



# Cropland C erosion and burial: Is buried soil organic matter biodegradable?



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## ABSTRACT

Erosion of soil in cropland results in the decoupling of soil organic matter (SOM) from its location of provenance. Deposition of soil downslope from eroded landscape areas can cause burial of SOM deep in the soil profile, often below the zone of cultivation. We assessed and compared the biodegradability of SOM from different landscape areas at six locations across Canada. We evaluated buried SOM located below the existing cultivated layer in depositional areas and SOM in the soil profile of eroded landscape areas. We measured the C in light fraction (LF) SOM, and conducted <sup>14</sup>C analysis and laboratory incubations on the whole soil. Measurement of <sup>14</sup>C indicated that most of the surface and buried soils contained significant quantities of post-bomb C (i.e., C fixed from the atmosphere by plants and incorporated into the soils over the last 60 years). Buried soils contained equal or greater masses of LF C than that in surface soils, indicating the presence of C which should have been readily-decomposable to microorganisms. However, biodegradable C in buried soils was less than that in the surface soils even though the buried soils usually had equal or larger quantities of total soil C and LF C. Our results suggest that the deposition of soil in depositional landscape positions in eroded landscapes constrained decomposition and mineralization of SOM. This suggests that erosion and depositional processes may be responsible for creating a net sink of C in these landscape positions. Previously eroded landscape areas that had lost their SOM to lower landscape areas appeared to be gradually accumulating recent C from the atmosphere, while the redistributed SOM is being retained in the system through burial and protection in depositional landscape areas.

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## 1. Introduction

Soil organic matter (SOM) located deep in the soil (below 30 cm) may represent over half of the total organic C stock in many areas of the world (Rumpel and Kögel-Knabner, 2011; Tarnocai et al., 2009; Chaopricha and Marin-Spiotta, 2014). Mechanisms by which C is incorporated to depths below 30 cm include deposition by plant roots and exudates, movement of dissolved organic matter through preferential flow channels, and bioturbation (Rumpel and Kögel-Knabner, 2011). Redistribution of soil in agricultural landscapes by erosion and deposition is another mechanism by which SOM is buried. Erosion by water and tillage processes results in the redistribution of SOM, whereby upper slopes can lose a majority of their C while footslope and toeslope landscape positions can gain large quantities of soil C (Van Oost et al., 2007; VandenBygaart et al., 2012). This redistribution of soil C may be an important mechanism in global C dynamics, but studies assessing the net effect on atmospheric C have not been conclusive (Van Oost et al., 2007; Berhe et al., 2007).

Erosion and redistribution by water, wind and tillage processes can move large quantities of SOM (VandenBygaart et al., 2012; Van Oost

et al., 2007). These processes occur in varying degrees across agricultural landscapes because over 60% of the earth's land surface consists of topography with greater than 8% slope (Berhe and Kleber, 2013). A recent study showed that depositional landscape areas can gain from 2 to 4 Mg C ha<sup>-1</sup> yr<sup>-1</sup> and some areas can gain over 200 Mg C ha<sup>-1</sup> in 50 years of agricultural management (VandenBygaart et al., 2012). This deposition can lead to the burial of C-rich A horizon material deep in the soil profile, and these profiles can exceed 90 cm in depth, potentially burying huge quantities of SOM. A consequence of these erosion processes is the decoupling of SOM from its original source of assimilation such that convex upper landscape areas have reduced SOM due to soil loss, while lower landscape positions can have deep soil profiles where soil has been deposited. Since this soil contains a large proportion of the stock of SOM in the landscape, the nature and fate of this buried material are of interest due to its potential influence on C dynamics (Rumpel and Kögel-Knabner, 2011; Berhe et al., 2007; VandenBygaart et al., 2012).

Recent studies have indicated that buried SOM in depositional environments is more resistant to decomposition than soils where erosion and deposition are not occurring. Doetterl et al. (2012) showed that buried SOM had a higher mean residence time than surface soils, and that this was due to its physical protection within

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microaggregates and association with silt-sized mineral particles. Wang et al. (2014) concluded that the majority of depositional SOC in some Belgian landscapes responds to burial by decomposing over centennial timescales.

The LF represents an actively cycling pool of SOM containing partially decomposed plant residues and microbial biomass, and is relatively easily decomposable compared to other portions of the SOM (Janzen et al., 1992; Gregorich et al., 1995). This light material, along with fine minerals, is preferentially moved in erosional processes and accounts for a substantial amount of SOM. In long-term laboratory incubations Haile-Mariam et al. (2008) observed that the LF was labile and susceptible to decomposition with about 65% of it lost over 800 days, and the majority lost within 400 days.

VandenBygaart et al. (2012) sampled soil profiles along transects from upper to lower landscape areas in agricultural soils of Canada to assess the recent effects of soil erosion and redistribution on soil C dynamics. Utilizing  $^{137}\text{Cs}$  as a dating tool they were able to reconstruct the landscape history of soil and soil C redistribution in the past half century. Here we expand that study by assessing the biodegradability of buried SOM located below the existing cultivated layer in depositional areas, as well as that within the soil profile in eroded areas. This was achieved through measuring LF SOM,  $^{14}\text{C}$  analysis and laboratory incubations of whole soil. We hypothesized that the LF would comprise the majority of labile C and would provide information about the biodegradability of the SOM that mineralized under ideal conditions over a 75 day laboratory incubation.

