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## The impact of land use change on C turnover in soils

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**Abstract.** Measurements of CO<sub>2</sub> flux, soil temperature, and moisture content of selected natural and disturbed soils in central California were made on a monthly basis from August 1994 to October 1995 in an attempt to detect the effects of temperature, moisture, and land use change on CO<sub>2</sub> production in soils. Soil CO<sub>2</sub> flux displayed a strong negative correlation with soil temperature and a positive correlation with soil moisture at the natural site. However, at the disturbed site, the linear correlation between CO<sub>2</sub> flux and temperature/moisture was insignificant. The negative correlation between soil CO<sub>2</sub> flux and soil temperature is in contrast to what has been observed in other ecosystems but is typical for Mediterranean ecosystems in which grasses are biologically active only during cool months. Comparison of carbon (C) inventories of paired natural and disturbed soils indicates that both cultivation and logging have resulted in a significant decrease in total soil C content. The reduction in soil C storage is about 26% for the cultivated soil and around 30% for the logged soil. Most of the C loss is from the upper horizons. Radiocarbon (<sup>14</sup>C) measurements of both recent and archived soil samples demonstrate large differences in C input rate and turnover time between natural and disturbed soils. The average turnover times of organic matter are longer in disturbed soils than in the corresponding natural soils as a result of preferential loss of C from “active” soil C pools. In both natural and disturbed soils, the average turnover times of organic matter increase with depth from decades or less in shallow horizons to hundreds of years or even thousands of years in deeper horizons. Our results show that land use change can have significant impact on soil C cycle and that shallow soil horizons are most susceptible to disturbance because of shorter turnover times of organic C in these horizons.

### 1. Introduction

Soils are the largest C pool in the terrestrial environment. The amount of C stored in soils is about twice the amount of C present in the atmosphere and is about 3 times the amount of C stored in the living plants [Schlesinger, 1990, 1995; Kimble and Stewart, 1995]. Consequently, any change in the size of soil C pool could alter the atmospheric CO<sub>2</sub> concentration and affect the global C balance [Schlesinger, 1991; Jenkinson *et al.*, 1991; Trumbore *et al.*, 1996]. Humans are constantly modifying the landscape and soils in many ways, yet the effects of these modifications on soil C dynamics and on the global C cycle are poorly known. Land use change impacts soil by altering the soil environment, which in turn affects microbial growth and decomposition processes that transform plant-derived C to soil organic matter and CO<sub>2</sub>. Because of a lack of fundamental knowledge of soil carbon dynamics, the magnitude and timing of the response of the soil carbon

reservoir to changes in land use are a large source of uncertainty in global carbon cycle models. The quantification of the changes in soil C storage and turnover time resulting from land use change is a prerequisite to understanding the C exchange rate between the soil and the atmosphere.

In this study, we monitored soil CO<sub>2</sub> flux as well as soil temperature and moisture variations in selected soils (both natural and disturbed) in central California from August 1994 to October 1995. We also analyzed C and <sup>14</sup>C content of soil organic matter in both recent and archived samples of these soils. Our objective is to establish the relationship between CO<sub>2</sub> production in soils and environmental parameters and to examine how land use change affects C storage and turnover in soils.

### 2. Methods

#### 2.1. Field and Laboratory Methods

Two pairs of natural and disturbed soils along the western slope of the Sierra Nevada Mountain range in central California were selected for this study. The first pair of soils are located at the foothill at ~ 470 m elevation. These soils have formed in Mesozoic granodiorite. The mean annual temperature and precipitation at this elevation are 17.8°C and 31 mm, respectively. The natural site (Fallbrook) is an oak

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annual grassland and the nearby disturbed site (which is referred to as Fallbrook-OCD) is a lemon orchard that was a grassland until 1972. The second pair of soils are in a forested area, at an elevation of 2414 m. Here the parent material for the soils is Tertiary basalt. The natural site (Windy) is in an old growth forest reserve, the "Teakettle Experimental Area," and the disturbed site (Windy-cut) is in a nearby logged area where open spaces due to logging have been invaded by dense shrubs. The precipitation at this elevation is  $\sim 102 \text{ mm yr}^{-1}$  and mean annual temperature is  $5.6^\circ\text{C}$ .

