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Source: Arctic, Antarctic, and Alpine Research, 37(1) : 108-117

Published By: Institute of Arctic and Alpine Research (INSTAAR), University of Colorado

URL: https://doi.org/10.1657/1523-0430(2005)037[0108:PSOMTI]2.0.CO;2

Arctic, Antarctic, and Alpine Research, Vol. 37, No. 1, 2005, pp. 108–117

Potential Soil Organic Matter Turnover in Taylor Valley, Antarctica

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Abstract

Antarctic Dry Valley ecosystems are among the most inhospitable soil ecosystems on earth with simple food webs and nearly undetectable fluxes of carbon (C) and nitrogen (N). Due to the lack of vascular plants, soil organic matter concentrations are extremely low, and it is unclear how much of the contemporary soil C budget is actively cycling or a legacy of paleolake production and sedimentation. While recent work indicates multiple sources of organic matter for dry valley soils, the composition and kinetics of organic pools remain poorly characterized. We examined soil organic matter pools and potential C and N turnover in soils from within six sites located across three hydrological basins of Taylor Valley, Antarctica that differed in surface age, microclimate and proximity to legacy (paleolake) sources of organic matter. We estimated potential C and N mineralization, and rate kinetics using gas exchange and repeated leaching techniques during 90-d incubations of surface soils collected from valley basin and valley slope positions in three basins of Taylor Valley. Soil organic C content was negatively correlated with the ages of underlying tills, supporting previous descriptions of legacy organic matter. Carbon and N mineralization generally followed 1st order kinetics and were well described by exponential models. Labile pools of C (90 d) were 10% of the total organic C in the upper 5 cm of the soil profile. Labile N was 50% of the total N in surface soils of Taylor Valley. These results show that a large proportion of soil C and particularly N are mineralizable under suitable conditions and suggest that a kinetically defined labile pool of organic matter is potentially active in the field during brief intervals of favorable microclimate. Climate variation changing the duration of these conditions may have potentially large effects on the small pools of C and N in these soils.

Introduction

The McMurdo Dry Valleys of Southern Victoria Land, Antarctica, are an arid, polar desert with some of the most inhospitable terrestrial environments on earth. Soil communities are dominated by microbes with few taxa of invertebrates and no vascular plants (Friedmann et al., 1993; Virginia and Wall, 1999). Soil organic carbon (C) concentrations are very low, 0.01 to 0.5 mg C g soil^{-1}, and a large proportion of the soil organic matter budget is thought to be a legacy of paleolake sediments (Moorhead et al., 1999; Burkins et al., 2000). While low rates of primary production by micro-algae have been reported for endolithic communities (Johnston and Vestal, 1991; Friedmann et al., 1993), a large proportion of contemporary organic matter in some valleys is derived from lake sediments deposited 10 to 20 ka ago when proglacial lakes occupied larger areas of the dry valley landscape (Stuvier and Denton, 1977; Denton et al., 1989; Burkins et al., 2000; Hall and Denton, 2000; Hall et al., 2000, 2001, 2002; Lyons et al., 2000).

Burkins et al. (2000) reported δ^{13}C and ^{15}N values for Taylor Valley soils intermediate between organic matter that resembles lacustrine (lake algal mats) and terrestrial sources (endolithic communities of micro-algae), suggesting multiple sources of organic matter, presumably of different ages and turnover characteristics. The degree to which lacustrine vs. terrestrial sources contribute to contemporary soil organic matter varies over elevation gradients due to the proximity of contemporary and legacy source pools, i.e., the Ross Sea, lakes and ephemeral glacial melt wetlands (Burkins et al., 2000; Moorhead et al., 2003). Similarly, biomass and diversity of invertebrate communities vary within and among valleys due to gradients of soil moisture, organic matter and salinity (Freckman and Virginia, 1997; Virginia and Wall, 1999; Courtright et al., 2001). It remains unclear how much of the contemporary organic matter is actively cycling through soil foodwebs, but *in situ* CO$_2$ flux measurements reveal surprisingly rapid C turnover (Parsons et al., 2004). For example, Burkins et al. (2001) reported a mean residence times for soil organic C of only 23 yr, which is low, considering that much of the contemporary soil organic matter is thought to be a legacy of lacustrine sediments deposited during the Last Glacial Maximum. Decadal mean residence times for soil organic matter indicate that C and N may turnover more rapidly than previous descriptions indicated (Matsumoto et al., 1990), and that soil foodwebs may be actively cycling recently fixed C in addition to legacy sources of organic matter. Dry valley ecosystems are very sensitive to small changes in temperature (Doran et al., 2002a). Therefore understanding the turnover kinetics of C and N is necessary to developing hypothesis about the influence of climate variability on soil organic matter dynamics.

In this paper we present estimates of labile C and N pools for soils collected from high and low elevations in Taylor Valley among sites previously characterized by isotopic composition suggestion mixed sources of organic matter, including lacustrine and endolithic sources (Burkins et al., 2000). Labile organic matter pools are kinetically defined using 90-d laboratory incubations (Nadelhoffer, 1990; Kaye et al., 2002). Since proximity to potential source pools may influence the amount and composition of SOM and N (Burkins et al., 2000, Fritsen et al., 2000), we selected sites close to contemporary lakes or ponds and higher elevation sites above the level of previous inundation by Lake Washburn (~300 m a.s.l., Fig. 1). Our objectives were to characterize potentially active pools of soil C and N over a gradient of

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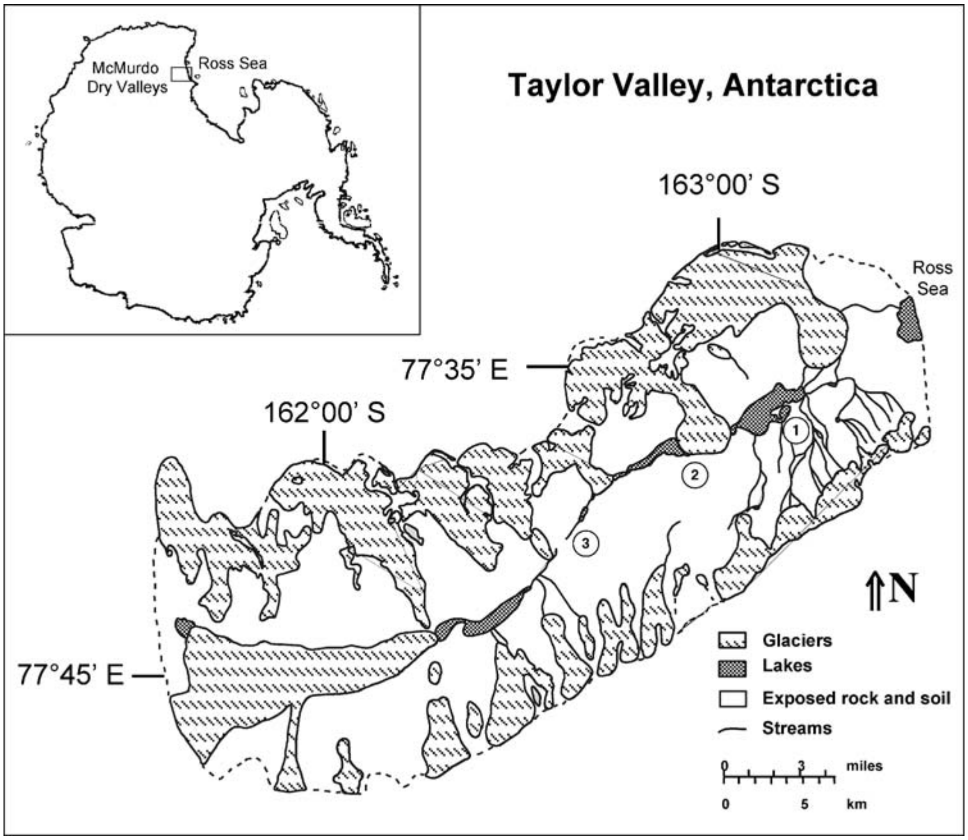

