharges (Meybeck, 1982; 1993; Probst, 1992; Ludwig et al., 1996) or hydrological and GIS-based models that link organic C outputs to climatic and physiographic settings on the catchments such as precipitation, slope, vegetation cover, and soil properties (Post et al., 1982; Esser and Kohlmaier, 1991; Probst, 1992; De Roo, 1993; Browne, 1995). However, these approaches generally provide little information on the nature and the source of organic matter exported from the catchments and its further evolution in the hydrographic network. Moreover, providing a direct link between SOC erosion and suspended organic matter loads is a difficult assignment. Soil erosion rates monitored with field plot experiments (Morgan, 1986; Lang, 1992; Loughran and Campbell, 1995) often do not match the sediment delivery ratio of stream flows, reflecting sorting, deposition, or remobilization processes that may take place along slopes and river banks on the watersheds (Meade, 1988; Dedkov and Mozzherin, 1992; Milliman and Syvitski, 1992; Trimble and Crosson, 2000). Accordingly, scaling up field plot measurements to catchments for SOC erosion budgets and models involves large and unexplained variability with respect to soil and sediment composition. Another major uncertainty on the impact of soil erosion on the C cycle is the mineralization rate of suspended organic matter in the hydrological network (range: 0 to 100%, see references in Lal, 1995). It is likely that the extent of organic carbon degradation is controlled by the nature of organic matter (i.e., labile vs. resistant or protected compounds, Hedges et al., 2001). The behavior in surface environments of refractory organic matter derived from geological basement sources (such as shales or other organic-matter bearing rocks) contrasts sharply with that of vegetation debris, charcoals, or humic substances generated by soil erosion. Discrimination between lithic and soil sources of organic matter in stream flows is important to constrain both soil erosion and global C cycle budgets (Meybeck, 1993; Kao and Liu, 1996; 2000; Raymond and Bauer, 2001a; 2001b; Megens et al., 2002). The mineralization rate of suspended organic matter is also controlled by redox conditions, in particular for water reservoirs located at the outlet of the watersheds, which may constitute ultimate receptacles for runoff and undergo severe oxygen depletion with high organic matter supply from the drainage areas (Likens, 1972; Chapra and Dobson, 1981; Stumm and Morgan, 1996). With lower mineralization rates than in soils and high residence times, riparian zones, wetlands, and lacustrine environments may possibly behave as C sinks with respect to

atmospheric CO₂ (e.g., Dean and Gorham, 1998) and bias organic C budgets based on direct link between soil erosion on catchments and riverine transport.

Isotope tracers such as radionuclides or stable isotopes may support identification and quantification of the different pools of organic matter generated by erosion on watersheds and the sources of suspended organic matter in stream flows and river discharges (Mariotti et al., 1980; 1984; 1991; Hedges et al., 1986; Wada et al., 1987; Cai et al., 1988; Ittekkot, 1988; Bird et al., 1992; 1994; Walling et al., 1993; Thornton and McManus, 1994; Onstad et al., 2000; Masiello and Druffel, 2001) or provide information to refine SOC budgets (Arrouays et al., 1995; Balesdent et al., 1988; Balesdent and Mariotti, 1996; Ritchie and McCarty, 2003). This chapter presents the results of several isotopic studies carried out on two cultivated watersheds with steep slopes located in tropical regions. The objectives were to (1) link the erosion status and the organic C content of cultivated soils using ¹³⁷Cs and total organic carbon measurements for the Houay Pano catchment (Laos), (2) better constrain the source of suspended organic matter during flood events by monitoring the composition of suspended loads in runoff and stream flows with stable N and C isotope measurements for the Rio Boconó watershed (Venezuela), and (3) evaluate the impact of a water reservoir located at the outlet of the Rio Boconó watershed (Venezuela) on suspended organic C exportation to the floodplain. The principles of the isotopic methods used in this study are summarized in the text along with each case study.

20.2 PHYSIOGRAPHIC SETTINGS AND MATERIAL SAMPLING

20.2.1 The Houay Pano Catchment in Northern Laos

Soil sampling for ¹³⁷Cs and organic C measurements was carried out for selected soils of the Houay Pano catchment (67 ha), located near Luang Prabang in northern Laos (19°51'00"–19°51'45"N, 102°09'50"–102°10'20"E). Soils were collected for 10 cm depth increments, using a corer with an internal 4.3 × 4.7 cm area, especially designed by the LSCE (laboratoire des Sciences du Climat et de l'Environnement). Catenas mainly cultivated with Job's tears (*Coix lacryma Jobi*) were sampled in October 2000 along two toposequences located on both sides of the main thalweg (Figure 20.1). Soils of the catchment are Entisols (18.5%; clay soils with medium

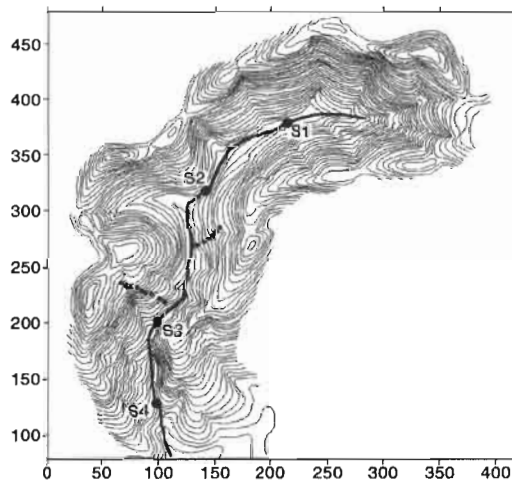


Figure 20.1 Location of soils sampled for ¹³⁷Cs and TOC measurements along two toposequences on the Houay Pano catchment (Laos). S1 to S4 refer to the main weirs. Grey circles correspond to soil locations.

fertility, pH = 6.4), Ultisols (33.1%; clay soils with medium fertility, pH = 5.5), and Alfisols (48.5%; heavy clay soils with medium fertility, pH = 6.2), settled on a geological basement, composed of Permian to Upper Carboniferous argillite series (shales, mudstones, and fine-grained sandstones) overlaid by limestone cliffs (NAFRI, 2001). After removal of organic matter, soils sampled for this study contain 43 to 50% of clays (< 2 μm), 34 to 42% of "fine silts" (2 to 50 μm) and 14 to 18% "coarse silts" and sands (50 to 2000 μm). The catchment is characterized by steep slopes. Slope gradients range 3 to 150% and average 60% for altitudes between 425 to 718 m above sea level. Mixed-deciduous and dry forests, mainly located on hilltops, account for less than 26% of land cover. Farmers mainly grow upland rice (*Oryza savita*), maize (*Zea mays*), and Job's tears, using slash-and-burn shifting cultivation with rotating fallows. Detailed surveys and interviews with farmers, cross checked by aerial photographs, showed that during the past four decades, catenas were generally cultivated for 7 to 8 years and maintained under fallow the remaining time. Mean annual temperature, precipitation, and evapo-transpiration (1986 to 1996 record) were 25.3°C, 1403 mm, and 1022 mm, respectively (NAFRI, 2001). Between 81% and 99% of total rainfall currently occurs during the rainy season, from April to October, with a maximum supply in July and August. Average precipitation records for Luang Prabang (LP), where precipitation data are available before 2000, can be used for Houay Pano (HP) where this study was conducted ($P_{LP} = 1.07 P_{HP}$, $r = 0.94$, $n = 39$ with $P =$ monthly precipitations in mm). Between 1954 and 1976, during the period of maximum ^{137}Cs fallout in Southeast Asia (see further in the text), precipitation averaged 1139 ± 225 mm in Luang Prabang, with an excursion to 1564 mm in 1963 (Bricquet et al., 2001). In 2000, the mainstream discharge of the Houay Pano catchment averaged 9.4 l s^{-1} with a total sediment yield of 7 t ha^{-1} (NAFRI, 2001).

20.2.2 The Rio Boconó Watershed in Northwest Venezuela

Erosion of SOC was monitored under natural rainfall conditions at various space and time scales by collecting runoff samples generated on experimental field plots and in stream flows during flood events. Located in northwestern Venezuela (Figure 20.2a, $08^{\circ}57' - 09^{\circ}31' \text{N}$, $70^{\circ}02' - 70^{\circ}34' \text{W}$), this watershed covers about 8% of the Venezuelan Andes (1620 km^2). Its outlet in the Llanos floodplain is closed by the Peña Larga dam built in 1983 (area, 122 km^2 , volume, 2850 10^6 m^3 , maximum depth, 72 m, Figure 20.2b). This large reservoir, designed for water and electricity production, also provided the opportunity to monitor offsite effects of SOC erosion on water quality (Bellanger et al., 2004b). Runoff experiments were performed on 30 m^2 experimental field plots (Felipe-Morales et al., 1977; Lal, 1990) on a bare field plot, formerly under maize and maintained manually free of vegetation, a maize field plot, and a coffee field plot (*Coffea arabica*). After removal of organic matter, soils are composed of 24 to 28% clays (< 2 μm), 18 to 26% silts (2 to 63 μm) and 46 to 56% sands (63 to 2000 μm). Coffee is currently cultivated on hill slopes at intermediate altitudes (500 to 1500 m), under 4 to 5 m high tree covers (Ataroff and Monasterio 1997), whereas maize fields are often established near the main streams. Runoff and suspended sediment discharge were monitored between April 15 and September 15 for each field plot with weekly measurements of water volume and sediment yield in the collection tanks. In addition, a single flood event (June 14, 1998) was covered in detail with a sampling frequency of 5 to 30 min. All suspended sediments were collected in polyethylene bottles in runoff and stream flows of the watershed during flood events and at several depths in the water column of the Peña Larga reservoir.

The tropical climate conditions prevailing in the region are highly influenced by altitude, ranging 200 to 4000 m above sea level within the watershed. The average temperature decreases with altitude by 0.6°C per 100 m. Precipitation records also display decreasing trends, of 1500 to 2500 mm in the lower parts of the watershed to less than 1000 mm for the higher altitude Paramó grasslands (2700 to 4000 m; Cornières, 1998). About 87% of total rainfall takes place during the rainy season, between April and October with a maximum in June and July (MARNR, 1991 in Rodríguez, 1999). Mean annual temperature, precipitation, and evapo-transpiration at the sampling