Soil samples were collected by horizon to the base of shallow pits dug for each representative soil for total carbon,  $^{13}\text{C}$ , and  $^{14}\text{C}$  analysis. Bulk density of soils at various depths were measured via a coring apparatus. In addition, "prebomb" soil samples from each site, obtained from an archive at University of California, Berkeley, were also analyzed for total carbon,  $^{13}\text{C}$ , and  $^{14}\text{C}$  content. At the low-elevation sites (Fallbrook and Fallbrook-OCD), soil temperature and moisture at 10, 25, 50, and 100 cm depths were monitored throughout a year using permanently installed sensors (manufactured by ELE International, Inc.), and  $\text{CO}_2$  fluxes at the soil surface were measured on monthly basis using a LI-COR  $\text{CO}_2$  analyzer [Norman et al., 1992; Ben-Asher et al., 1994; LI-COR, 1990].

Soil samples for organic carbon analysis were handpicked of visible plant and root debris.  $\text{CO}_2$  was produced by combustion of the sample with CuO and silver foil under vacuum at  $875^\circ\text{C}$  for 2 hours and purified cryogenically. Weight percentage of C in a sample was determined from the  $\text{CO}_2$  yield. Total soil C inventory was determined from the %C and bulk density measurements (corrected for field-estimated gravel contents). The purified  $\text{CO}_2$  was reduced to graphite with  $\text{H}_2$  over Fe, and its  $^{14}\text{C}/^{13}\text{C}$  ratio was measured on an accelerator mass spectrometer (AMS) at the Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory. Carbon 14 data are reported as  $\Delta^{14}\text{C} = [A_{\text{SN}}/A_{\text{ABS}} - 1] \times 1000\%$ , where  $A_{\text{SN}}$  is the specific activity of a sample (which is proportional to the  $^{14}\text{C}/\text{C}$  ratio in the sample) normalized to  $\delta^{14}\text{C} = -25\%$  using  $^{13}\text{C}/^{12}\text{C}$  ratios measured in a stable isotope ratio mass spectrometer and  $A_{\text{ABS}}$  is the absolute international standard activity which is equal to 0.95 times the specific activity of the international standard (NBS oxalic acid distributed by National Institute of Standards and Technology) normalized to  $\delta^{13}\text{C} = -19\%$  and the year 1950 [Stuiver and Polach, 1977]. Analytical precision ( $1\sigma$ ) for  $^{14}\text{C}$  analysis is better than 9%, and for C content it is 8%.

## 2.2. Using $^{14}\text{C}$ to Study the Turnover Time of Soil Organic Matter

The "bomb"  $^{14}\text{C}$  produced during the nuclear era is a powerful tracer to study organic matter turnover in soils [Harrison et al., 1993a, b; Hsieh, 1989, 1992, 1993, 1996; O'Brien and Stout, 1978; O'Brien, 1984; Trumbore et al., 1989, 1990, 1996; Trumbore, 1993; Wang et al., 1996a, 1997a]. Nuclear weapons testing during the period from 1954 to 1962 resulted in a massive input of  $^{14}\text{C}$  into the atmosphere. After the test ban agreement in 1963, this excess bomb  $^{14}\text{C}$  has been steadily decreasing toward the prebomb levels because of its uptake by the oceans [Broecker and Olson, 1960] and the terrestrial ecosystems [Trumbore et al., 1996] and also owing to dilution by " $^{14}\text{C}$ -dead"  $\text{CO}_2$  from fossil fuel burning.

Temporal variations of this "excess"  $^{14}\text{C}$  in the atmosphere are well documented [Manning et al., 1990; Levin et al., 1985], and these variations are accurately reflected in the  $^{14}\text{C}$  content of plant tissues [Suess, 1955; Stuiver, 1965; Wang et al., 1997b]. In order to translate the bomb  $^{14}\text{C}$  signatures into organic matter turnover times, it is necessary to model the pattern of the  $^{14}\text{C}$  uptake into the soil organic matter.