FIGURE 1. The McMurdo Dry Valley region in Victoria Land, Antarctica (inset), and Taylor Valley, a National Science Foundation Long-Term Ecological Research Site in Southern Victoria Land, Antarctica, and locations where soils were sampled for this study near (1) Lake Fryxell, (2) Lake Hoare, and (3) Mummy Pond.

soil age and proximity to legacy sources of organic material, and to evaluate the relationships among organic matter content and potential turnover with physical and chemical properties of Taylor Valley soils.

Materials and Methods

SITE DESCRIPTION

The McMurdo Dry Valleys are a series of ice-free valleys in the Trans-Antarctic Mountain Range, adjacent to the McMurdo Sound in southern Victoria Land, Antarctica. The USA National Science Foundation Long-Term Ecological Research (LTER) network maintains a research program in Taylor Valley with research focusing on the linkages among the physical and biological components of an ecosystem close to the tolerance limits for life. The climate in this region is extremely cold and dry, with brief periods (10–75 d) during the Austral summer when temperatures exceed 0°C. Mean annual temperature ranges from –14.8 to –30.0°C, and annual precipitation is less than 100 mm water equivalent (Doran et al., 2002b).

TABLE 1

Location and till chronology of study sites in the Taylor Valley.

Site	Latitude S	Longitude E	Elevation (m)	Till sequence	Surface age (ka)
Fryxell floor	77°37′	163°15′	67	Ross Sea[a]	10–24
Fryxell terrace	77°38′	163°17′	300	Ross Sea[a]	10–24
Hoare floor	77°38′	162°57′	80	Ross Sea[a]	10–24
Hoare terrace	77°39′	162°55′	335	Ross Sea[a]	10–24
Mummy floor	77°39′	162°41′	158	Taylor II-Bonney[b]	75–98
Mummy terrace	77°40′	162°42′	293	Taylor III[b]	200–210

[a] Hall et al. (2000).

[b] Higgins et al. (2000), Burkins et al. (2000).

Soils of Taylor Valley typically occur on poorly developed mid-Holocene to Pliocene tills, composed of diorites, granites, and sandstone (Campbell and Claridge, 1987; Bockheim, 1997). Soils in this region are poorly weathered, typically with greater than 95% sand by weight, high pH and low organic matter content (Campbell and Claridge, 1987). Dry valley soils are variously classified as Cryosols, Cryopsammets, and recently in the Gelisol soil order (Pastor and Bockheim, 1980; Campbell and Claridge, 1987; Bockheim, 1997; Beyer et al., 1999), though no published soil maps are available for this region.

We selected six sites spanning a range of climate and edaphic characteristics corresponding to low (valley floor) and high elevation (above Lake Washburn's influence) sites near Lake Fryxell, Lake Hoare, and Mummy Pond in Taylor Valley during December 1999 (Table 1, Fig. 1). These sites were selected and stratified by elevation on the basis of previous reports of δ^{13}C and δ^{15}N signatures consistent with multiple sources of organic matter, principally lacustrine and endolithic materials (Burkins et al., 2000). Samples were collected from high- and low-elevation sites, with one low-elevation site close to the valley floor, and one higher elevation site located on visible terraces near 300 m a.s.l. on north-facing slopes in each of the three major hydrological basins of Taylor Valley (Lyons et al., 1998; Fig. 1, Table 1).

Taylor Valley soils are largely derived from Pleistocene tills and represent a series in age and underlying substrate (Denton et al., 1989; Bockheim, 2002; Hall et al., 2000; Hall and Denton, 2000; Higgins et al., 2000). Inundation by Glacial Lake Washburn 10–20 ka ago resulted in accumulation of lacustrine sediments over much of Taylor Valley floor up to elevations of 300 m (Denton et al., 1989; Hall et al., 2000). Soils from the Fryxell and Hoare basins are underlain by the Ross Sea Drift and are estimated to be 10–24 ka (Bockheim, 1997; Hall et al., 2000; Hall and Denton, 2000). Soils from the Mummy Pond vicinity are underlain by glacial tills originating from the Taylor II-Bonney (75–95 ka) and Taylor III (200–210 ka) tills (Hall and Denton, 2000; Higgins et al., 2000; Bockheim, 2002).

SOIL COLLECTION AND ANALYSIS

We collected five replicate soil samples from the top 5 cm of the soil profile from coordinates in a 10 × 10 m grid at each site. Grids were located on level to gently sloping areas, generally representative of the soil surfaces in the near vicinity. Soils were collected into WhilPak bags and stored in ice chests for transport to the Crary Laboratory, McMurdo, where they were stored at 4°C, and processed within 48 h of collection. Soils were passed through a No. 10 sieve to remove rock and soil fragments >2 mm, and subsampled for gravimetric soil moisture, pH, electrical conductivity, and inorganic N content. Soil moisture was determined based upon weight loss from a 50-g field-moist sample dried in a forced-air convection oven at 105°C over 48 h to characterize soil moisture content at the time of sampling and to calculate the oven dry weight equivalent of field moist samples. We estimated soil salinity by measuring the electrical conductivity of a 1:5 soil solution of sieved soil in de-ionized water using a Corning 311 conductivity meter temperature calibrated with a 0.01 M KCl solution. Soil pH was measured on a 1:2 soil:de-ionized water slurry with an Orion model 720A pH meter. We measured initial KCl-extractable NO_3^- and NH_4^+ concentrations on 25-g subsamples shaken for 30 min in 50 ml of 2.0 M KCl, an orbital shaker, filtered through Whatman No. 40 paper and analyzed on a Lachat Autoanalyzer (Keeney and Nelson, 1982). We determined soil texture on a 30-g subsample by the hydrometer method (Gee and Bauder, 1986). A subsample of each soil was air-dried, ground in a ball mill, and 75 mg dry soil were analyzed on a Carlo Erba 1500 elemental analyzer to determine total C and N content (Nelson and Sommers, 1982). An additional 75 mg subsample was acidified with 50% HCl to remove carbonates and analyzed as above to determine organic C content.