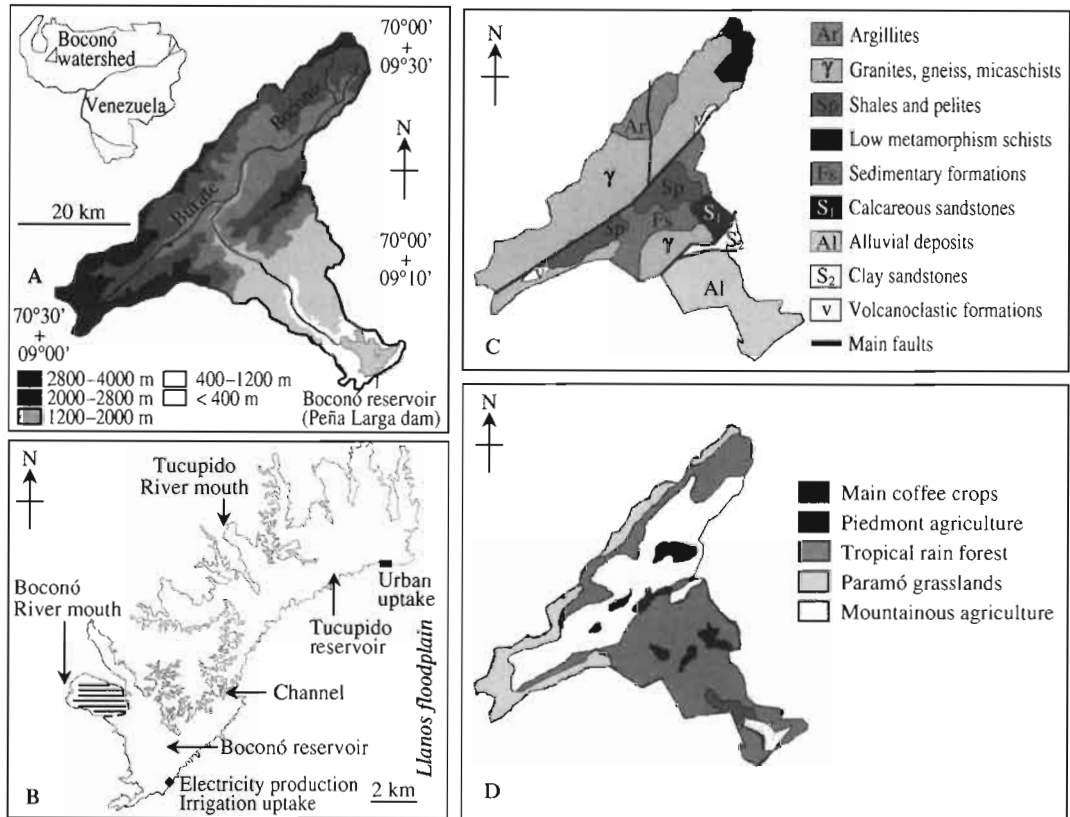


Figure 20.2 Location of the Rio Boconó watershed in Venezuela with main physiographic characteristics (a) altitudes, (b) physiographic map of the Peña Larga water dam at the outlet of the Rio Boconó watershed with open and black circles refer to sampling locations in September 1999. (Modified from Bellanger, B., S. Huon, P. Steinmann, F. Chabaux, F. Velasquez, V. Vallès, K. Arn, N. Clauer, and A. Mariotti. 2004a. *Applied Geochemistry*, 19, 1295–1314. With permission.) (c) Simplified lithological map, and (d) simplified vegetation cover. (Modified from Bellanger, B., S. Huon, F. Velasquez, V. Vallès, C. Girardin, and A. Mariotti. 2004b. *Catena*, 58, 125–150. With permission.)

area location (Corojó subcatchment, 1200 m above sea level) are 21°C , 1600 mm, and 980 mm, respectively (Rómulo-Quintero, 1999). Ultisols are the most common soil types found on the Boconó watershed (Pérez, 1997). They are derived from a geological basement composed of Paleozoic meta-sedimentary rocks and Quaternary alluvial deposits for the southern part of the watershed whereas, crystalline rock formations mainly outcrop on the northern part (Figure 20.2c). Approximately 42% of the watershed is occupied by rain forest (500 to 1700 m) and dry mountainous forest (1500 to 2700 m). Paramó-type vegetation and fallows make up 17.5% of the area. The remaining land is used for agriculture (i.e., maize, coffee, and living productions) often established on steep slopes (Figure 20.2d). The main stream of the watershed is the Boconó River whose discharge averages $77.7\text{ m}^3\text{ s}^{-1}$ (1952 to 1973 record; MARNR, 1991 in Rodríguez, 1999) with a mean sediment yield of $7\cdot 10^6\text{ t yr}^{-1}$ (Cornières, 1998).

20.3 ANALYTICAL METHODS

Total organic carbon (TOC), total nitrogen (TN), $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ isotope ratios were measured, after carbonate removal, on the same sample aliquot by EA-IRMS (Carlo-Erba NA-1500 NC Elemental Analyser on line with a Fisons Optima Isotope Ratio Mass Spectrometer; Girardin and

Mariotti, 1991). Carbonate removal for TOC analyses was performed using 1N HCl under pH control (above 4.0), in order to reduce possible leaching effects (Huon et al., 2002). Results for isotope abundance are reported in per mil (‰) relative to Pee Dee Belemnite (PDB) standard and relative to air N₂, for $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$, respectively (Coplen et al., 1983). TOC and TN concentrations are reported in mg g⁻¹ of dry sample (equivalent to weight %). During the course of this study, analytical precision was better than the means ($\pm 1\sigma$): $\pm 0.1\%$ for $\delta^{13}\text{C}$, $\pm 0.3\%$ for $\delta^{15}\text{N}$, 0.1 mg g⁻¹ for TOC and 0.05 mg g⁻¹ for TN. Data reproducibility was controlled by replicated analysis of samples (50%) and by a tyrosine standard that yielded $-23.2 \pm 0.1\%$ and $10.1 \pm 0.30\%$ for $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$, respectively (mean values for tyrosine: $\delta^{13}\text{C} = -23.2\%$, $\delta^{15}\text{N} = 10.0\%$).

All soil samples were ground using a hand mortar and sieved at 2 mm in order to remove coarse vegetation debris, stones, and other coarse lithic aggregates. Fine-size fractions of soils (i.e., < 50 μm or < 200 μm) were separated by wet sieving from grounded soil sample aliquots. For suspended sediments no grinding was necessary and fine fractions were directly recovered from sieving of samples collected in streamflows and runoff flows and stored in polyethylene bottles. In both cases the fractions are representative of the total fine fraction of soils or sediments. Dry bulk soil "densities" (in g cm⁻³) were calculated using the ratio of lyophilized soil weight to sampled volume. ¹³⁷Cs activities were measured from subsamples (60 to 80 g) that were put in tightly closed plastic boxes and submitted to 24 hours γ -counting. Coaxial HP Ge N-type detectors were used for γ -spectrometry (8000 channels, low background). Efficiencies and backgrounds were periodically controlled with sediment and soil standards (Soil-6, IAEA-135, IAEA-375, and KCl).

20.4 RESULTS AND DISCUSSION

20.4.1 Monitoring SOC Erosion with ¹³⁷Cs for Cultivated Soils of the Houay Pano Catchment in Laos

20.4.1.1 Application of ¹³⁷Cs Measurements in Soil Erosion Studies

The erosion status of soils can be assessed using fallout ¹³⁷Cs labeling techniques. ¹³⁷Cs is an artificial radionuclide (half-life of 30.17 years) produced by atmospheric testing of thermonuclear weapons during the 1950s and 1960s in the Northern Hemisphere (UNSCEAR, 1969; 1993; 2000) and after the Chernobyl reactor accident in 1986 (Cambray et al., 1989; SPARTACUS, 2000). During these events, large amounts of ¹³⁷Cs were released and dispersed in the stratosphere. Most of the fallout occurred between 1954 and 1976 with a maximum supply in 1963. Once ¹³⁷Cs reaches land surface, it is strongly adsorbed by fine-grained soil particles and no longer exchangeable (Rogowski and Tamura, 1965; Ritchie and McHenry, 1973; Cremers et al., 1988). Therefore, the redistribution of ¹³⁷Cs is likely to be controlled only by erosion, transport, and sedimentation of soil particles with limited migration with soil depth, usually not exceeding around 30 cm, due to local bioturbation and convective-diffusive processes (Ritchie et al., 1974; Brown et al., 1981a; 1981b; Jong et al., 1983; Ritchie and McHenry, 1990; Walling and Quine, 1992; Loughran et al., 1987; Higgitt and Walling, 1993; He and Owens, 1995).

In cultivated soils, ¹³⁷Cs is mixed to the plow depth and surface concentrations are lower than for noncultivated soils (Quine et al., 1999). Undisturbed sites should have ¹³⁷Cs inventories that reflect the amount of ¹³⁷Cs fallout minus the loss due to radioactive decay. They provide reference values used as baselines for assessing the local erosion and deposition status of soils (Sutherland, 1994). Estimates of soil redistribution rates on cultivated and uncultivated soils may be derived from simple proportional (de Jong et al., 1983) to more complex mass balance models that link the amount of remaining soil ¹³⁷Cs activity with the assumed reference inventory, time, tillage dilution, particle size, and time-variant fallout ¹³⁷Cs input (Katchanovski and de Jong, 1984; Zhang et al., 1990; Quine, 1995; Walling and He, 1999; 2001). For each site, the calculated erosion rates

account for both tillage erosion (downslope redistribution of soil particles due to tillage, mainly weeding operations, e.g., for tropical soils, Turkelboom et al., 1999; Dupin et al., 2002) and erosion induced by water runoff (overland flow and rill erosion, e.g., Lal, 1990).

20.4.1.2 Estimates of Soil Erosion Rates on the Houay Pano Catchment

Soils of the Houay Pano catchment are theoretically suitable for ^{137}Cs investigations because most of the fields were under forests or long-term fallows when ^{137}Cs fallout took place in the Southeast Asian regions. Cultivation only started at the beginning of the 1970s with a maximum intensification during the 1990s. One of the major uncertainties for the assessment of soil erosion rate is the reference ^{137}Cs inventory used to discriminate undisturbed from eroded soils. Three different methods were used to constrain this reference value: (1) the average ^{137}Cs inventory of “assumed” undisturbed soils, located on long-term fallow fields on the catchment; (2) the “global” ^{137}Cs fallout value provided by the Csmode11 developed by Walling and He (2001) and; (3) the ^{137}Cs deposition inventory derived from average precipitation records in Vietnam (Hien et al., 2002). The ^{137}Cs inventory of three assumed undisturbed soils provided an average value ($\pm 1\sigma$) of $613 \pm 48 \text{ Bq m}^{-2}$ (range: 585 to 671 Bq m^{-2}). Based on the average annual precipitation of 1564 mm for the Houay Pano catchment in 1963 (Bricquet et al., 2001), the year of maximum fallout, the ^{137}Cs deposition inventory should be approximately 596 Bq m^{-2} (Hien et al., 2002) but decreases to $474 \pm 88 \text{ Bq m}^{-2}$ if the overall 1954 to 1976 precipitation record is taken into account (average $\pm 1\sigma$: $1168 \pm 263 \text{ mm}$, Bricquet et al., 2001). The ^{137}Cs reference inventories calculated with Csmode11 (Walling and He, 2001) are slightly higher, 635 Bq m^{-2} and 559 Bq m^{-2} for 1963 and 1954 to 1976 precipitation records, respectively. Both methods assume 100% homogenous deposition of ^{137}Cs fallout. However, the extent of radionuclide deposition bears some uncertainties with respect to local rainfall intensity (i.e., for ^7Be and ^{210}Pb ; Caillet et al., 2001). Since the average ^{137}Cs value measured for undisturbed soils and the rainfall-derived ^{137}Cs reference inventories for 1963 are consistent (596 Bq m^{-2} , 613 Bq m^{-2} , and 635 Bq m^{-2}), the average of the three values (615 Bq m^{-2}) was assumed to provide a first order estimate of the ^{137}Cs fallout reference for the Houay Pano catchment (with respect to year 1963). The ^{137}Cs inventories determined with the 1954 to 1976 precipitation records (474 and 559 Bq m^{-2} ; average, 517 Bq m^{-2}) are lower than the levels derived from undisturbed soils (range: 585 to 671 Bq m^{-2}). However, this value cannot be rejected because it accounts for the overall period of ^{137}Cs fallout in Southeast Asia. The results obtained with the low reference estimate (517 Bq m^{-2}) will also be discussed as an alternative solution for soil redistribution rate estimates.