The most common approach is to model a soil or any of its horizons as a time-dependent box. Assuming that organic matter decomposition follows the first-order reaction and is the only mechanism for carbon loss in soils, the variation in organic C and  $^{14}\text{C}$  in a soil or any of its horizons with time can be described by the following equations [Trumbore et al., 1989, 1990; Trumbore, 1993; Harrison et al., 1993b; Wang et al., 1996b, 1998]:

$$\frac{\partial C}{\partial t} = \Phi - kC \quad (1)$$

$$\frac{\partial C^{14}}{\partial t} = \Phi^{14} - (k+\lambda)C^{14} \quad (2)$$

where C and  $C^{14}$  are the organic C and  $^{14}\text{C}$  content ( $\text{mol cm}^{-3}$ ) of the soil layer; the  $\Phi$  and  $\Phi^{14}$  represent the net input of organic C and  $^{14}\text{C}$  ( $\text{mol cm}^{-3} \text{ yr}^{-1}$ ), respectively, in a soil or a soil horizon. The second terms on the right side of (1) and (2) represent loss of organic carbon by microbial decomposition and radioactive decay,  $k$  is the decay rate ( $\text{yr}^{-1}$ ) of organic matter, and  $\lambda$  is the decay constant of  $^{14}\text{C}$  ( $0.0001245 \text{ yr}^{-1}$ ). These equations can be written simply as [Trumbore et al., 1989, 1990; Trumbore, 1993; Harrison et al., 1993b]

$$C_t = C_{t-1} + \Phi - kC_{t-1} \quad (3)$$

$$C_t^{14} = C_{t-1}^{14} + \Phi^{14} - (k+\lambda)C_{t-1}^{14} \quad (4)$$

where  $C_{t-1}$  is C content in the previous year,  $C_t^{14}$  is  $^{14}\text{C}$  content in the previous year,  $k$  is organic C decay rate or  $1/\text{turnover time}$  ( $\tau$ ).

When archived samples of a soil are available for C and  $^{14}\text{C}$  analyses, (3) and (4) can be used to determine the average input rate  $\Phi$  and turnover time ( $1/k$ ) of soil organic matter by best fitting the observed changes in C and  $^{14}\text{C}$  content from the archived to present soils.

Another approach is described by Hsieh [1993]. Hsieh's "active pool" model also assumes that (1) organic matter decomposition and radioactive decay of  $^{14}\text{C}$  are the only mechanisms responsible for C and  $^{14}\text{C}$  loss in a soil C pool and (2) decomposition of organic matter follows the first-order reaction. The second assumption can be stated mathematically as

$$N = N_0 \exp(-kt) \quad (5)$$

Equation (5) can be understood by imagining that organic C added to a soil each year could be painted with a distinct color. If the organic C compounds added to a soil in a particular year were painted red,  $N_0$  would be the initial amount of red C in the soil and  $N$  would be the amount of red C left in the soil after time  $t$ . Under the above assumptions, the specific activity of the soil C pool can be expressed as [Hsieh, 1993]

$$A_a = \frac{\sum_{i=b}^y \{\Phi_i \exp[-k(y-i)]\} \{^{14}\text{C}_i \exp[-\lambda(y-i)]\}}{\sum_{i=b}^y \Phi_i \exp[-k(y-i)]} \quad (6)$$

where

- $A_a$  the specific  $^{14}\text{C}$  activity of the C pool;
- $\Phi_i$  the organic C input to the C pool in year  $i$ ;
- $k$  the decomposition rate of organic matter;
- $^{14}\text{C}_i$  the specific  $^{14}\text{C}$  activity of the new C added in year  $i$ , which is the same as that of the atmosphere for that particular year;
- $y$  the year of sampling;
- $b$  the initial year of model simulation;
- $\lambda$  decay constant of  $^{14}\text{C}$ .

The nominator and denominator in (6) represent the specific  $^{14}\text{C}$  activity and the amount of C accumulated, respectively, since year  $b$ . Sensitivity test shows that  $A_a$  is not sensitive to random fluctuations in  $\Phi_i$  and can therefore be approximated by

$$A_a = \frac{\sum_{i=b}^y \{ \exp[-(y-i)/\tau] \} \{^{14}\text{C}_i \exp[-\lambda(y-i)]\}}{\sum_{i=b}^y \exp[-(y-i)/\tau]} \quad (7)$$

where  $\tau$  is turnover time (i.e.,  $\tau=1/k$ ).