INCUBATIONS

Carbon and N mineralization, and 90-d labile C and N pools were measured using a modified microlysimeter incubation technique developed by Stanford and Smith (1972) and described more recently by Nadelhoffer (1990). We incubated 75 g soil sealed in a 150-ml Falcon Filtration unit (cat no. 35-7102, Franklin Lakes, NJ) and repeatedly measured CO_2 production (weekly) and inorganic N (NH_4^+ and NO_3^-) accumulation (biweekly) over the course of 90-d incubations. Incubation vessels were individually tested for leaks prior to the incubation by submerging filter units in water and applying pressure with a rotary vacuum pump. To measure C mineralization, filtration units were sealed for 48 h prior to weekly sampling and the amount of CO_2 accumulation in the headspace of the unit was measured using a Li-Cor 6200 Portable Infrared Gas Analyzer modified for direct injection of small (10 ml) gas volumes (Anonymous, 1999). Carbon mineralization was calculated from the change in headspace concentration of CO_2 relative to a blank, and converted to gravimetric units using the ideal gas law (Robertson et al., 1999) and the following equation:

$$C_m = \frac{(\mathrm{CO_2}\cdot 12\cdot P)}{R\cdot T}\cdot\left(\frac{V}{M}\right), \tag{1}$$

where C_m is C mineralization rate expressed as μg C·g $\mathrm{soil}^{-1}\cdot\mathrm{d}^{-1}$, CO_2 is the change in headspace CO_2 concentration (μL CO_2 L $\mathrm{headspace}^{-1}$ d^{-1}), 12 is the molecular weight of C per μmol CO_2, P is the barometric pressure in atmospheres, R is the universal gas constant (0.0820575 L atm·$[°\mathrm{K}\cdot\mathrm{mole}]^{-1}$), T is the incubation temperature in °K, V is the headspace volume of the soil microcosm in L (total flask volume − soil mass·bulk $\mathrm{density}^{-1}$), and M is the mass of incubated soil in g.

We measured the accumulation of inorganic N by leaching soil microlysimeters with 100 ml of 2 M KCl under low tension and analyzing the leachate for inorganic N as described above. Following leaching, soils microlysimeters were flushed with 100 ml of deionized water and a minus-N trace nutrient solution to recharge lost nutrients other than inorganic N. Incubations were initiated at 10% gravimetric soil moisture and 20°C, but subsequent leaching and recharging brought soils up to field capacity. Estimates of N accumulation were blank corrected, expressed on a gravimetric basis and plotted vs. d of incubation.

TABLE 2

Soil physical and chemistry properties ± 1 standard deviation for high and low elevation sites in Taylor Valley.

Site	Depth of thaw[a] (cm)	Gravimetric moisture content (%)	Fines[b] (%)	pH	Conductivity (μS cm^{-1})
Fryxell floor	30.5 ± 1.0	2.1 ± 0.2	3.0 ± 0.6	9.75 ± 0.25	260 ± 120
Fryxell terrace	42.5 ± 1.0	1.7 ± 0.2	2.5 ± 1.1	9.29 ± 0.32	49 ± 12
Hoare floor	52.1 ± 0.5	2.5 ± 0.7	1.6 ± 1.2	10.04 ± 0.19	430 ± 270
Hoare terrace	27.3 ± 0.5	3.7 ± 0.5	0.7 ± 0.4	9.10 ± 0.09	49 ± 16
Mummy floor	27.6 ± 0.5	1.2 ± 0.2	1.6 ± 1.3	9.40 ± 0.62	58 ± 13
Mummy terrace	24.4 ± 0.5	0.8 ± 0.2	1.7 ± 1.6	9.71 ± 0.13	77 ± 25

[a] Depth to ice cemented ground.

[b] Clay sized particles by weight (%).

An exponential model was fit to the C and N mineralization rates over the 90 d, and parameters and rate constants estimated from nonlinear regression using the following equation:

$$C_m = kC_a e^{(-kt)}, \tag{2}$$

where C_m is the mineralization rate at d t (μg C g soil^{-1} d^{-1}), k is the turnover coefficient (d^{-1}), C_a is the 90-d labile pool (μg C g soil^{-1}), and e is the base of the natural logarithm, 2.718. A resistant fraction of C was estimated as the residual fraction (i.e., un-mineralized pool) or

$$C_r = SOC_t - C_a, \tag{3}$$

where C_r is the residual fraction, SOC_t is the total soil organic C as determined by elemental analysis (see above), and C_a is the 90-d labile pool as determined from the exponential fit to C mineralization over the 90-d incubations. Rate constants and parameters were fit to rates of net N mineralization from the microlysimeters using the same analytical approach.

STATISTICAL ANALYSES

We tested for differences in soil properties, organic matter, and rates of C and N mineralization over the sites sampled using Pearson Correlation and a general linear model in the SAS v8.0 software package (SAS Institute, Cary, NC). We used nonlinear regression to characterize the temporal trends in C and N mineralization over the 90-d incubations using nonlinear regression procedure in SAS v8.0 (SAS Institute, Cary, NC). A pseudo-R^2 was calculated as 1 − (sum of $\mathrm{squares}_{\mathrm{residual}}$/sum of $\mathrm{squares}_{\mathrm{corrected\ total}}$) for fits to the nonlinear models.

Results

SOIL PROPERTIES

Soil chemical and physical properties varied among the sites studied; however there were no consistent differences between soils collected from low and high elevations within individual basins (Table 2). Soils were generally dry (1–4% gravimetric moisture content), coarse textured (<3% clay sized particles by weight), and alkaline (9.1–10.0 pH). Salinity varied from low electrical conductivities in Mummy soils (58–77 μS cm^{-1} in 1:5 soil dilutions) to ranges of 50 to

TABLE 3

*Soil inorganic and organic carbon and nitrogen content. Means with the same letter are not significantly different (*P < 0.05*).*

Site	NO_3^- (µg N·g soil^{-1})	NH_4^+ (µg N·g soil^{-1})	Total N (mg N g soil^{-1})	Inorganic C (mg C g soil^{-1})	Organic C (mg C·g soil^{-1})
Fryxell floor	0.69 ± 0.95	1.47 ± 0.93^a	0.037 ± 0.01^a	0.65 ± 0.13	0.31 ± 0.02
Fryxell terrace	0.08 ± 0.11	1.30 ± 0.69^a	0.043 ± 0.01^a	2.02 ± 0.16	0.37 ± 0.01
Hoare floor	0.02 ± 0.02	0.04 ± 0.04^c	0.020 ± 0.01^b	1.29 ± 0.05	0.29 ± 0.03
Hoare terrace	0.15 ± 0.25	0.02 ± 0.02^c	0.017 ± 0.00^b	0.81 ± 0.15	0.31 ± 0.04
Mummy floor	N.D.[†]	0.16 ± 0.13^b	0.024 ± 0.00^b	0.28 ± 0.12	0.24 ± 0.02
Mummy terrace	0.02 ± 0.01	0.20 ± 0.13^b	0.022 ± 0.01^b	0.46 ± 0.04	0.23 ± 0.02

[†] None detected. Concentrations were below detection limits.