Plots of ^{137}Cs vs. TOC inventories for the first 30 cm of 15 Ultisol–Alfisol soils sampled along two toposequences indicate that the distribution of ^{137}Cs inventories is significantly linked to the total SOC content. Equivalent patterns are displayed for each soil horizon (Figure 20.3a–c). A common process apparently relates the soil erosion status with the amount of SOM in the top 30 cm of soils (Mabit and Bernard, 1998; Ritchie and McCarty, 2003; Figure 20.3c–d). The K content of these 15 soils is rather low (range, 0.15 to 0.37%; average $\pm 1\sigma$: $0.24 \pm 0.06\%$; Table 20.1) with respect to other soils of the watershed (range, 0.15 to 1.43%, $n = 49$; average $\pm 1\sigma$: $0.60 \pm 0.36\%$, data not shown) and, thus, provides constrain for “homogenous” K-bearing clay content, a first-order proxy for fine-size particles content in soils. Therefore, it is likely that the distribution of ^{137}Cs in the selected soils is apparently not linked to differences in soil retention properties. No relationship between ^{137}Cs inventory and local slope gradient could be put forward, which indicates that the redistribution of soil particles is not directly linked to slope. Surprisingly, TOC (Figure 20.4a) and ^{137}Cs (Figure 20.4b) inventories decrease with the amount of fine-size ($< 50 \mu\text{m}$) particles in the soils. Given that (1) ^{137}Cs is preferentially bound to fine-size soil particles (i.e., clay particles) and (2) TOC concentration and ^{137}Cs activity both decrease with soil depth (Table 20.1), these trends are best explained by selective removal of topsoil layers by runoff and tillage erosion. Local outcrop of deeper soil layers with lower SOC and ^{137}Cs levels but higher fine-size particle content explain

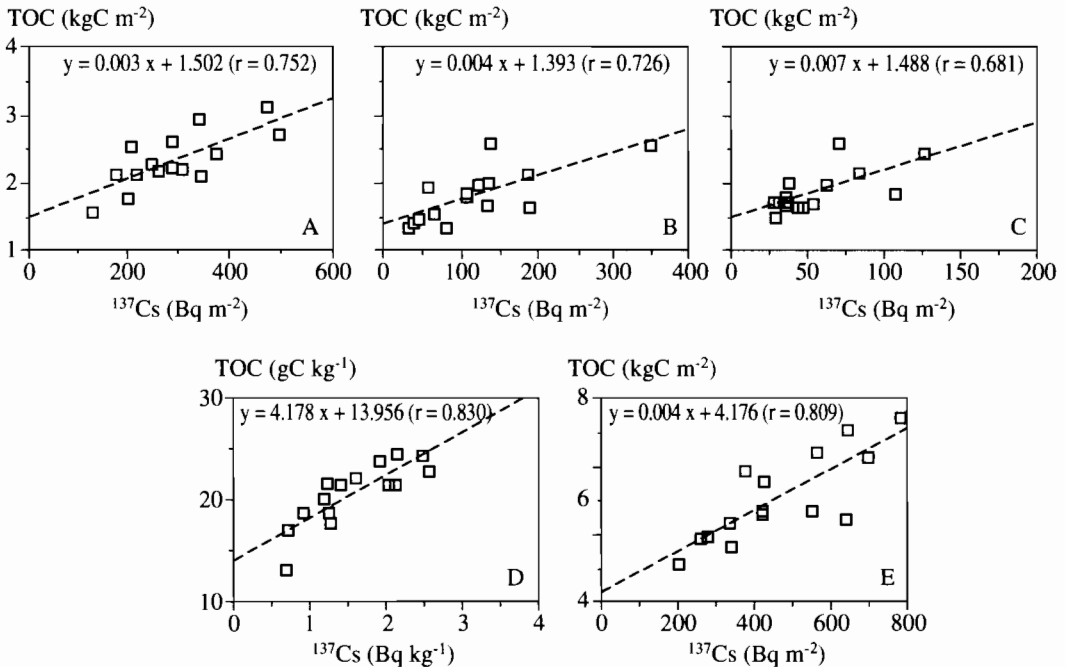


Figure 20.3 Plots of TOC (< 200 μm) vs. ^{137}Cs inventories of 15 cultivated soils of the Houay Pano catchment (Laos) for several soil depths (a) 0 to 10 cm, (b) 10 to 20 cm, (c) 20 to 30 cm, (d) and (e) integrated 0 to 30 cm. Regression coefficients are displayed between brackets for each plot.

Table 20.1 Average Data for the Top 30 cm of 15 Soils Sampled on the Houay Pano Catchment

Soil depth (cm)	< 50 μm (wt.%)	K (wt.%)	Soil density (g cm^{-3})	^{137}Cs (Bq kg^{-1})	TOC (mgC g^{-1})	^{137}Cs (Bq m^{-2})	TOC (kgC m^{-2})
0–10	78.5 ± 7.2	0.26 ± 0.06	0.84 ± 0.09	3.5 ± 1.3	28.4 ± 3.9	284 ± 105	2.3 ± 0.4
10–20	84.2 ± 3.2	0.24 ± 0.06	0.90 ± 0.13	1.4 ± 0.9	20.7 ± 3.1	121 ± 81	1.8 ± 0.4
20–30	83.1 ± 5.4	0.22 ± 0.06	1.15 ± 0.09	0.5 ± 0.3	16.4 ± 2.6	55 ± 30	1.9 ± 0.3

Note: $\pm 1\sigma$, standard deviation.

the observed pattern. Traditional farming practices do not involve plowing but two to four hoeing and weeding operations to a maximum depth of 2.5 cm (range, 2 to 3 cm) for each year of cultivation. The resulting tillage erosion, i.e., superficial redistribution of soil particles with down slope soil movements, may account for the large variability of ^{137}Cs inventory observed along the slopes and explain local outcrop of deeper soil horizons (Turkelboom et al., 1999; Dupin et al., 2002).

The ^{137}Cs inventories indicate that most of the soils display lower ^{137}Cs specific activities than the assumed reference (615 Bq m^{-2}). Based on the first-order relationship that links TOC and ^{137}Cs inventories, a reference TOC inventory of 6.64 kgC m^{-2} for 30 cm of soil (maximum penetration depth of ^{137}Cs , Figure 20.3e) was derived. Using the average soil density of the 15 soils ($\pm 1\sigma$: $0.98 \pm 0.08 \text{ g cm}^{-3}$; range: 0.84 to 1.15 g cm^{-3} , Table 20.1), the TOC inventory corresponds to an average TOC concentration of 21.93 mgC g^{-1} . This value is in the range of expected TOC concentrations for the undisturbed soils that decrease with depth from ca. 20 to 30 mgC g^{-1} (0 to 10 cm) to ca. 16 to 20 mgC g^{-1} (20 to 30 cm). Assuming a TOC concentration of 21.93 mgC g^{-1} in 1963 when these soils were still under fallow, the redistribution of soil particles along slopes provided, in 2000, average ($\pm 1\sigma$) soil and SOC net losses and gains of $-1.28 \pm 0.66 \text{ kg m}^{-2} \text{ yr}^{-1}$ with $-28.1 \pm 14.5 \text{ gC m}^{-2} \text{ yr}^{-1}$ and $0.66 \pm 0.46 \text{ kg m}^{-2} \text{ yr}^{-1}$ with $14.4 \pm 10.1 \text{ gC m}^{-2} \text{ yr}^{-1}$, respectively (Table 20.2). Although these averages bear high uncertainties (expressed as high standard deviations), soil

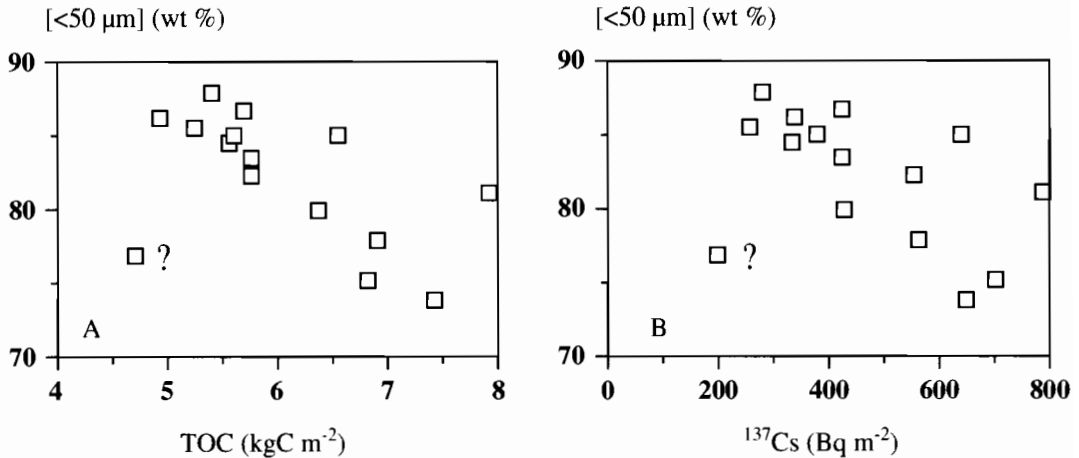


Figure 20.4 Plots of: (a) TOC and (b) ^{137}Cs inventories vs. fine-size ($< 50 \mu\text{m}$) fraction content of 15 cultivated soils (0 to 30 cm) of the Houay Pano catchment (Laos).

Table 20.2 Soil and TOC Redistribution Rate Estimates for 15 Soils Sampled on the Houay Pano Catchment

Soil Sample Number	TOC Inventory (kgC m^{-2})	TOC Change since 1963 ⁽¹⁾ ($\text{gC m}^{-2} \text{yr}^{-1}$)	Predicted Soil Redistribution ⁽²⁾ ($\text{kg m}^{-2} \text{yr}^{-1}$)	Soil Redistribution Rate Using ^{137}Cs ⁽³⁾ ($\text{kg m}^{-2} \text{yr}^{-1}$)	Predicted TOC Redistribution ⁽⁴⁾ ($\text{gC m}^{-2} \text{yr}^{-1}$)	Δ TOC ⁽⁵⁾ ($\text{gC m}^{-2} \text{yr}^{-1}$)
HP33	5.26	-37.29	-1.70	-0.64	-13.96	-23.33
HP34	6.55	-2.47	-0.11	-0.40	-8.72	6.24
HP35	5.72	-24.92	-1.14	-0.31	-6.75	-18.17
HP36	6.36	-7.63	-0.35	-0.30	-6.51	-1.11
HP37	5.07	-42.49	-1.94	-0.48	-10.53	-31.96
HP38	5.76	-23.83	-1.09	-0.31	-6.69	-17.14
HP39	5.55	-29.47	-1.34	-0.49	-10.79	-18.67
HP40	6.91	7.39	0.34	-0.07	-1.60	8.98
HP41	7.61	26.27	1.20	0.29	6.34	19.93
HP42	7.35	19.26	0.88	0.05	1.16	18.10
HP43	6.81	4.66	0.21	0.15	3.17	1.49
HP44	4.71	-52.18	-2.38	-0.91	-19.88	-32.30
HP45	5.61	-27.89	-1.27	0.04	0.92	-28.80
HP46	5.77	-23.47	-1.07	-0.09	-1.92	-21.55
HP47	5.24	-37.93	-1.73	-0.70	-15.34	-22.60
Mean	6.02	-16.30	-0.77	-0.28	-6.07	10.73
$\pm 1\sigma$	0.87	23.47	1.07	0.34	7.40	17.94

Note: Negative and positive signs refer to soil or TOC losses and gains, respectively. Due to low uncertainties on ^{137}Cs activity and TOC concentration measurements, all redistribution rates are displayed with less than 5% error.

⁽¹⁾Ratio of the difference between reference (6.64 kgC m^{-2}) and measured TOC inventories to time in years since 1963 (1963 to 2000 = 37 years).

⁽²⁾Ratio of TOC change to reference TOC concentration (21.93 mgC g^{-1}).

⁽³⁾Estimated using mass balance model 2 of *csmodel1* (Walling and He, 2001) with a ^{137}Cs reference inventory of 615 Bq m^{-2} , an average plough depth of $24.7 \text{ kg m}^{-2} \text{ cm}$ for 30 cm of soil and using year 1970 for the start of cultivation.

⁽⁴⁾Ratio of soil redistribution rate to reference TOC concentration (21.93 mgC g^{-1}).

⁽⁵⁾Difference between TOC change (column 3) and predicted TOC redistribution (column 6).