Since the  $^{14}\text{C}$  activities of the atmosphere are well documented, (7) can be used to predict the relationship between the  $^{14}\text{C}$  content of an "active" C pool and its turnover time for a specific sampling year (Figure 1). The turnover time  $\tau$  is equivalent to the "mean life" commonly used by isotope geochemists [Faure, 1986]. It represents the average life expectancy of a C atom in a C pool. If we define  $T_{1/2}$  as the "half-life" of organic C (i.e., the time needed for half of the initial C to decay away), then the following relationships can be derived from (5):

$$T_{1/2} = \ln 2/k$$

$$= \tau \ln 2,$$

or

$$\tau = T_{1/2} / 0.693.$$

When  $t = 10 T_{1/2}$  or  $t = 6.93 \tau$ , the initial C has been reduced by a factor of  $2^{-10}$  due to decomposition. In other words, after 10 half-lives or about seven turnover times, almost all the initial C has been turned over. In practice, the period of model simulation ( $y - b$ ) should be at least 3 times longer than the turnover time in order for (7) to yield a stable solution. Hsieh's [1993] model, which also takes advantage of the excess bomb  $^{14}\text{C}$ , applies to active C pools with turnover time of decades or less (Figure 1). In this approach, archived samples are not required. If a soil horizon is dominated by an active C pool with a turnover time of decades or less, the above two approaches should yield the same results.

In addition to the two methods discussed above, other approaches such as the "paired plot method" [Hsieh, 1992] have also been applied to the study of soil C cycle and will not be discussed here because of their irrelevance to this study.

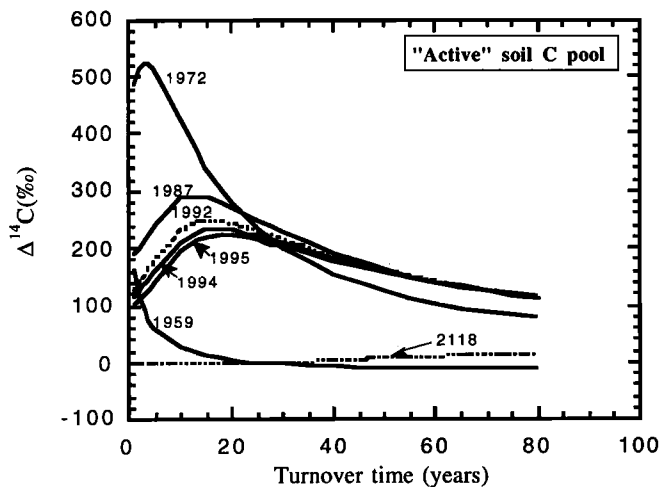
### 3. Results and Discussion

#### 3.1. $\text{CO}_2$ Production in Natural and Disturbed Soils

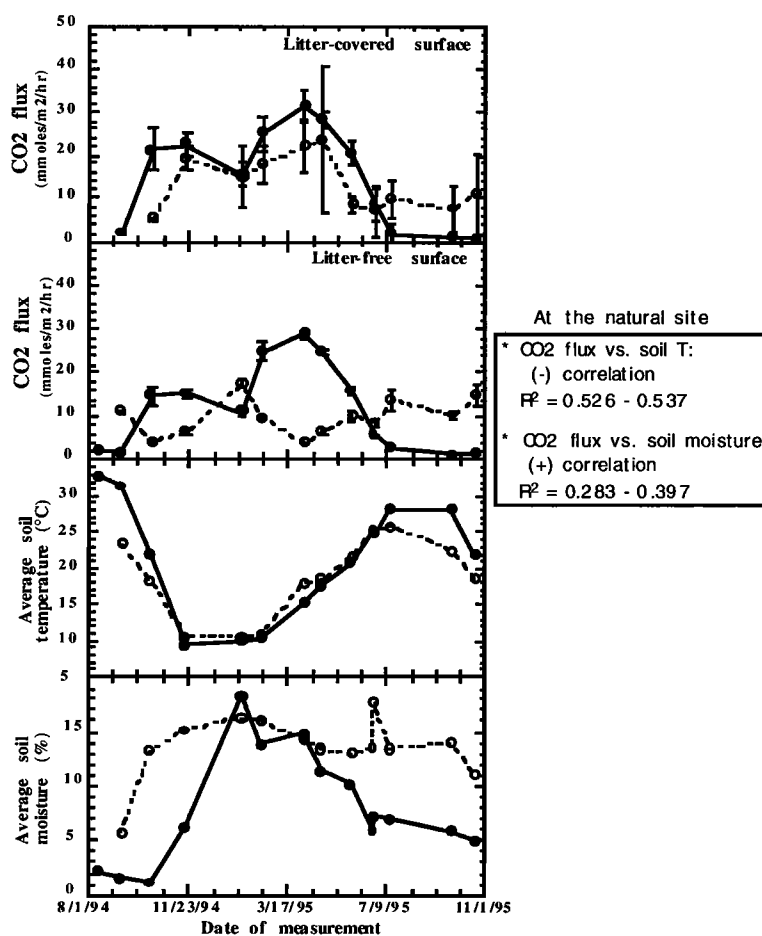
$\text{CO}_2$  is produced in soils primarily by root respiration and microbial decomposition of organic matter. The  $\text{CO}_2$  flux defines the rate of C cycling through soils, thereby constraining estimates of C inputs from aboveground and belowground biomass production and root respiration rates. The impacts of human activities on soil  $\text{CO}_2$  flux out of soils are poorly documented and vary among sites.