430 µS cm^{-1} in soils collected from Fryxell and Hoare ($F = 5.57$, $P = 0.0103$). Similarly, soil carbonate content varied from low concentrations of 0.28 to 0.46 mg C·g soil^{-1} in Mummy soils to higher concentrations of 0.65 to 2.00 mg C·g soil^{-1} in Hoare and Fryxell soils ($F = 36.79$, $P < 0.0001$, Table 3).

Soil organic C and N content was greatest in Fryxell soils ($F = 10.47$, $P = 0.0005$, $F = 19.51$, $P = 0.0001$, respectively; Fig. 2 and Table 3). Initial inorganic N concentrations varied from below detection limits for NO_3^- in soils from near Mummy Pond, to 1.5 µg N-NH_4^+·g soil^{-1} in Fryxell soils (Table 3). Soil NO_3^- levels were highly variable, with no significant differences apparent among the sites ($F = 1.98$, $P = 0.16$). Ammonium concentrations varied over several orders of magnitude, and were greatest in Fryxell soils, followed by Mummy, and Hoare soils ($F = 19.49$, $P < 0.0001$). No consistent differences in C and N content, for organic or inorganic fractions were evident between soils collected from low and high elevations with individual basins (Table 3).

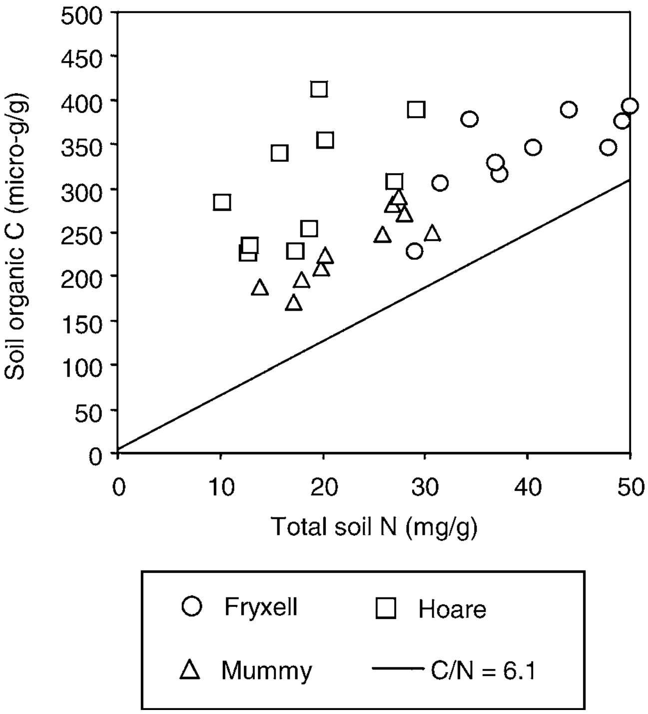

FIGURE 2. Soil organic C vs. total soil N concentration in soils collected near valley floors and from higher elevation (~300 m) on north-facing slopes in Taylor Valley, Antarctica. The solid line is the general ratio of C:N for balanced growth of heterotrophic bacteria (Goldman and Dennett, 2000).

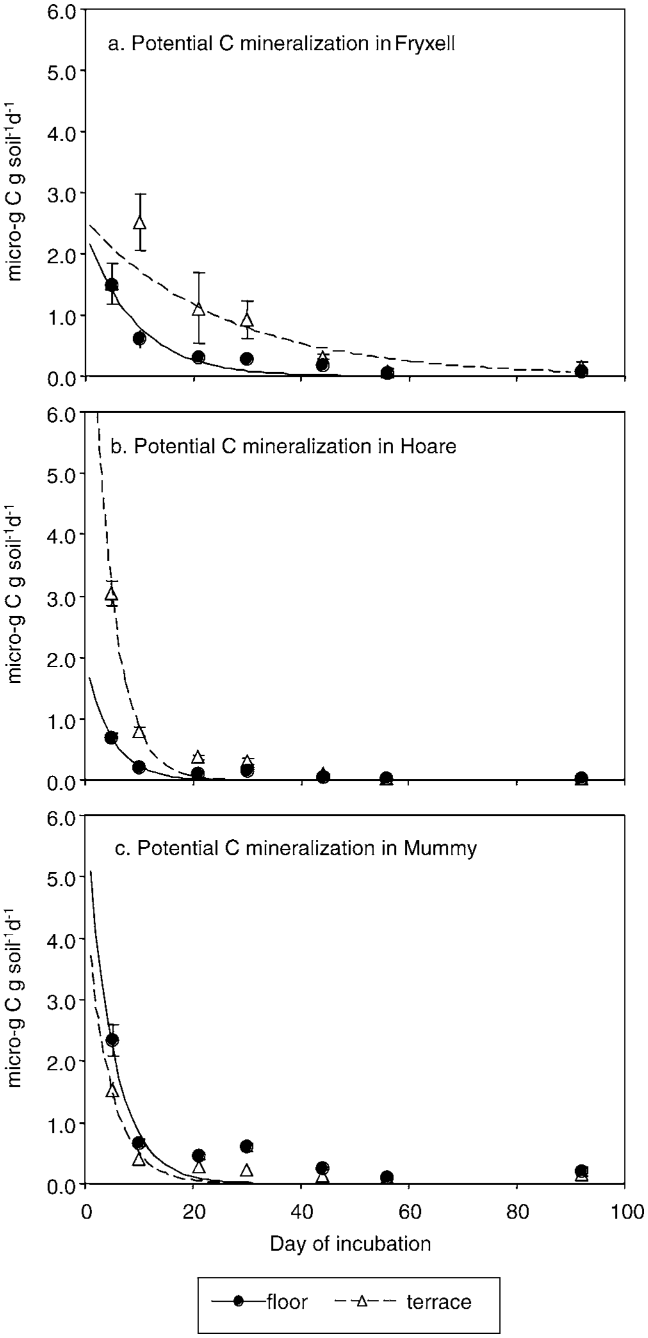

FIGURE 3. Potential C mineralization (µg C·g soil^{-1}·d^{-1}) in 90-d incubations of soils collected from low (valley floor) and high elevations (> 300 m a.s.l.) sites near (a) Lake Fryxell, (b) Lake Hoare and (c) Mummy Pond in Taylor Valley. Rate parameters and constants for 1st order models of potential C mineralization are given in Table 4. Error bars are 1 standard error of the mean.

CARBON AND NITROGEN TURNOVER

Potential C mineralization followed 1st order kinetics during the 90-d incubations (Fig. 3). Nonlinear regressions were significant for all incubations, and generally accounted for a large proportions of variance

TABLE 4

Regression parameters and rate constants ± 1 standard error of the mean for the single exponential model describing soil respiration (Fig. 3) with the following equation: C mineralization = k $C_a e^{(-kt)}$. Parameters with the same letter are not significantly different (P < 0.05).