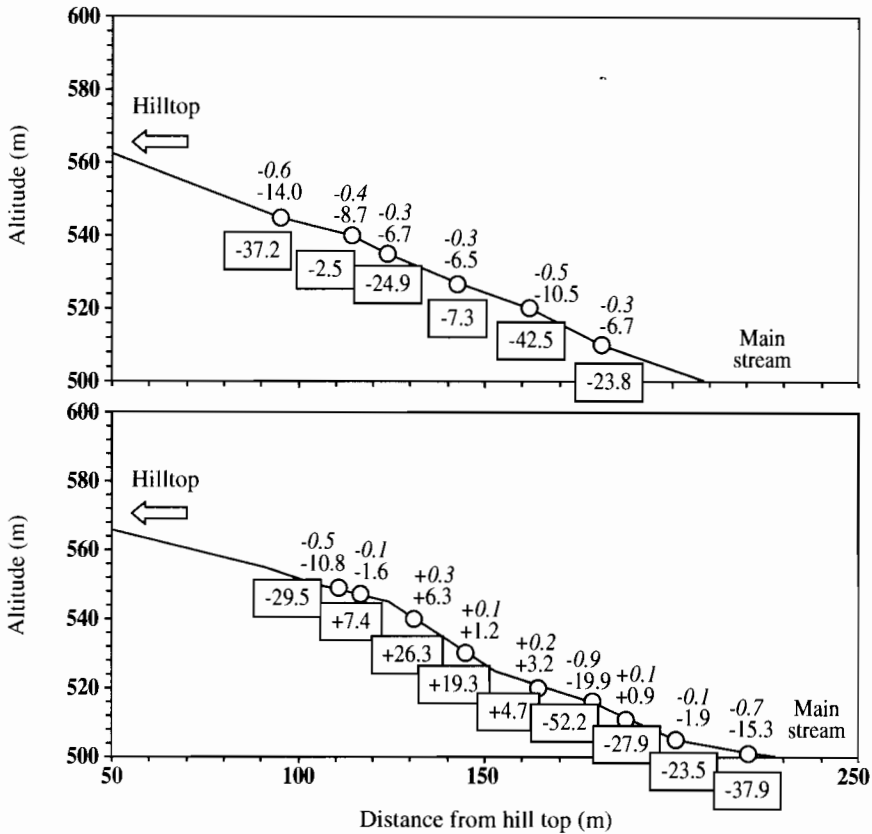


Figure 20.5 Distribution of soils along the two toposequences of the Houay Pano catchment (Laos). TOC changes (in $\text{gC m}^{-2} \text{yr}^{-1}$) are included in rectangles. Soil redistribution rates (in $\text{kg m}^{-2} \text{yr}^{-1}$) in italic and predicted TOC redistribution (in $\text{gC m}^{-2} \text{yr}^{-1}$) are reported above the slope line.

gains and losses are not randomly distributed along slopes. For one of the two toposequences (Figure 20.5b), soil and TOC depletions are found on hilltops and in bottom parts of the slopes whereas accumulations are observed for high to midslope soils. This unexpected pattern may be explained by the chronology of soil occupation along the slopes of the catchment. In 1963, all fields were under fallow and cultivation started in the early 1970s for catenas located near the main stream, with progressive extension to hilltops during the 1980s and 1990s. Therefore, the duration of cultivation apparently controls the extent of soil particles redistribution calculated with the TOC inventory. Direct monitoring of soil erosion on selected subcatchments of the Houay Pano watershed provided exportation budgets of -0.573 kg m^{-2} and -0.058 kg m^{-2} for soils under traditional cultivation and fallow grounds, respectively (NAFRI, 2001). Making the simple assumption that soil erosion rates can be estimated on the basis of 7 years of cultivation with 30 years of fallow during the past 37 years (time elapsed between 2000 and 1963), an overall soil erosion rate estimate of $-0.16 \text{ kg m}^{-2} \text{yr}^{-1}$ was calculated. This value is in the same order of magnitude as the soil loss rate estimates made with mass balance model 2 (range: $-0.07 \text{ kg m}^{-2} \text{yr}^{-1}$ to $-0.91 \text{ kg m}^{-2} \text{yr}^{-1}$; Table 10.2; Csmoell, Walling, and He, 2001).

However, modeled redistribution rates are much lower than those predicted with soil TOC depletion. With mass balance model 2, removal and accumulation rates averaged ($\pm 1\sigma$) $-0.43 \pm 0.25 \text{ kg m}^{-2} \text{yr}^{-1}$ and $0.13 \pm 0.11 \text{ kg m}^{-2} \text{yr}^{-1}$, respectively (Table 20.2), vs. $-1.28 \pm 0.66 \text{ kg m}^{-2} \text{yr}^{-1}$ and $0.66 \pm 0.46 \text{ kg m}^{-2} \text{yr}^{-1}$, respectively, using TOC-¹³⁷Cs inventories. Lower redistribution rates indicate additional losses of SOC with respect to ¹³⁷Cs-bound particle redistribution. This deviation most likely accounts for other processes controlling the SOC content such as

mineralization of SOM (i.e., direct conversion to CO_2 through soil respiration) and soil leaching (i.e., dissolved organic carbon [DOC] exportation by runoff). But this interpretation is more puzzling for accumulation sites, for which TOC contents rise with respect to the assumed initial inventory (6.64 kgC m^{-2}) and to hypothesized CO_2 and DOC releases. The soil redistribution rates calculated with the low ^{137}Cs reference inventory previously outlined (517 Bq m^{-2}) averaged ($\pm 1\sigma$): $-0.36 \pm 0.21 \text{ kg m}^{-2} \text{ yr}^{-1}$ and $+0.25 \pm 0.16 \text{ kg m}^{-2} \text{ yr}^{-1}$ for erosion and accumulation, respectively. The soil redistribution budget that can be derived from these values (using 2.5 cm depth as plowing depth, Walling and He, 2001) indicate that local soil losses may be compensated by local soil accumulations along slopes and is not corroborated by direct soil erosion estimates at the scale of the catchment ($-0.16 \text{ kg m}^{-2} \text{ yr}^{-1}$).

The results obtained for the Houay Pano catchment suggest that the redistribution pattern of ^{137}Cs and the SOC content in the first 30 cm of cultivated soils are significantly linked (i.e., Mabit and Bernard, 1998; Ritchie and McCarty, 2003). However, SOC erosion rates estimated using TOC depletion since 1963 are higher than those calculated with mass balance models based on ^{137}Cs inventories. This deviation is best explained by decoupling processes. Soil particles redistribution along slopes is not the only pathway for SOC depletion that is also controlled by mineralization of SOM through respiration and DOC releases by runoff. Although TOC losses attributed to soil erosion are highly variable from site to site (ca. 8 to 85% of total loss, average $\pm 1\sigma$: $36 \pm 21\%$), coupling ^{137}Cs measurements with SOC concentration measurements reduces uncertainty on soil redistribution rates derived from fallout ^{137}Cs models. However, major uncertainties on the assessment of SOC erosion (and accumulation) remain: (1) the reference ^{137}Cs value used to discriminate erosion and deposition sites by comparison with undisturbed soils and (2) the depth of the cultivation layer required to compute soil erosion rates (Csmode11 of Walling and He, 2001), which was assumed to be low (ca. 2.5 cm) according to the traditional farming methods used on the Houay Pano catchment. Better constrain on these two parameters may allow quantitative determination of SOC budgets that account for both physical and biogeochemical processes, i.e., soil stability, SOC mineralization, and SOC storage capacities.

20.4.2 Monitoring Erosion of Organic Carbon Derived from Soil and Rock Sources in Runoff and Stream Flows with Stable C and N Isotopes for the Boconó Watershed in Venezuela

20.4.2.1 Principles of Natural $^{13}\text{C}/^{12}\text{C}$ and $^{15}\text{N}/^{14}\text{N}$ Labeling of Soils and Suspended Sediments

The $\delta^{13}\text{C}$ composition of SOM is directly inherited (within 1‰) from plant cover $\delta^{13}\text{C}$ in surface horizons (Balesdent and Mariotti, 1996). The ^{13}C natural abundance in plants is mainly controlled by the $\delta^{13}\text{C}$ of the inorganic C source (subject to variations in time and space), the photosynthetic pathway utilized (C_3 , C_4 , CAM) and, to a lesser extent, by environmental conditions (temperature, humidity, pCO_2). Aerial plants with a C_3 photosynthetic cycle (85% of plant species) are strongly ^{13}C -depleted with a mean $\delta^{13}\text{C}$ of -26% due to large isotopic fractionation between CO_2 ($\delta^{13}\text{C} = -7.8\%$, present-day value) and plant organic carbon (Deines, 1980; O'Leary, 1988; Farquhar et al., 1989). In contrast, plants with C_4 photosynthesis discriminate less against $^{13}\text{CO}_2$ and display a mean $\delta^{13}\text{C}$ of -12% (Farquhar, 1983; O'Leary, 1988). The $\delta^{13}\text{C}$ of SOM in tropical regions is primarily influenced by the relative contribution of C_3 vs. C_4 plants, but is also affected by isotopic fractionation that occurs during the decomposition of plant tissues. This latter effect results from the metabolic activity of decomposer organisms or from differential decomposition of biochemical fractions of the plants that are isotopically distinct (Galimov, 1985; Fogel and Cifuentes, 1993). Accordingly, the composition of SOM in undisturbed soils generally displays a ^{13}C -enrichment pattern with depth (up to 1 to 3‰) associated with SOC mineralization (Melillo et al., 1989). Similar trends with soil depth are also displayed for $^{15}\text{N}/^{14}\text{N}$ ratios (Mariotti et al., 1980; Tiessen et al., 1984; Yoneyama,

1996). However, changes in the concentration and the $\delta^{13}\text{C}$ of atmospheric CO_2 due to fossil fuel combustion may also partly explain the ^{13}C -enrichment of stabilized SOM (e.g., Marino and McElroy, 1991). In the past 20 years, $\delta^{13}\text{C}$ values have been successfully used as tracers for SOM dynamics studies in particular with respect to vegetation changes (i.e., C_3 - C_4 plants or vice versa; Cerri et al., 1985; Balesdent et al., 1987; Cerling et al., 1989; Martin et al., 1990; Mariotti and Peterschmitt, 1994; Desjardins et al., 1994; Arrouays et al., 1995; Schwartz et al., 1996; Balesdent, 1996; Balesdent and Mariotti, 1996). However, caution must be used when applying decompositional models to simulate $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ changes because the chemical and mineralogical variability of soils also possibly controls the rate of SOM decomposition and the extent of isotopic fractionation with soil depth (e.g., Ehleringer et al., 2000; Buchmann and Kaplan, 2001; Krull et al., 2003).

Numerous studies have also used the potential of ^{13}C and ^{15}N natural abundance measurements in suspended sediments to derive information on the origin of particulate organic carbon (POC) in rivers (e.g., Hedges et al., 1986; Mook and Tan, 1991; Mariotti et al., 1991; Bird et al., 1992; 1994; Martinelli et al., 1999). Suspended sediment $\delta^{13}\text{C}$ are generally used to estimate the relative contribution of land-derived (allochthonous) vs. phytoplanktonic-derived (autochthonous) organic C sources. However, degradation of organic matter or interactions with dissolved phases during transport (e.g., Richey et al., 1980; 1988) may alter the $^{13}\text{C}/^{12}\text{C}$ content of suspended material and complicate the interpretation based on an ideal two components mixing. Since organic matter in most plants fractionate against air $^{15}\text{N}/^{14}\text{N}$, fresh or "fossil" vegetation debris transported in suspended sediments may be discriminated from SOM on the basis of their $\delta^{15}\text{N}$ values (e.g., Mariotti et al., 1984; Kao and Liu, 1996; Kao and Liu, 2000; Huon et al., 2002). Due to limited $^{15}\text{N}/^{14}\text{N}$ fractionation during burial of rapidly accumulating organic matter in sedimentary basins, refractory organic C derived from geological sources should have $\delta^{15}\text{N}$ (and $\delta^{13}\text{C}$) inherited from plant compositions, contrasting with that of mineralized SOM more enriched in ^{15}N and ^{13}C (Precambrian-Jurassic sedimentary records, in Tyson, 1995; Hayes et al., 1999). Additional constraints on the origin of riverine particulate organic carbon (POC) have been obtained by coupling $\delta^{13}\text{C}$ values with C/N ratio, $\Delta^{14}\text{C}$, POC/Chlorophyll-a ratio or lipid characterization measurements (Kennicut et al., 1987; Hedges et al., 1994; Thornton and Mc Manus, 1994; Barth et al., 1998; Onstad et al., 2000; Masiello and Druffel, 2001; Meyers, 1997; Raymond and Bauer, 2001; Krusche et al., 2002).