The western slope of the Sierra Nevada has the typical Mediterranean climatic pattern of dry summers and wet winters. In this environment, plants grow seasonally in cool months when adequate soil moisture is available. At the low elevation sites (Fallbrook (FB) and FB-OCD), in the foothills of the Sierra Nevada, we monitored the  $\text{CO}_2$  production as well as soil temperature and moisture contents at 10, 25, 50, and 100 cm depths in both natural and cultivated soils throughout a 1-year cycle. Both soils displayed strong seasonal variations in soil temperature (Figure 2). In the summer and fall, the average soil temperatures at the cultivated site were a few degrees lower than at the natural site owing to irrigation and shading by mature lemon trees. The average moisture content also varied seasonally (Figure 2). In the natural soil, which depended entirely on precipitation for soil moisture, the water content ranged from less than 5 wt. % in the late summer and fall to about 18% in the spring, following the winter rains. At the disturbed site, soil moisture remained high throughout most of the year because of irrigation.

Soil  $\text{CO}_2$  flux showed a strong seasonal cycle, with the highest  $\text{CO}_2$  flux at the grassland site in the late fall and in the spring, in response to more available moisture and the growth of the annual grasses (Figure 2). Although soil  $\text{CO}_2$  flux varies across the landscape, there is a strong negative correlation between  $\text{CO}_2$  flux and average soil temperature and



**Figure 1.** The  $\Delta^{14}\text{C}$  values (‰) of soil organic matter predicted by Hsieh's [1993] "active pool" model. Curves represent the relationships between the  $^{14}\text{C}$  content of an active pool and its turnover time for various sampling years from 1959 to 2118.



**Figure 2.** Seasonal variations of soil CO<sub>2</sub> flux, average soil temperature, and soil moisture at low-elevation sites (~470 m). Solid dots represent the natural grassland soil, and open circles denote the agricultural soil. Error bars represent the standard deviation of a minimum of 28 measurements. Temperature and moisture measurements were made at 10, 25, 50, and 100 cm depths within the soil profiles using permanently installed sensors.

a positive correlation between CO<sub>2</sub> flux and average soil water content at the natural grassland site (Table 1). The correlation coefficient  $R^2$  for the linear flux-temperature models are 0.526 and 0.537 for litter-free and litter-covered surfaces, respectively, indicating that temperature alone can explain ~53% to 54% of the total variation in CO<sub>2</sub> flux in the natural grassland. For the linear flux-moisture models the  $R^2$  are 0.397 for litter-free surfaces and 0.283 for litter-covered surfaces, suggesting that moisture alone can explain ~28% to ~40% of the total variation in CO<sub>2</sub> flux at the grassland site. In contrast, there is no significant linear correlation with either soil temperature or moisture at the agricultural site (Table 1).  $F$  tests show that multiple regression models including both temperature and moisture do not have significant improvements over the simple linear models in terms of explaining the variation of CO<sub>2</sub> flux.

The negative correlation between CO<sub>2</sub> flux and soil temperature is in contrast to what has been observed in other ecosystems [Kucera and Kirkham, 1971; Kowalenko et al., 1978; Raich and Schlesinger, 1992; Howard and Howard, 1993; Raich and Potter, 1995], but is typical for Mediterranean ecosystems in which grasses are biologically active only during cool, wet months [Amundson and Smith,

1988]. The annual CO<sub>2</sub> flux out of the soil from humus decomposition and root respiration averaged 1.4 kgC m<sup>-2</sup> yr<sup>-1</sup> for the grassland soil and 1.2 kgC m<sup>-2</sup> yr<sup>-1</sup> for the agricultural soil (Table 2). These values are higher than those predicted by the CO<sub>2</sub> flux models of Raich and Potter [1995].