Site	Coefficient (C_a) 90-d labile C µg C·g soil^{-1}	Proportion of SOC mineralized	k (d^{-1})	Pseudo-R^2	P
Fryxell floor	21.95^{b†} ± 2.85	0.07 ± 0.003	0.11^{b} ± 0.03	0.77	0.0001
Fryxell terrace	65.43^{a} ± 23.97	0.18 ± 0.02	0.039^{c} ± 0.02	0.51	0.0012
Hoarse floor	9.44^{bc} ± 0.78	0.03 ± 0.001	0.22^{a} ± 0.053	0.88	0.0001
Hoarse terrace	42.38^{a} ± 2.83	0.14 ± 0.003	0.25^{a} ± 0.036	0.95	0.0001
Mummy floor	31.10^{ab} ± 2.96	0.13 ± 0.003	0.20^{a} ± 0.075	0.79	0.0001
Mummy terrace	21.03^{b} ± 2.25	0.10 ± 0.003	0.22^{a} ± 0.076	0.77	0.0001

[†] Regression parameters with different letters have nonoverlapping 95% confidence intervals.

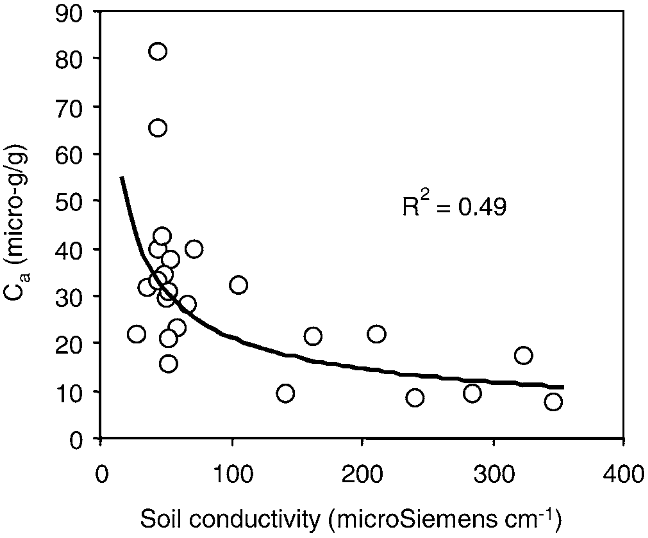

FIGURE 4. Plot of potentially mineralizable C (C_a) during 90 incubations of soils collected from Taylor, Valley vs. soil conductivity (1:5 dilution). The logarithmic fit (y = –12.30Ln(x) + 82.03) to conductivity accounts for 49% of the variation in labile C (P = 0.014).

(i.e., >75%) in the observed rates of C mineralization (Table 4). Soils collected from the Lake Fryxell basin had the highest concentrations of 90-d labile C among the sites studied (but also the poorest fit to the nonlinear model, Table 4), followed by Mummy and Hoare soils. Carbon mineralization from incubations of Hoare soils produced the best fits to the nonlinear models (pseudo-R^2s = 0.88–0.95). Ninety-day labile pools of C (C_a) ranged from 10 to 65 µg C·g soil^{-1}, which represented 3 to 18 % of the total soil organic C. Ninety-day labile pools of C (C_a) were negatively correlated with the conductivity of 1:5 soil extracts. A log function of soil conductivity accounted for 49% of the variability in the 90-d labile C fraction ($P = 0.014$, Fig. 4). Similarly, there was a negative correlation between C_a and soil pH (Pearson correlation coefficient = –0.68, $P < 0.002$).

The mass of inorganic N leached from the microlysimeters decreased over 90 d (Fig. 5). This trend was mainly due to large pulses of NH_4^+ leached during the first weeks followed by a rapid, 1st order decline in NH_4^+ leaching over the remaining course of the incubations. Rates of NO_3^- leaching were universally low, and did not follow 1st order kinetics. Nitrification was evidently not limited by the availability of inorganic N in these incubations since there was no indication of a decline in the production of NO_3^-, particularly for incubations of Fryxell and Hoare soils. Sample contamination is a present concern among N applications for Antarctic soils, given the extremely broad range in NO_3^- concentration (Barrett et al., 2002). However, we discount contamination as a contributing factor in the apparent pool-independent dynamics of NO_3^- leaching since concentrations of NO_3^- in the leachate were consistently more than 3× the concentration of NO_3^- in incubation and extractant blanks.

Net N mineralization was very well described by 1st order kinetic models for all soils (pseudo $R^2 = 0.94$–0.99), with the exception of those collected from the high-elevation Fryxell site, where 90-d pools (N_a) were also the lowest (Fig. 6, Table 5). Rates of potential nitrification were considerable lower than net N mineralization and net NH_4^+ accumulation in all soils (data not shown). Ammonium production accounted for 90 to 100 % of the N flux in these incubations of Taylor Valley soils. The 90-d labile fractions of N (N_a) ranged from 8 to 25 µg N·g soil^{-1}, which represented 26 to 66 % of the total soil N pools. Correlations between N mineralization rate parameters and soil physiochemical properties were universally weak (Pearson correlation coefficients <0.50).

The C:N ratios of 90-d labile pools (C_a/N_a) were low, generally below published C:N ratios for balance growth of heterotrophic bacteria, i.e., the Redfield ratio, 6.125 by weight (Goldman and Dennett, 2000; Fig. 7a). In contrast, the C:N ratios of the residual fraction were similar to the C:N of bulk soil organic matter, and universally above the Redfield Ratio (Fig. 7b). Total soil organic C and residual organic C were both negatively correlated with published estimates of surface till ages (Fig. 8). Exponential fits to these corresponding surface till ages accounted for 69 and 75% of the variance in total and residual soil organic C, with turnover coefficients of 0.0021 and 0.0020 yr^{-1}, respectively ($P = 0.03$, $P = 0.005$). These turnover coefficients indicate mean residence times of 200 to 476 yr for total and residual C.

Discussion

Potential fluxes of C and N were up to an order of magnitude greater than estimates of *in situ* turnover (Burkins et al., 2001; Parsons et al., 2004), consistent with a large body of incubation data showing a stimulating effect of disturbing soil structure and incubating under continuous favorable conditions (e.g., Herlihy, 1979; Raison, 1987; Hook and Burke, 1995). While not indicative of rates in the field, potential estimates of C and N turnover can provide important information about sizes and kinetics of active soil organic matter pools. For example, large proportions of readily mineralizable C and N suggest that dry valley organic matter is potentially more active than previously thought, based upon assumptions made from descriptions of soil organic matter sources (Friedmann et al., 1993; Moorhead et al., 1999; Burkins et al., 2000). Indeed, *in situ* estimates of CO_2 efflux suggest a more rapid turnover of C than is possible if legacy organic matter alone were the source for contemporary foodwebs, given the decadal ranges of mean residence time estimates, and the ancient provenance of legacy sources (Burkins et al., 2001; Parsons et al., 2004).