20.4.2.2 Monitoring Erosion of Fine-Size (< 50 μm) SOC with Field Plot Experiments

During the rainy season (April to September 1998), mean soil loss and mean runoff can only be correlated for the bare field plot ($r = 0.83$, Figure 20.6). Cumulative losses of organic C in the fine-size (< 50 μm) fraction amounted 29.5 gC m^{-2} and 2.3 gC m^{-2} for the bare field plot and the coffee field plot, respectively (Figure 20.7, corresponding to soil losses of 1.8 kg m^{-2} and 0.05 kg m^{-2} , respectively). During the same period, cumulative runoff was 60 times higher for the bare field plot than for the cultivated field plots. Maximal runoff and in SOC loss occurred in June and July. Differences in the magnitude of soil loss and runoff between cultivated and bare field plots reflect the classical protective effect of the vegetation cover that reduces raindrop impact and runoff generation (Roose, 1977; Wischmeier and Smith, 1978). The vegetation cover apparently also controlled the grain size distribution of suspended sediments (Wan and El Swaify, 1997; Hairshine et al., 1999) and induced a sorting effect for cultivated soils. During the single rainfall event experiment, soils from the coffee field plot released almost exclusively fine-size sediments (mean weighted < 50 μm size fraction contribution: 93%, Figure 20.8) originating from the breakdown of surface soil aggregates (sheet erosion, Lal, 1990) and transported by limited runoff. In contrast, suspended sediments exported from the bare field plot by higher runoff had a grain size distribution that matches that of topsoil horizons (mean weighted < 50 μm size fraction contribution: 45% for suspended sediments vs. 47% for 0 to 20 cm soil horizons (Figure 20.8) and accounted for additional rill erosion processes (Lal, 1990). During rainfall, the overall fine (< 50 μm) SOC exportation amounted to 1.3 gC m^{-2} and 0.03 gC m^{-2} for the bare and the coffee field plots, respectively.

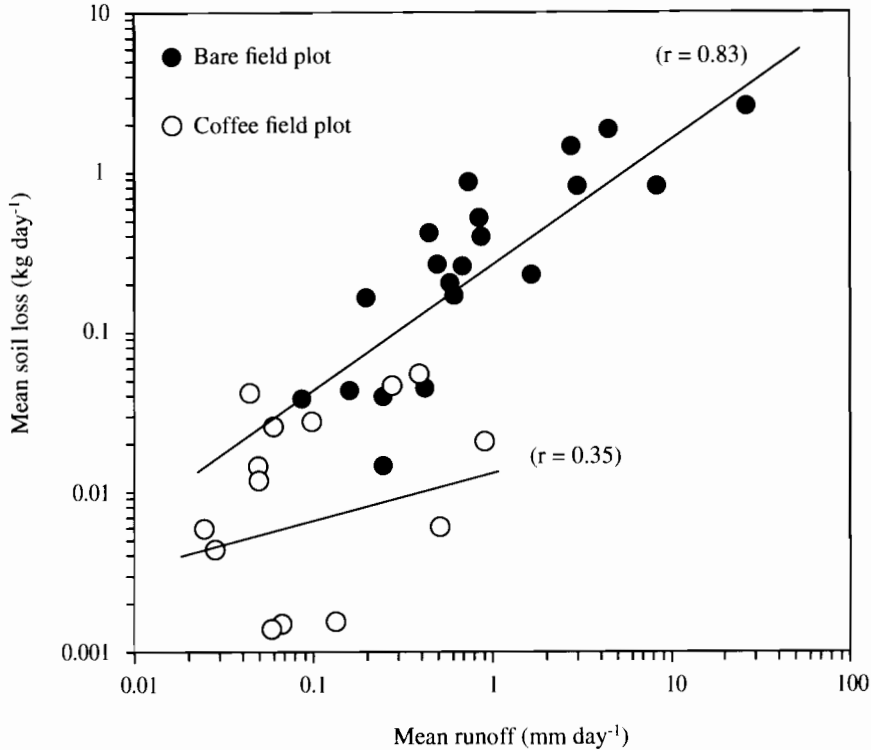


Figure 20.6 Plot of mean soil loss vs. mean runoff for the bare and the coffee field plots on the Rio Boconó watershed (Venezuela) during the rainy season. (From Bellanger, B., S. Huon, F. Velasquez, V. Vallès, C. Girardin, and A. Mariotti. 2004a. Experimental $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ study of soil organic carbon loss by erosion in the Venezuelan Andes. *Catena*, 58, 125–150. With permission.)

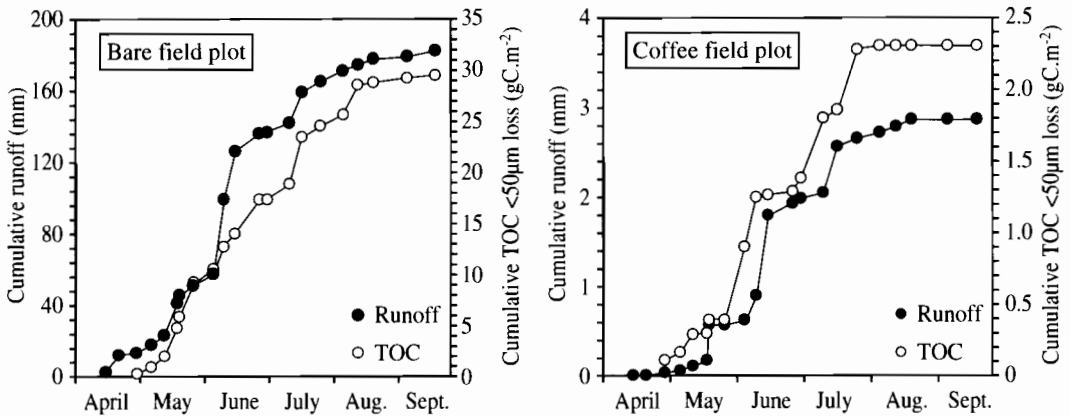


Figure 20.7 Plots of cumulative runoff and fine size TOC loss for the two experimental field plots of the Rio Boconó watershed (Venezuela) during the rainy season.

The composition ($\delta^{13}\text{C}$, $\delta^{15}\text{N}$, TOC, TOC/TN) of fine-size ($< 50 \mu\text{m}$) suspended organic matter is displayed as a function of suspended sediment concentration in Figure 20.9 for the bare field plot. For low sediment concentrations, the composition of fine organic matter reflected the transport of vegetation debris mobilized from top soil horizons and characterized by high organic C contents, high TOC:TN ratios and ^{13}C - and ^{15}N - depleted compositions with respect to SOM (Meyers, 1997). An equivalent pattern is displayed for the coffee field plot (not shown here, see in Bellanger

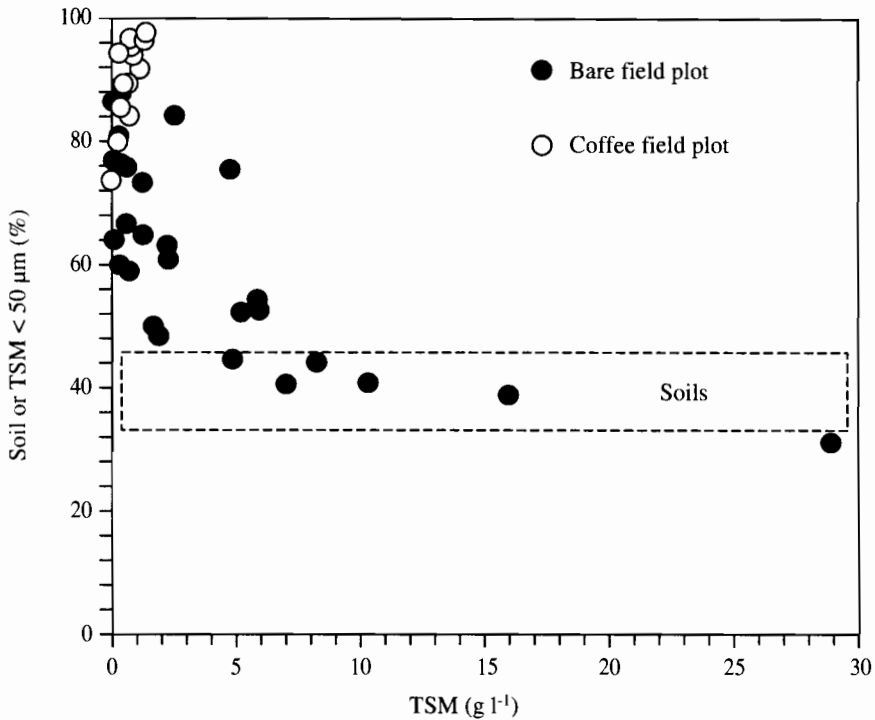


Figure 20.8 Plots of fine-size particles in suspended sediments vs. TSM delivery by soils of two experimental field plots of the Rio Boconó watershed (Venezuela) during a single rainfall event (June 14, 1998). Soil composition is indicated by the dashed line. TSM = total suspended matter concentration. (From Bellanger, B., S. Huon, F. Velasquez, V. Vallès, C. Girardin, and A. Mariotti. 2004a. Experimental $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ study of soil organic carbon loss by erosion in the Venezuelan Andes. *Catena*, 58, 125–150. With permission.)

et al., 2004b). For higher runoff and suspended sediment concentrations, approximately above 30 ml s^{-1} and 0.5 g l^{-1} (95% of suspended matter yield for the bare field plot), respectively, the composition of fine-size ($< 50 \mu\text{m}$) organic matter evolves toward constant values that closely reflect the composition of SOM, fingerprinting isotopic contrasts inherited from the vegetation cover. A constant composition for fine-size ($< 50 \mu\text{m}$) suspended particles can be assumed for most of the sediment yield, which provides quantitative information on SOC erosion. Monitoring the release of POC from coarser size fractions will only be meaningful for steady state conditions and during periods of high runoff. Sorting effects related to erosion intensity suggest that only fine-size organic fractions transported in stream flows will accurately reflect the contribution of soil fine fractions.