At the high-elevation forested sites (Windy and Windy-cut), measurements were only made in July and October (Table 2). The average CO<sub>2</sub> flux was 2.1 kgC m<sup>-2</sup> yr<sup>-1</sup> at the old-growth forest site and 2.2 kgC m<sup>-2</sup> yr<sup>-1</sup> at the logged site in July and was significantly reduced at both sites in October when the temperature was lower. However, we do not have enough data to determine the correlation coefficient between the CO<sub>2</sub> flux and temperature and moisture at these sites.

### 3.2. Effect of Land Use Change on C Storage in Soils

It is well documented that cultivation results in a substantial loss of organic C [Johnson, 1992; Davidson and Ackerman, 1993; Post and Mann, 1990; Paul et al., 1997], but inconsistent results have been obtained on the effects of logging on C storage and turnover in soils. Most studies show no significant change ( $\pm 10\%$ ) with logging, a few studies show large net losses, and a few studies also show a net gain

**Table 1.** Least Squares Linear and Multiple Regression Models Describing the Relationship Between Soil CO<sub>2</sub> Flux and Average Soil Temperature *T* and Average Water Content *M*

Site	Model			R <sup>2</sup>	<i>p</i> <sup>a</sup>	<i>n</i>
	<i>a</i>	<i>b</i>	<i>c</i>			
Natural soil, litter free <sup>b</sup>	30.338	-0.90522		0.526	0.005	12
	2.2698		1.179	0.397	0.021	12
	23.193	-0.711	0.397	0.547	0.019	12
Natural soil, litter covered	36.701	-1.1045		0.537	0.0067	11
	5.2721		1.1451	0.283	0.0752	11
	35.029	-1.059	0.093	0.538	0.031	11
Disturbed soil, litter free	9.5598	0.005036		4.3 × 10 <sup>-5</sup>	0.984	11
	11.569		-0.1455	0.009	0.768	11
	13.21	-0.05	-0.2	0.012	0.947	11
Disturbed soil, litter covered	25.136	-0.65384		0.317	0.0712	11
	-10.116		1.6951	0.155	0.2313	11
	21.565	-0.617	0.21	0.319	0.2156	11

Here soil CO<sub>2</sub> flux =  $a + bT + cM$ . Soil CO<sub>2</sub> fluxes are in mmole m<sup>-2</sup> h<sup>-1</sup>, *T* is in degrees Celsius, *M* is weight percent of water content. R<sup>2</sup> is correlation coefficient, *p* is observed significance level, and *n* is number of mean monthly flux values used in the regression model; each mean monthly flux value is an average of at least 28 flux measurements.

<sup>a</sup> Correlation is significant when  $p < 0.05$ .

<sup>b</sup> Measurements taken after surface litters were removed.

following logging [Johnson, 1992]. In any case, the dynamics of organic C following disturbance is poorly known.

Comparison of C inventories of the natural grassland soil (FB) and the agricultural soil (FB-OCD) clearly shows that agricultural practices have resulted in a significant reduction in the amount of C stored in soils (Figure 3), which is consistent with other studies. Most of the C loss occurred in the upper horizons (upper 30 cm), and below ~30 cm depth, both natural and disturbed soils have about the same C content. If we assume that the disturbed soil had the same total amount of organic C in 1972 (the time of land conversion) as now at the natural site, the amount of C lost would be 1535 gC m<sup>-2</sup> (from 5839 gC m<sup>-2</sup> in 1972 to 4304 gC m<sup>-2</sup> in 1994), which amounts to a total reduction of ~26% in C storage in 22 years at the disturbed site. The loss of C over the 22 years of cultivation shows the combined effects of rapid decomposition of soil organic matter coupled with a significant reduction in C additions under cultivation.

The C content of the old-growth forest soil is also very different from that of the logged forest soil (Figure 3). Our data show that logging has caused a significant decrease in soil C storage with most of the C loss from the upper soil horizons