Estimates of labile C (~10% of total soil organic C) compare well with published data on microbial biomass in dry valley soils, suggesting a large role for soil biota in labile organic matter fractions. Vestal (1988) reported a viable microbial biomass 0.3 to 9.6% of total organic carbon in soils from Linnaeus Terrace in the Upper Wright Valley (elevation 1600 m). While this range is slightly lower than the values for labile C fractions reported in Table 4, it is likely that in addition to living microbial biomass, the labile fraction includes cellular exudates and dead microbial material. Such substances may represent the "fresh litter" inputs in this system, given the absence of higher plants. The high proportion of labile C in dry valley soils is likely due to the algal/

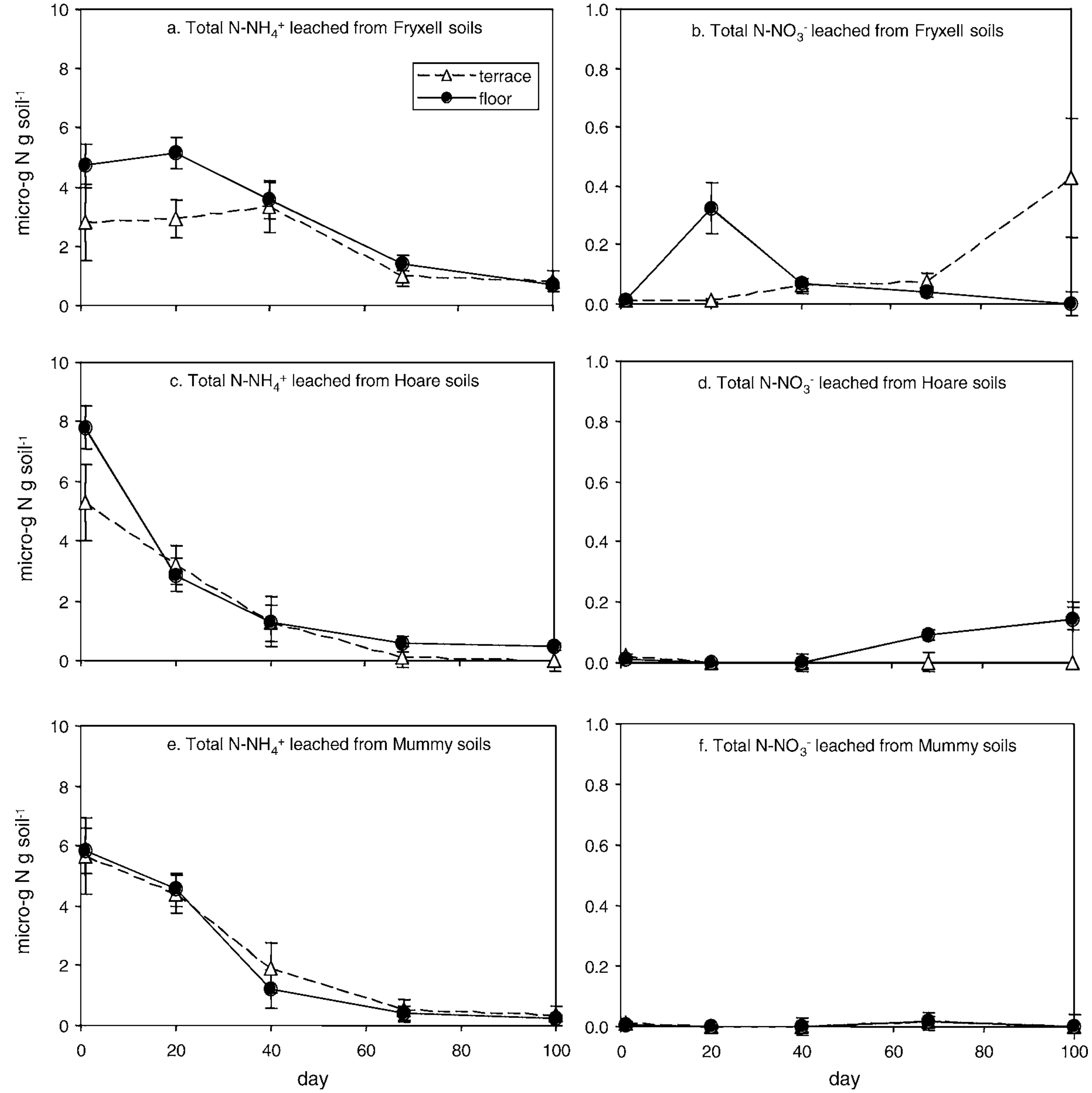

FIGURE 5. Mean ammonium and nitrate content (μg N·g soil^{-1}) of leachates from microlysimeter incubations of Taylor Valley soils. Microlysimeters were leached biweekly with 2 M KCl. Error bars are 1 standard error of the mean.

microbial source of organic matter. While previous studies have described multiple sources of this organic matter for dry valley soils (i.e., marine, lacustrine, and endolithic), all sources consist of algal and microbial biomass (with very minor contributions from moss and lichens in certain locally confined environments). Algal and microbial tissue contains no structural materials such as lignin or cellulose that restrain decomposition rates; hence the fast potential turnover of organic C in dry valley soils is more indicative of litter decomposition than mineralization of soil organic matter (Parton et al., 1987; Agren et al., 1991).

In contrast to most temperate soil ecosystems, decomposition of organic matter in terrestrial dry valley ecosystems is not limited by substrate quality or nutrient deficiencies. Under such conditions, microbes and soil fauna would be poised to respond to pulses of liquid water availability during the brief austral summer, and could support relatively high, though transient rates of CO_2 flux. Previous work in the dry valleys has shown that biotic production in both aquatic and terrestrial ecosystems are sensitive to even small changes in temperature (Doran et al., 2002a). Considering that dry valley organic matter is composed of a large proportion of potentially active C, compared to temperate systems where active pools seldom account for more than 2–5% of the total organic C (Zak et al., 1994; Paul and Clark, 1996; Paul et al., 1999, Kaye et al., 2002), it is possible that small changes in summer temperature could change soil organic matter dynamics and C storage in the dry valleys.

Similarly, a large proportion (50%) of the soil N pool was mineralized from Taylor Valley soils. Though concentrations of inorganic N in Taylor Valley are low in comparison to dry temperate systems, potential turnover during laboratory incubations was much higher (Barrett and Burke, 2000; Kaye et al., 2002), suggesting that the potential for N mineralization in the field is significant under favorable microclimates. Moreover, the low C:N ratio of 90-d mineralizable fractions (Fig. 7a) indicates that N is available in excess of microbial metabolic and growth requirements, i.e., the balance of C and N mineralization did not conform to the ratios predicted by balanced microbial growth. Such profligate N turnover during incubations suggests that N is not limiting to soil microbes in Taylor Valley.