20.4.2.3 Monitoring Sources of Fine-Size ($< 50 \mu\text{m}$, $< 200 \mu\text{m}$) Suspended Organic Carbon in Stream Flows during Flood Events

Monitoring changes in the source of organic matter in suspended sediments of stream flows was carried out for two flood episodes (Cornières, 1998; Bellanger, 2003). The first event (June 14, 1998) of medium intensity had a return period of 1 year and generated for the main stream, the Boconó River before the confluence with the Burate (Figure 20.2a), a maximum water discharge of $110 \text{ m}^3 \text{ s}^{-1}$ (Figure 20.10a). During the second event (July 9 to 10, 1995), water discharges rose to $240 \text{ m}^3 \text{ s}^{-1}$ and $290 \text{ m}^3 \text{ s}^{-1}$ for the Rio Boconó and the Rio Burate (Fig. 20.11a), respectively, with a return period of 10 years. Total suspended sediment concentrations (TSM) and water discharges (Q) are correlated through power functions with no hysteresis ($\text{TSM} = 5 \cdot 10^{-4} \times \text{Q}^{1.9}$, r^2

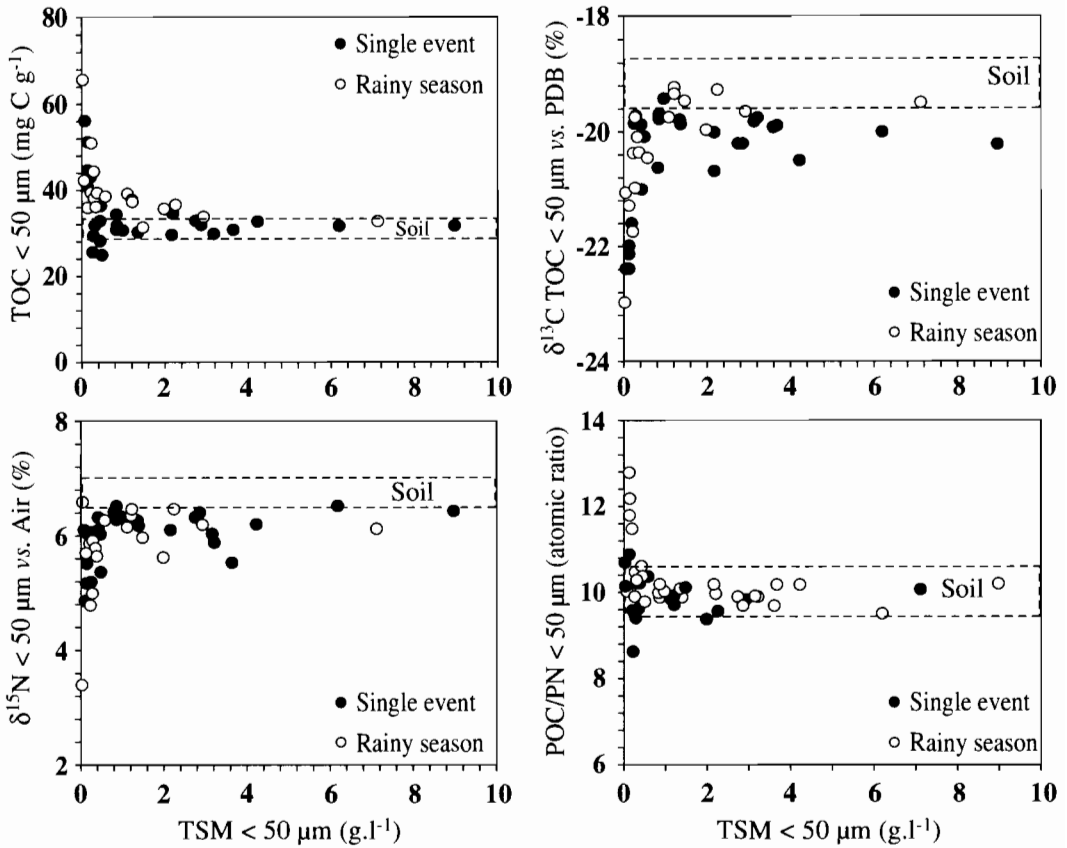


Figure 20.9 Plots of changes in TOC, $\delta^{13}\text{C}$, $\delta^{15}\text{N}$ and TOC/TN in fine-size suspended sediments released by soils of the bare field plot on the Rio Boconó watershed (Venezuela). Soil composition is indicated by the dashed line. TSM = total suspended matter concentration. (From Bellanger, B., S. Huon, F. Velasquez, V. Vallès, C. Girardin, and A. Mariotti. 2004a. Experimental $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ study of soil organic carbon loss by erosion in the Venezuelan Andes. *Catena*, 58, 125–150. With permission.)

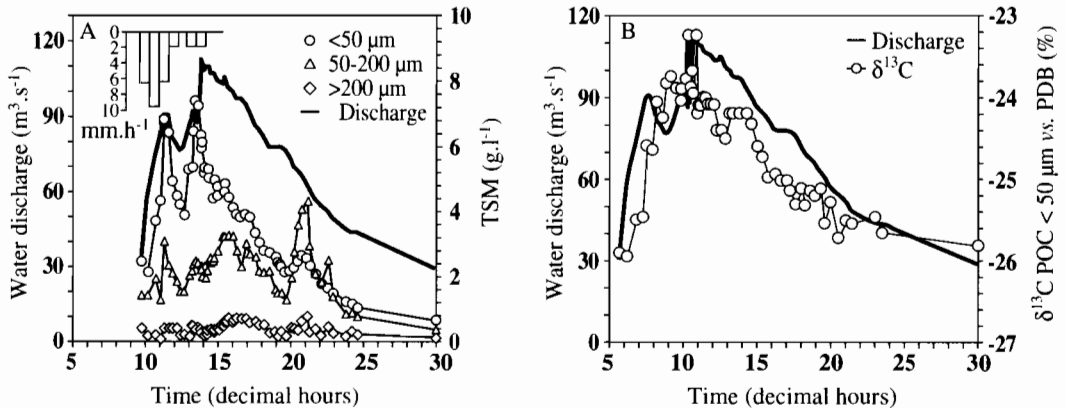


Figure 20.10 Plots of water discharge and (a) suspended matter grain size compositions and (b) $\delta^{13}\text{C}$ of fine-size POC for the Rio Boconó (Venezuela) during the June 14, 1998 flood event. TSM = Total suspended matter concentration. Precipitations are reported in mm h^{-1} .

= 0.97 for the Rio Burate in 1995; $TSM = 3.5 \cdot 10^{-3} \times Q^{1.35}$, $r^2 = 0.77$, and $TSM = 4 \cdot 10^{-2} \times Q^{1.14}$, $r^2 = 0.69$, for the Rio Boconó in 1995 and 1998, respectively). These relationships indicate that erosion along slopes of the watershed and remobilisation of riverbank sediments was synchronous to water discharge (Nordin, 1985). Therefore, monitoring the composition of POC in suspended sediments may provide both information on erosion intensity and on the contributing sources of organic carbon if they have sufficiently contrasted isotopic signatures.

During the June 1998 flood event of the Rio Boconó, fine- (< 50 μm) and coarse- (> 200 μm) size fractions account for 60 and 6% of total suspended sediment load, respectively. The $\delta^{13}\text{C}$ composition of fine (< 50 μm) suspended organic matter varies linearly with suspended sediment concentration (Figure 20.10b and Figure 20.12a). This relationship reflects a correlative change with erosion intensity of the type of suspended organic matter. For high suspended matter concentrations (high erosion levels) the source of organic C is most likely derived from SOM, characterized by enriched ^{13}C compositions with respect to depleted ^{13}C compositions for vegetation debris (Meyers, 1997). This interpretation is also supported by $\delta^{15}\text{N}$ changes in the composition of fine-size (< 50 μm) suspended sediments. When plotted in a $\delta^{15}\text{N}$ - $\delta^{13}\text{C}$ mixing diagram (Fig. 20.12c; see discussion in Ganeshram et al., 2000), suspended POC compositions are scattered between a vegetation-derived organic matter pool with $\delta^{15}\text{N}$ values close to atmospheric N_2 composition (0‰) and soil-derived organic matter pools enriched in ^{15}N with respect to the composition of vegetation debris (Figure 20.12c). Additional information may also be derived from POC concentrations in fine-size suspended matter. Low POC contents reflect enhanced supply of SOM during peak flow and falling water stages whereas; high POC concentrations involve more vegetation debris during rising and falling water stages (Figure 20.12b). However, precise monitoring of the contribution of each SOM pool may still remain difficult because the $\delta^{13}\text{C}$ values (range, $-25.5\% < \delta^{13}\text{C} < -23.5\%$) that can be assigned to soil and vegetation debris end-members may overlap. For this flood event that lasted 24 hours, we derived SOC exportation budgets of $166 \cdot 10^6 \text{ gC}$ and $139 \cdot 10^6 \text{ gC}$ for < 200 μm and < 50 μm size fractions of suspended sediments, respectively. Scaling up these values to the upper Boconó catchment's (574 km^2) provides specific erosions of 0.29 gC m^{-2} for < 200 μm size fractions and 0.24 gC m^{-2} for < 50 μm size fractions. These budgets are in the range of the organic C budgets derived from the field plots experiments (1.3 gC m^{-2} and 0.03 gC m^{-2} for the bare and the coffee field plots, respectively; see above) carried out during the same flood event (June 14, 1998). Given that (1) less than 20% of the catchment is occupied by bare fields (excluding outcrops of geological basement rocks) and (2) the experiments carried out for bare and coffee field plots account for high and low erosion conditions with respect to cultivated and fallow lands, monitoring fine-size suspended fractions in stream flows provides organic C budgets that are consistent with the results obtained with the field plot experiments. It is therefore possible to derive SOC erosion budgets from fine-size suspended POC concentrations in stream flows as far as only top SOM is involved by erosion processes and transported by runoff to stream flows during the flood event.

During the July 1995 flood event, the discharge of the Rio Burate rose by a factor 10 whereas the suspended sediment concentration increased by a factor 150 (Figure 20.11a). This event is more intense than the previous one. The fine-size (< 200 μm) fraction accounts for 93% of total suspended sediment yield. Monitoring ^{15}N abundances in these fine sediments with $\delta^{15}\text{N}$ measurements provides minimum values when water discharge is maximal, indicating a concomitant change in the composition of suspended sediments (Figure 20.11b). When plotted against suspended matter concentration, TOC, TOC/TN, and $\delta^{15}\text{N}$ values follow decreasing hyperbolic trends with increasing water discharge and suspended sediment load (Figure 20.13a–b). The decrease of TOC concentrations in suspended sediments is usually interpreted as a dilution trend toward a minimum value (ca. 4 to 5 mgC g^{-1} in our study) that either reflects: (1) enhanced contribution of mineral particles (i.e., quartz, micas, etc.) with respect to organic matter, derived from erosion of the geological basement on the watershed (Meybeck, 1982); or (2) supply of SOM from deeper soil horizons, more depleted in organic C than top soil horizons (Ludwig et al., 1996). In our study, the TOC/TN

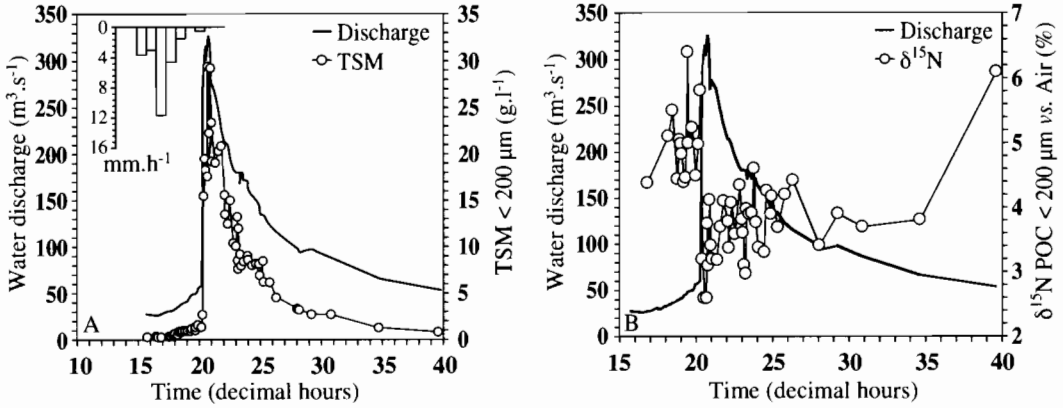


Figure 20.11 Plots of water discharge and (a) TSM delivery and (b) $\delta^{15}\text{N}$ of suspended fine-size (< 200 μm) fractions for the Rio Burate (Rio Boconó watershed, Venezuela) during the July 9 and 10, 1995 flood event. TSM = total suspended matter concentration.