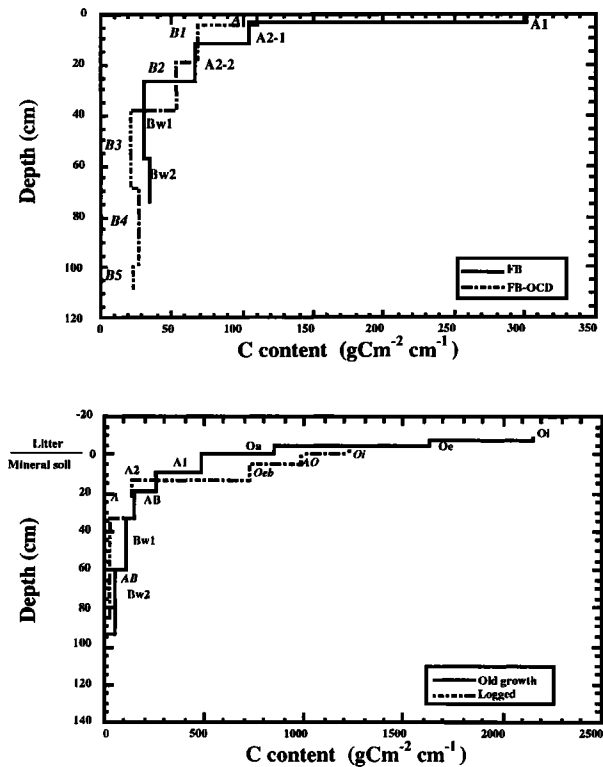
(Figure 3). Big temperature differences in the summer between logged and natural sites (Table 2) have certainly contributed to the C loss in the logged soil because much higher summer temperatures would have accelerated the decomposition of organic matter at the logged site. The amount of C lost from the logged soil (0-128 cm) is ~30% (from 27,480 to 19,265 gC m<sup>-2</sup>) assuming that the old-growth forest soil is at steady state and the soil at the logged site had the same C content before logging as now at the natural site. The local land use history in our logged soil study area is not precisely known. The youngest tree ring of a stump near our site has a Δ<sup>14</sup>C value of 380‰, suggesting that the tree was cut either around 1962 or 1974. The average loss rate would be ~249 gC m<sup>-2</sup> yr<sup>-1</sup> if the site was logged in 1962 or ~391 gC m<sup>-2</sup> yr<sup>-1</sup> assuming that the site was disturbed in 1974. In either case, the amount of C lost from the soil is substantial and, again, surface soils prove to be most sensitive to disturbance.

### 3.3. Effect of Land Use Change on <sup>14</sup>C Distribution in Soils

There are large differences in <sup>14</sup>C content of soil organic matter between the natural and the disturbed soils (Figure 4).

**Table 2.** CO<sub>2</sub> Flux at the Natural and Disturbed Sites

Site	Average CO <sub>2</sub> Flux (kgC m <sup>-2</sup> yr <sup>-1</sup> )		Mean Annual T, °C	Precipitation, mm	Land Use
	Litter Free	Litter Covered			
Fallbrook	1.32	1.56	17.8	31	natural grassland
Fallbrook-OCD	1.01	1.4	17.8	31	lemon orchard
Windy	2.05, July	2.11, July	16.1, July	102	old-growth forest
	1.0, Oct.	0.77, Oct.	12.1, Oct.	102	
Windy-cut	2.17(July)	2.18(July)	30(July)	102	logged forest
	0.57(Oct.)	0.4(Oct.)	11.1(Oct.)	102	

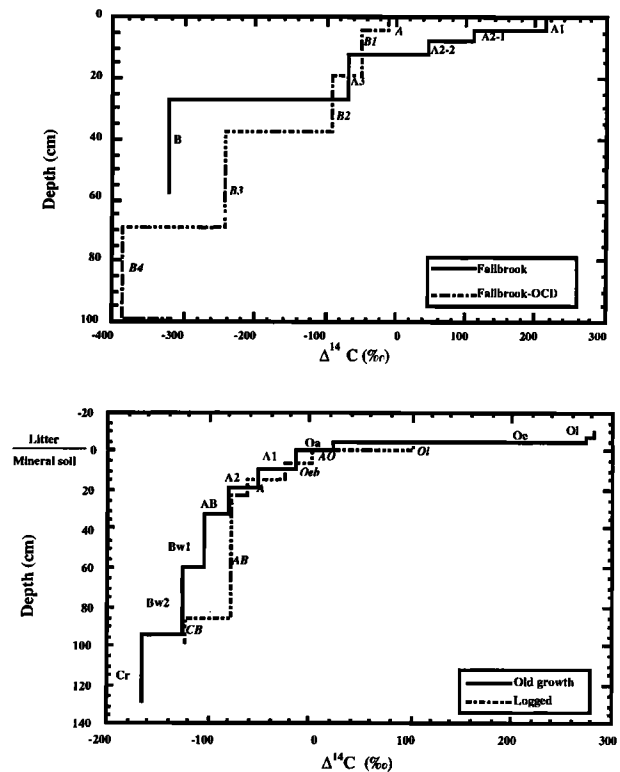


**Figure 3.** Comparison of C inventories between the natural ((top) Fallbrook (FB) and (bottom) old growth) and disturbed ((top) FB-OCD and (bottom) logged) soils, showing that (1) surface horizons, including the litter layer (the O horizon), of the disturbed soils are much thinner and contain much less C than their natural counterparts, and (2) subsoils at both natural and disturbed sites (below ~40 cm) have about the same amount of C.