Nitrogen cycling in the dry valley soils has received relatively little attention. Most accounts have remarked on the broad ranges of

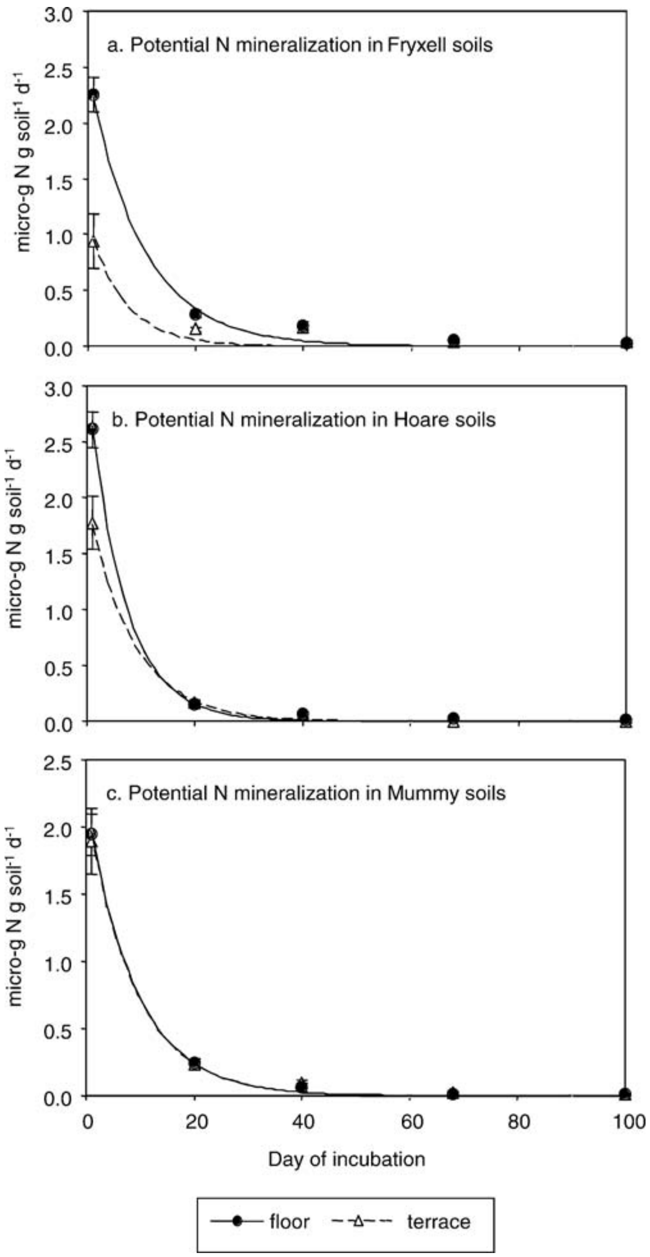

FIGURE 6. *Potential net nitrogen mineralization (μg N·g soil^{-1}·d^{-1}) over 90-d microlysimeter incubations of soils collected from low elevation (valley floors) and higher elevation (~300 m) on north-facing slopes in Taylor Valley, Antarctica. Microlysimeters were leached biweekly with 100 ml of 2 M KCl to remove inorganic N. Nitrogen mineralization was calculated as the rate of ammonium and nitrate accumulation per day over the course of the incubations. Rate parameters and constants for 1st order models of potential net N mineralization are given in Table 4.*

inorganic N and high concentrations of soluble NO_3^- (Wada et al., 1981; Campbell and Claridge, 1987; Bockheim, 1997). The high NO_3^- concentrations (>1000 ppm) in ancient, high-elevation soils are of particular note (Barrett and Virginia, unpublished). In contrast, in the soils described here, NO_3^- and NH_4^+ were both important constituents of the inorganic pool, though fluxes of N were dominated by release of NH_4^+, and nitrification was quite low. These low rates of potential nitrification in the laboratory are consistent with previous studies that suggest abiotic processes (atmospheric deposition) are largely responsible for accumulation and cycling of NO_3^- in the dry valleys (Bockheim, 1997; Lyons et al., 2003).

TABLE 5

Regression parameters and rate constants ± 1 standard error of the mean for the single exponential model describing soil mineralization (Fig. 5) with the following equation: N mineralization = $k\ N_a e^{(-kt)}$. Parameters with the same letter are not significantly different ($P < 0.05$).

Site	Coefficient (C_a) 90-d labile C μg C·g soil^{-1}	Proportion of SOC mineralized	k (d^{-1})	Pseudo-R^2	P
Fryxell floor	24.55^{a†} ± 2.68	0.40 ± 0.08	0.10^{b} ± 0.01	0.99	0.0001
Fryxell terrace	7.35^{c} ± 1.49	0.26 ± 0.06	0.15^{a} ± 0.01	0.51	0.0002
Hoarse floor	20.46^{ab} ± 2.60	0.66 ± 0.10	0.15^{a} ± 0.02	0.99	0.0001
Hoarse terrace	16.55^{b} ± 2.64	0.50 ± 0.10	0.12^{ab} ± 0.02	0.97	0.0001
Mummy floor	19.99^{ab} ± 3.99	0.48 ± 0.11	0.11^{ab} ± 0.02	0.94	0.0001
Mummy terrace	19.71^{ab} ± 1.46	0.51 ± 0.05	0.11^{ab} ± 0.01	0.99	0.0001

† Regression parameters with different letters have nonoverlapping 95% confidence intervals.

While the Taylor Valley soils described here do not have unusually large concentrations of NO_3^- or inorganic N in general, the rapid turnover and low C:N of mineralizable pools indicate that available N probably exceeds biotic demand. The presumably abiotic source (i.e., deposition) of much of the NO_3^- found in dry valley soils, and the low C:N ratio of mineralizable pools suggests that sources of C and N, and controls over their respective turnover are very different in the dry valleys relative to temperate ecosystems. Ongoing work is addressing sources and potential loss vectors of soil N in the dry valleys.

IMPLICATIONS FOR IN SITU *C AND N FLUXES*

Variation in soil organic matter dynamics among the study sites can be attributed to differences in local microclimate and edaphic characteristics. Physical soil properties are important influences over the distribution of soil biodiversity (Kennedy, 1993; Freckman and Virginia, 1997; Virginia and Wall, 1999; Courtright et al., 2001), biotic activity and survival (Nkem and Wall, unpublished), soil respiration (Parsons et al., 2004), and C turnover dynamics. For example, negative correlations between 90-d labile C and soil conductivity and pH suggest that biotic activity may be inhibited by high pH and soluble salt concentrations, even under the favorable conditions experienced during laboratory incubations (Fig. 4). Further, the poor fit of 1st order kinetic models to 90 d C mineralization in some soils may be due to physiochemical constraints upon soil biota, rather than C-substrate limitations (Fig. 3a).