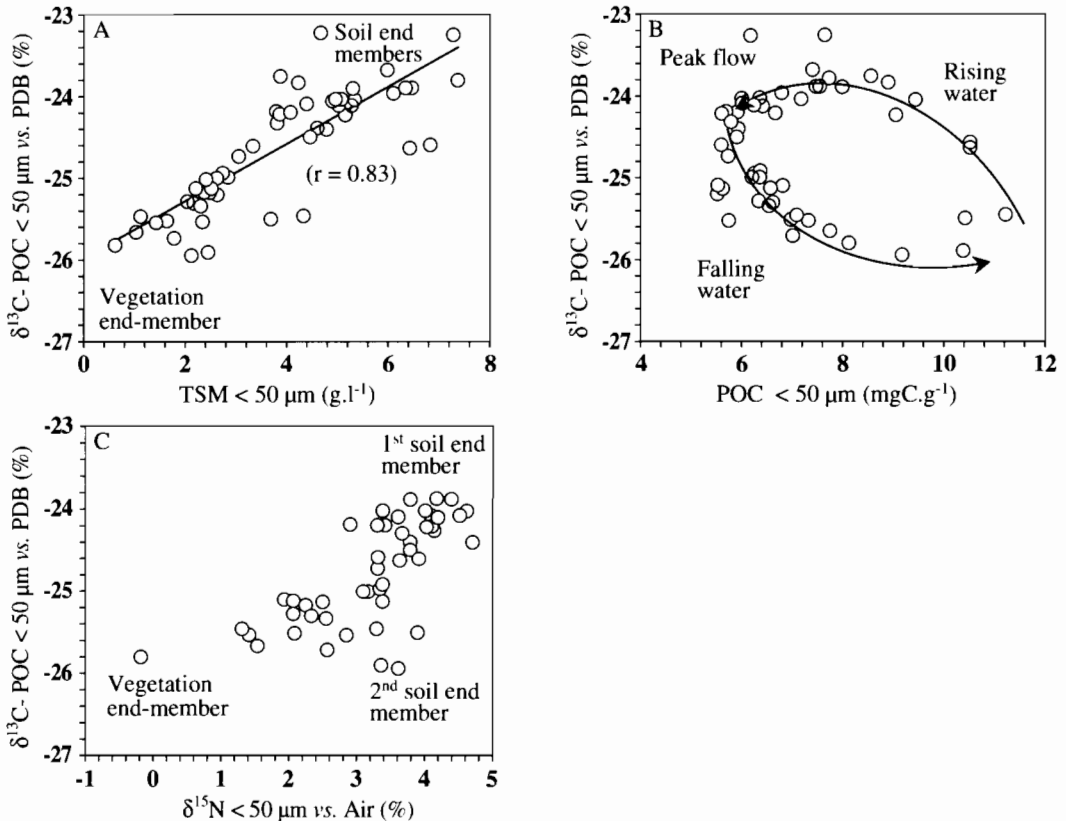


Figure 20.12 Plots of $\delta^{13}\text{C}$ values vs. (a) TSM, (b) POC, and (c) $\delta^{15}\text{N}$ for fine (< 50 μm) suspended fractions during the June 14, 1998 flood event of the Rio Boconó (Venezuela). TSM = total suspended matter concentration. Estimated soil and vegetation end-members compositions are reported.

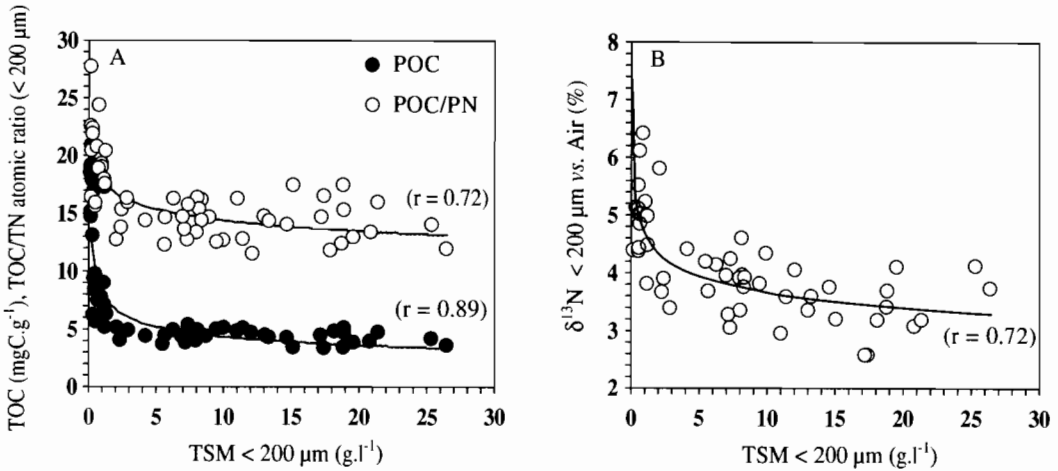


Figure 20.13 Plots of: (a) TOC and TOC/TN vs. TSM and B) $\delta^{15}\text{N}$ vs. TSM for fine-size (< 200 μm) suspended fractions during the June 14, 1998 flood event of the Rio Burate (Boconó watershed, Venezuela). TSM = total suspended matter concentration.

Table 20.3 Average Composition of Fine Organic Matter in Soils, Vegetation, and Rocks of the Rio Burate Watershed

< 200 μm	TOC (mgC g^{-1})	TN (mgN g^{-1})	TOC/TN	$\delta^{15}\text{N}$ vs. AIR (%)	$\delta^{13}\text{C}$ vs. PDB (%)
Soils (0 to 20 cm)	35 ± 10	3 ± 1	22 ± 7	5.5 ± 0.5	Variable*
Vegetation debris	460 ± 50	30 ± 10	17 ± 7	0 ± 2	Variable*
Shales-lutites	15 ± 3.8	1.0 ± 0.2	14 ± 5	2.5 ± 0.5	-25.5 ± 0.3

* Due to occurrence of both C_3 and C_4 plant communities on the watershed.

and $\delta^{15}\text{N}$ of fine (< 200 μm) organic matter of carboniferous shales (organic-matter bearing rocks; Table 20.3) located along the banks of the Rio Burate (Figure 20.2c) match those obtained during the period of high suspended sediment transport corresponding to high erosion stages. SOM cannot be the exclusive supplier of organic C because TOC/TN and $\delta^{15}\text{N}$ values in suspended sediments are higher and lower, respectively, than for soil surface horizons. The contribution of the different pools of organic matter can be quantified using isotopic mass balance equations if suspended sediments that compose the mixtures derive from end-members with contrasted compositions (Balesdent and Mariotti, 1996). The contribution of particulate nitrogen sources of vegetation debris may be assumed to be negligible with respect to the high turbidity of stream flow waters. Meaningful discrimination between lithic and soil sources of suspended organic matter is only obtained with TOC/TN and $\delta^{15}\text{N}$ values. The use of ^{13}C abundance measurements might be confusing because SOM might be inherited from either C_3 or C_4 plants and, because long-term cultivation also involves rotating fallow periods and local replacement of C_3 by C_4 plant covers and vice versa.

In an ideal mixture between soil- and rock-derived organic particles, the instantaneous N and $\delta^{15}\text{N}$ contributions of the two end-members in suspended (< 200 μm) sediments can be expressed by the following equations:

$$N_{\text{sample}} = N_{\text{soils}} + N_{\text{rocks}} \quad \text{with} \quad fN_{\text{rocks}} = N_{\text{rocks}}/N_{\text{sample}} = 1 - fN_{\text{soils}}$$

$$N_{\text{sample}} \times \delta^{15}\text{N}_{\text{sample}} = (N_{\text{soils}} \times \delta^{15}\text{N}_{\text{soils}}) + (N_{\text{rocks}} \times \delta^{15}\text{N}_{\text{rocks}})$$

$$\delta^{15}\text{N}_{\text{sample}} = [(1 - fN_{\text{rocks}}) \times \delta^{15}\text{N}_{\text{soils}}] + (fN_{\text{rocks}} \times \delta^{15}\text{N}_{\text{rocks}})$$

$$fN_{\text{rocks}} = (\delta^{15}N_{\text{sample}} - \delta^{15}N_{\text{soils}}) / (\delta^{15}N_{\text{rocks}} - \delta^{15}N_{\text{soils}})$$

The fraction of organic carbon derived from rock erosion (fC_{rocks}) can be calculated using the C:N specific ratios (equivalent to TOC/TN in Table 20.3) of each end-member in the mixture:

$$fC_{\text{rocks}} = C_{\text{rocks}} / C_{\text{sample}} = [N_{\text{rocks}} \times (C/N)_{\text{rocks}}] / [N_{\text{sample}} \times (C/N)_{\text{sample}}]$$

$$fC_{\text{rocks}} = fN_{\text{rocks}} \times [(C/N)_{\text{rocks}} / (C/N)_{\text{sample}}]$$

According to these calculations, the contribution of rock-derived particles (fC_{rocks}) in suspended sediments accounts for 0 to 80% of the (< 200 μm) POC transported during the flood event (Figure 20.14). Erosion of the geological basement is maximal during the period of high water discharge of the Rio Burate, reflecting enhanced rock and sediment contribution to the suspended load. However, these estimates bear a high uncertainty, mainly due to a rather low discrimination between the average $\delta^{15}N$ values of the two end-members ($\Delta\delta^{15}N = 3\%$, Table 20.3), with respect to the $\delta^{15}N$ variability in each pool ($\sigma = 0.5\%$, Table 20.3). The overall average weighted contribution of fine organic C derived from rocks accounts for 50% of the suspended (< 200 μm) POC load. During the flood event that lasted 24 hours, 254 10⁶ gC were exported in the < 200 μm size fraction of suspended sediments on the Rio Burate watershed (364 km²). Specific organic C erosion was ca. 0.70 gC m⁻², approximately half of it (0.35 gC m⁻²) originating from organic matter bearing rocks. As already outlined by Meybeck (1993) on a more global scale, mountainous catchments with large outcrops of sedimentary to low-grade metamorphic rocks may provide a significant fraction of POC to river discharge and bias organic carbon budgets solely based on SOM erosion. Moreover, the behavior in surface environments of more refractory organic matter released from geological basement sources (such as shales or other organic-matter bearing rocks) will contrast sharply with that of vegetation debris, charcoals or humic substances generated by soil erosion.

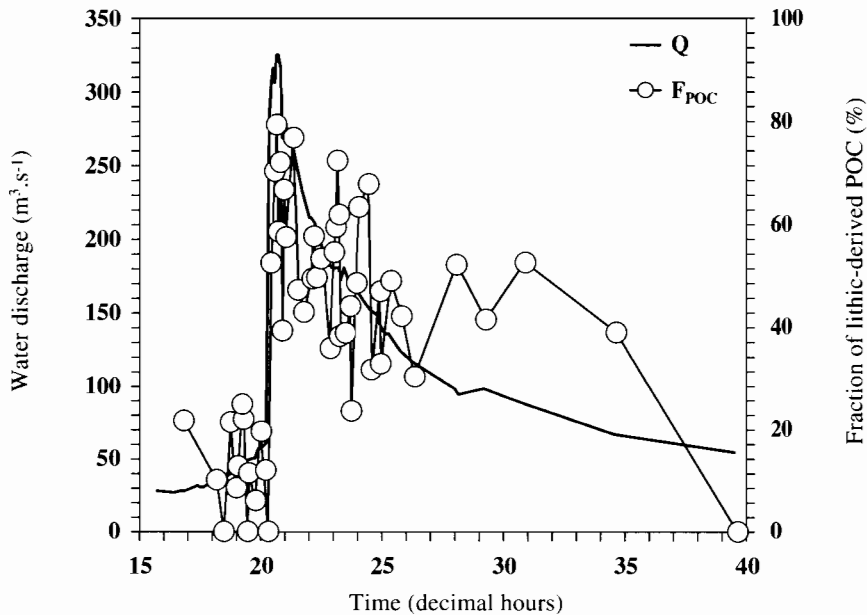


Figure 20.14 Plot of water discharge and fraction of lithic-derived POC for fine-size (< 200 μm) suspended fractions during the June 14, 1998 flood event of the Rio Burate (Boconó watershed, Venezuela). TSM = total suspended matter concentration.