## 2. Methods

### 2.1. Field sampling

The six sites were located in six different provinces in Canada (Alberta, Saskatchewan, Manitoba, Ontario, New Brunswick, P.E.I.) in fields where soil erosion and redistribution were observed and measured (Table 1; Table 2) (VandenBygaart et al., 2012). A full description of the field methods and site properties is given in VandenBygaart et al. (2012). Briefly a minimum of four soil profiles were excavated at different landscape positions along hillslopes including positions where soil erosion and deposition were evident. At each landscape position a soil pit was excavated and four replicate bulk samples (~1 kg, wet weight) were sampled in 5 cm increments to beneath the A horizon at each of the profile locations. Triplicate rings were extracted for bulk density measurements at all sample depths. The collected soils were air-dried, passed through a 2-mm sieve, and weighed. The bulk density of each layer was determined using the core volume and mass of soil for each layer. Textural analysis was conducted using the pipette method (Kroetsch and Wang, 2008). Approximately 20 g of each soil sample was dried at 60 °C and ground to <250  $\mu\text{m}$  using a puck mill vibratory pulverizer (Gy-Ro Mill 4299, Aimil Ltd, New Delhi). Soils were pre-screened to determine whether inorganic C was present and if present, the samples were subjected to a 6 M HCl washing and recovery prior to elemental analysis. Total soil C and N concentrations were determined on a Carla Erba elemental analyzer (NC 2500, Carlo Erba Instruments, Milan, Italy). Soil organic C storage was determined for each depth increment using the measurements of C concentration and bulk density.

**Table 1**

Locations of study catenae.

Province	Location	Latitude (N)	Longitude (W)	Cropping system	soil classification
Alberta	Mundare	53° 33' 44"	112° 17' 39"	Wheat	Orthic Black Chernozem
Saskatchewan	St. Denis	52° 12' 46"	106° 5' 52"	Wheat	Orthic Dark Brown Chernozem
Manitoba	Miami	49° 19' 33"	98° 22' 58"	Canola	Calcareous Black Chernozem
Ontario	Elora	43° 38' 59"	80° 24' 36"	Corn/soybean	Orthic Gray Brown Luvisol
New Brunswick	Grand Falls	47° 00' 6"	67° 41' 20"	Potato/forage	Orthic Humo-Ferric Podzol
P.E.I.	Middleton	46° 19' 6"	63° 34' 36"	Potato/forage	Orthic Humo-Ferric Podzol

**Table 2**

Background information for each of the eroded and aggraded soil profiles at each of the six study locations.

Location	Profile <sup>a</sup>	Erosion status	A-horizon thickness <sup>b</sup> (cm)	Deposition Rate <sup>b</sup> (cm yr <sup>-1</sup> )
Alberta	ALTA-1	Eroded	10	
	ALTA-2	Aggraded	60	0.31
Saskatchewan	SASK-1	Eroded	14	
	SASK-2	Aggraded	37	0.25
	SASK-3	Aggraded	81	0.55
Manitoba	MAN-1	Eroded	18	
	MAN-3	Aggraded	75	0.73
Ontario	ONT-1	Eroded	25	
	ONT-2	Aggraded	49	0.36
New Brunswick	NB-2	Eroded	35	
	NB-4	Aggraded	90	1.11
P.E.I.	PEI-3	Eroded	22	
	PEI-4	Aggraded	48	0.36

<sup>a</sup> Profile label used in VandenBygaart et al. (2012).

<sup>b</sup> Derived by VandenBygaart et al. (2012).

### 2.2. Radiocarbon

The degree to which bomb-fallout  $^{14}\text{C}$  is found in SOM can provide a direct measure of the C that has been assimilated from the atmosphere in the past 50 years (Trumbore, 2000).  $\Delta^{14}\text{C}$  is used to express bomb-sourced radiocarbon, and it represents the difference between the  $^{14}\text{C}/^{12}\text{C}$  ratio in the sample compared to that of a universal standard (oxalic acid, decay corrected for the year 1950) (Gaudinski et al., 2000). There are two timescales over which radiocarbon can be used as a dating tool. Firstly, radioactive decay of the  $^{14}\text{C}$  atoms to  $^{14}\text{N}$  occurs with a half-life of 5730 years, and thus measuring the quantity of  $^{14}\text{C}$  in a sample can yield an age for when the C was assimilated from the atmosphere (Trumbore, 2000). Secondly, during thermonuclear atmospheric bomb-testing in the Northern Hemisphere in the late 1950s and early 1960s, there was a sharp increase in the concentration of  $^{14}\text{C}$  in the atmosphere. A positive  $\Delta^{14}\text{C}$  depicts the presence of  $^{14}\text{C}$  produced by thermonuclear bomb-tests from 1955 to 1963 (Hua, 2009), suggesting that at least some C in the sample was extracted from the atmosphere after 1950 (Trumbore, 2009). Negative  $\Delta^{14}\text{C}$  values indicate that the C has been present long enough for significant radioactive decay to occur (Trumbore, 2000). The SOM consists of material with a broad range of ages and thus a  $\Delta^{14}\text{C}$  value is a result of an average of C that has exhibited radioactive decay and that which has been derived from recent bomb testing. Consequently  $\Delta^{14}\text{C}$  may be a useful tool for reconstructing historical C dynamics in eroded landscapes.

### 2.3. $^{14}\text{C}$ characterization

$^{14}\text{C}$  measurement was conducted by Beta Analytic Inc. of Miami, FL, USA by accelerator mass spectrometry (AMS). Samples were pretreated with HCl repeatedly to ensure the absence of carbonates. We report the radiocarbon data as  $\Delta^{14}\text{C}$ , corrected for mass-dependent isotopic fractionation using  $\delta^{13}\text{C}$  where the standard is 95% of the activity of NBS Oxalic Acid (Stuiver and Polach, 1977).  $\Delta^{14}\text{C}$  is zero when the sample has the same isotopic signature as the atmosphere had in 1950.

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