In addition to ranges of physiochemical properties, the sites used in this study represent a gradient of till age and proximity to legacy sources of organic matter. Soils developing at lower elevations, and/or on younger tills are closer to lacustrine and marine sources of organic matter and it follows that soil organic C content has generally been found to decrease with increasing elevation within Taylor Valley (Powers et al., 1998; Burkins et al., 2000; Fritsen et al., 2000). In this study, there were no significant differences in bulk soil properties or organic matter content between soils collected from higher elevations near 300 m, and those collected from valley floors at lower elevations (Table 3). However, the relationship between soil organic C and till ages (Fig. 8) suggests that there are differences in soil C fractions, and perhaps C turnover, among sites whose soils range from the those underlain by young tills of the Ross Sea Drift to Mummy soils underlain by the older Taylor II and III glacial drifts. While organic matter content did not vary consistently with elevation, the variation in

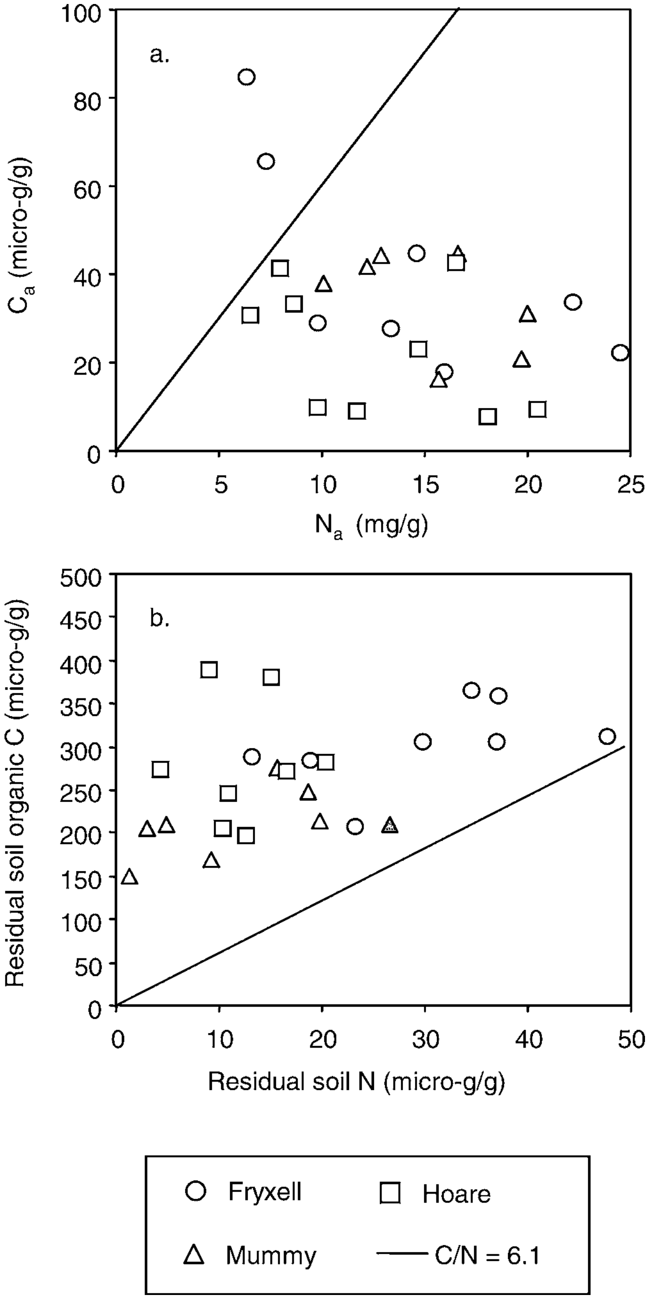

FIGURE. 7. Carbon:nitrogen ratios for (a) 90-d labile C and N pools, and (b) residual (unmineralized) pools as determined from 90 incubations. Solid line is the C:N for balanced growth of bacterial biomass (see text).

soil C fractions over a range of till ages coincides with previous studies that have used elevation as a proxy for age (e.g., Burkins et al., 2000). If soil organic pools are influenced by geomorphic processes (i.e., legacy sources), then the negative relationship between soil organic C fractions and surface age depicted by Figure 8 suggests that younger soils have had more recent inputs, and hence higher concentrations of C compared to older, perhaps more depleted soils developing on older tills. This relationship between soil C fractions and till age may be useful for inferring residence times (1/k) for soil organic C.

The turnover rates estimated by exponential fits between C content and till age predict mean residence times of 200 to 500 yr for soil organic C. This is much longer than the decadal (23–73) residence times based on observations of *in situ* CO_2 efflux from dry valley soils, but is more consistent with a multiple source model of soil organic matter derived in large part from legacy sources of organic matter deposited during the Last Glacial Maximum. Together, these rates of C turnover estimated from *in situ* and laboratory measurements of CO_2 flux, and inferred from variation in soil C fractions across different till ages, suggest that soil organic matter in Taylor Valley is composed of at least two kinetically distinct pools. A small pool with a short residence time may be capable of active cycling under favorable microclimates, while a larger, less kinetically active pool cycles over century to millennial time scales. These results suggest that active pools of organic matter may be sensitive to contemporary climate variability, while organic matter of ancient origin may be resistant to all but major episodes of climate change associated with variation in lake levels and landscape development.

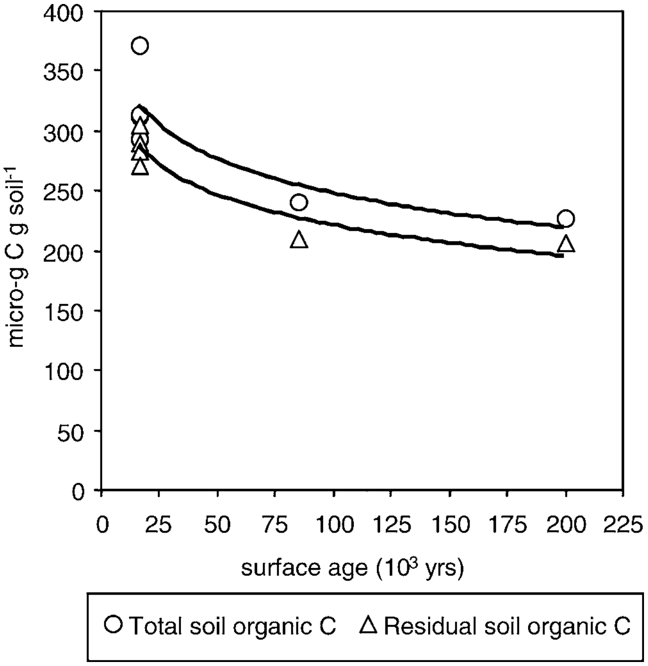

FIGURE 8. Plot of total soil organic C and residual organic C vs. till ages (Table 1). The following nonlinear fits: $y = 325e^{-0.002x}$, and $y = 290e^{-0.002x}$ accounted for 69 and 75% of the variance in the total, and residual organic C in dry valley soils incubated over 90 d.

Acknowledgments

We thank K. Catapano, S. Blecker, G. Steinhart, and K. Lavigne for assistance in the field and laboratory. Logistic support was provided by Antarctic Support Associates and Petroleum Helicopters Inc. This work was supported by the National Science Foundation McMurdo Long-Term Ecological Research (OPP 0096250).

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Revised ms submitted May 2004