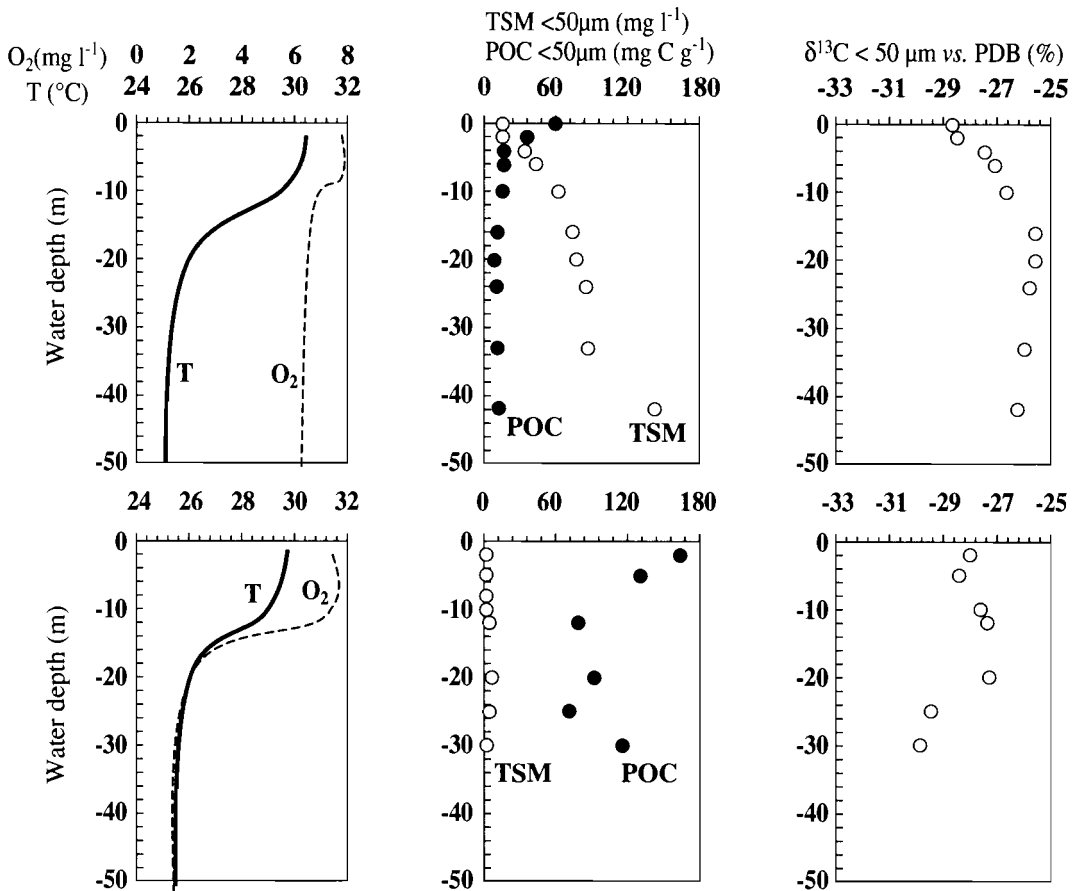


Figure 20.15 Plot of dissolved oxygen (O_2), temperature (T), total suspended matter and POC concentrations (TSM and POC), $\delta^{13}C$ for fine size fractions in the water column of the Boconó reservoir (upper graphs) and the Tucupido reservoir (lower graphs) in September 1999. TSM = total suspended matter concentration.

20.4.2.4 Monitoring the Evolution of Suspended Organic Matter in a Water Reservoir at the Outlet of the Watershed

The Peña Larga water dam can be divided into two parts, the Boconó and the Tucupido reservoirs that communicate through a narrow and shallow channel with a maximum depth of 20 m (Figure 20.2b). Thermal stratification in both reservoirs was stable throughout the year with, oxic conditions (dissolved $O_2 > 4.5 \text{ mg} \cdot \text{l}^{-1}$, water residence time: $\tau = 0.36$ years, Bellanger et al., 2004b) and hypoxic to anoxic conditions (dissolved $O_2 < 1 \text{ mg} \cdot \text{l}^{-1}$, water residence time $\tau = 6.2$ years; Bellanger et al., 2004b) below the thermocline, for the Boconó and Tucupido reservoirs, respectively. Suspended sediment loads are mainly controlled by Boconó River discharge with high and depth increasing concentrations for Boconó Reservoir and low and constant concentrations for Tucupido Reservoir. The composition of fine-size ($< 50 \mu\text{m}$) suspended organic matter in the oxic reservoir reflects that of the sources of suspended sediment in the water column: i.e., surface productivity above the thermocline (more depleted in ^{13}C , enriched in organic C, Figure 20.15) and suspended organic matter derived from erosion on the watershed below the thermocline (Figure 20.15; average composition for the Boconó river: $\delta^{13}C = -25.8\%$, $[\text{TOC}] = 10.3 \text{ mgC} \cdot \text{g}^{-1}$). In contrast, fine suspended sediments in the water column of the hypoxic to anoxic reservoir are more enriched in organic C and depleted in ^{13}C , indicating additional contribution of chemoautotrophic or methanotrophic

biomass to suspended material load (Harvey and Macko, 1997; Hollander and Smith, 2001; Lehmann et al., 2002). POC budgets indicate that during the past two decades erosion on the Boconó watershed involved the transfer of $32 \times 10^9 \text{ gC yr}^{-1}$ to the Peña Larga water dam (ca. $19.8 \text{ gC m}^{-2} \text{ yr}^{-1}$; Cornières 1998). Of this, approximately $21 \pm 5 \times 10^9 \text{ gC yr}^{-1}$ (66%) was buried in bottom lake and alluvial fan sediments and only ca. $2 \times 10^9 \text{ gC yr}^{-1}$ (6%) was exported from the water dam to the Llanos floodplain. A net amount of $8 \pm 5 \times 10^9 \text{ gC yr}^{-1}$ (ca. 25%, not including ca. $1 \times 10^9 \text{ gC}$ involved in the suspended load) has been mineralized in the Boconó basin (Bellanger et al., 2004b). These organic C burial rates are one order of magnitude higher than those usually reported for small lakes (Einsele et al., 2001) due to relatively high erosion rates on the watershed. Although providing a significant flux of dissolved C produced by in situ oxidation of suspended organic matter, the Peña Larga water reservoir reduced by nearly 94% the POC flux generated by erosion on the Rio Boconó watershed.

20.5 CONCLUSION

Monitoring the composition of organic C in soils and suspended sediments of stream flows with isotopic tracers provides significant information that may help to better constrain SOC erosion budgets. The redistribution pattern of fallout ^{137}Cs and the SOC content in the first 30 cm of cultivated soils of the Houay Pano catchment (Laos) are significantly linked. A common process apparently relates the soil erosion status with the amount of organic matter in the topsoil horizons. However, SOC erosion rates, estimated using total SOC depletion since 1963, are higher than those calculated with mass balance models based on ^{137}Cs inventories. With mass balance model 2 (Walling and He, 2001), erosion and accumulation rates along two toposequences averaged ($\pm 1\sigma$): $-0.43 \pm 0.25 \text{ kg m}^{-2} \text{ yr}^{-1}$ and $0.13 \pm 0.11 \text{ kg m}^{-2} \text{ yr}^{-1}$, respectively, vs. $-1.28 \pm 0.66 \text{ kg m}^{-2} \text{ yr}^{-1}$ and $0.66 \pm 0.46 \text{ kg m}^{-2} \text{ yr}^{-1}$, respectively, using TOC – ^{137}Cs inventories. The difference reflects additional organic C losses controlled by mineralization of SOM through respiration and dissolved organic C releases by runoff. Monitoring SOC erosion with ^{137}Cs may help to better constrain soil leaching and mineralization rates by subtracting from SOC inventories the effect of particles redistribution along slopes due to runoff and tillage erosion. The reliability of these estimates is mainly linked to the reference ^{137}Cs inventory used to discriminate erosion and deposition sites by comparison with undisturbed soils and to the cultivation depth used to compute soil erosion rates with mass balance models.

Monitoring fine-size ($< 50 \mu\text{m}$ or $< 200 \mu\text{m}$) suspended organic matter transported by runoff and stream flows with $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ measurements may allow quantification of the contributing pools of organic matter if the isotopic signature of each end-member in the mixture is sufficiently discriminating. During an important flood event, the contribution of particulate organic C derived from incisive erosion of organic matter bearing rocks of the geological basement of the Rio Boconó watershed (Venezuela) could be distinguished from that of topsoil horizons using $\delta^{15}\text{N}$ measurements on fine suspended sediments. The contribution from lithic sources was approximately equivalent to that of soils, which indicates that soil erosion budgets and models that only account for the contribution of SOC might be partly biased. The transport of fine-size ($< 50 \mu\text{m}$ or $< 200 \mu\text{m}$) suspended organic matter in stream flows during flood events was apparently conservative for different scales (30 m^2 , 364 km^2 , and 574 km^2) with soil and SOC specific erosions in the same order of magnitude, between those of bare fields (high soil erosion conditions) and coffee fields (low soil erosion conditions) end-members (Table 20.4).

Since fine-size fractions account for most of the organic C stored in soil surface horizons, monitoring fine organic matter in runoff and stream flows may provide significant constraint on SOC erosion, possibly allowing to scale up field plot measurements to catchments with appropriate grain size correction factors, which still need to be determined. Although water quality might be affected by high-suspended matter loads derived from erosion on the watershed, the extent of organic C released in the hydrographical network of the floodplain can be thoroughly reduced by

Table 20.4 Selected Soil and SOC Erosion Estimates for the Rio Boconó Watershed (Venezuela)

	Bare Soil (30 m ²)		Soil under Coffee (30 m ²)		Rio Boconó (574 km ²)	Rio Burate (364 km ²)
	Season Flood Event	Season Flood Event	Season Flood Event	Season Flood Event	Flood Event	Flood Event
Soil loss (kg m ⁻²)	1.8	0.09	0.05	0.0007	0.06	0.17
< 200 µm] (wt%)	—	62	—	96	90	93
< 50 µm] (wt%)	—	45	—	93	56	—
POC loss (gC m ⁻²)	—	—	—	—	—	—
< 200 µm (gC m ⁻²)	—	1.3	—	0.03	0.29	0.35
< 50 µm (gC m ⁻²)	29.5	1.2	2.3	0.03	0.24	—

water reservoirs located at the outlet of the watersheds. The Peña Larga water reservoir reduced by nearly 94% the suspended organic matter load discharged by the Rio Boconó through sediment burial and in situ mineralization processes and behaved as a major sink for C with respect to suspended POC supply from the watershed.

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On the cover: A typical landscape of red ferrallitic soils on the high plateau of central Madagascar near Antananarivo during the rainy season. The hilltop is covered by overgrazed grassland deeply eroded around the cattle trails which join the village and the springs in the valley. The hills lose carbon, nutrients, soil, and water, but might be nourishing the rice paddies below.

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SOIL EROSION AND CARBON DYNAMICS

In addition to depleting nutrients necessary for healthy crops, soil erosion processes can affect the carbon balance of agroecosystems, and thus influence global warming. While the magnitude and severity of soil erosion are well documented, fluxes of eroded carbon are rarely quantified. **Soil Erosion and Carbon Dynamics** brings together a diverse group of papers and data from the perspectives of world-renowned soil scientists, agronomists, and sedimentologists to resolve whether soil erosion on carbon is a beneficial or destructive process.

This book collects quantitative data on eroded organic carbon fluxes from the scale of the agricultural plot to that of large basins and oceans. It quantifies the magnitude of eroded carbon for different soil management practices as compared to normal carbon sequestration and discusses the fate of the eroded carbon and whether or not it is a source or sink for atmospheric CO₂. Finally, the book offers data reflecting the impact of soil erosion on soil, water, and air quality. Other important topics include solubilization, carbon transfer, and sediment deposition, as well as carbon dioxide emissions, global warming potential, and the implications of soil erosion on the global carbon cycle and carbon budget.

Features

- Defines basic concepts and general approaches to the global carbon cycle, carbon sequestration, erosion, and eroded carbon
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