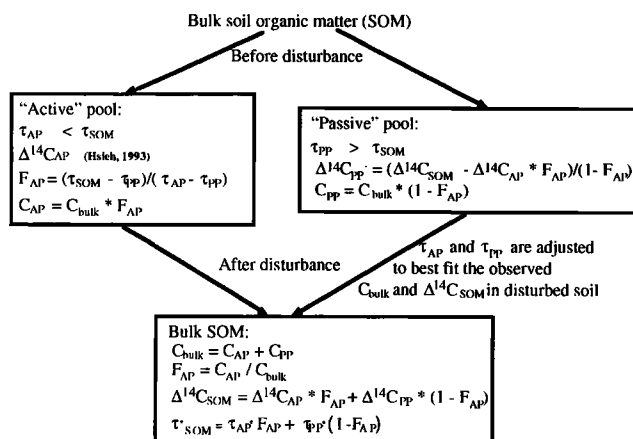
The  $\Delta^{14}\text{C}$  values of organic matter in the upper soil horizons (the upper 30 cm depth interval) at the disturbed sites are significantly lower than those of their natural counterparts, suggesting a preferential loss of C from younger C pools. However, organic matter in subsoil horizons (from ~30 to 69 cm in the cultivated soil and from ~30 - 80 cm in the logged soil) is relatively enriched in  $^{14}\text{C}$  at the disturbed sites, which suggests increased input of C in these soil horizons. This interpretation is supported by field observation of root distributions in these soils. At the natural grassland site, grass roots are abundant in the shallow horizons and the density of roots dropped drastically below ~30 cm depth. At the cultivated site, roots were present commonly in the B horizon, with almost no visible roots in the A horizon (i.e., the upper 5 cm). Therefore the shift from shallow-rooted grasses to deeply rooted trees have obviously shifted C inputs into various horizons. Although establishment of "permanent" tree crops is very important along the mountain front from the cities of Fresno to Bakersfield, it is only one of many types of agriculture in the San Joaquin Valley. It is therefore likely that not all soils will show this trend in C dynamics because many cropping systems are devoted to annual crops. At the logged site, the apparent increase in  $^{14}\text{C}$  content below 30 cm depth could be an artifact of the decay of dead tree roots that were high in  $^{14}\text{C}$ . Also, the rooting pattern

may have shifted from the old-growth forest because of the invasion of bushes at the logged site. It is also possible that the  $^{14}\text{C}$  enrichment in subhorizons is due in part to downward translocation of young C from surface horizons in dissolved or particulate forms.

If we consider soil organic matter as one single pool, decomposition of organic matter and radioactive decay of  $^{14}\text{C}$  (half-life of 5730 years) between the time of land use change and the time of sampling would not be sufficient to produce the observed C and  $\Delta^{14}\text{C}$  values in shallow soil horizons observed at the disturbed sites. A minimum of two pools with different turnover times is required to explain the observed changes in C and  $^{14}\text{C}$  content of soil organic matter in disturbed soils because soil organic matter is likely a complex series of compounds having different turnover times. Conceptually and mathematically, soil organic matter can be divided into multiple pools as long as mass balance requirements are met. For simplicity, we assume that (1) soil organic matter can be represented by two pools, an "active" pool with a turnover time less than the turnover time for bulk soil organic matter and a "passive" pool with a turnover time greater than the turnover time for bulk soil organic matter (Figure 5); (2) the  $^{14}\text{C}$  content of the active pool in natural soils follows the relationship described by Hsieh's [1993] model as shown in Figure 1, and the  $^{14}\text{C}$  content of the "passive" pool can be calculated from mass balance considerations; and (3) the



**Figure 4.** Carbon 14 content of bulk organic matter in natural ((top) Fallbrook and (bottom) old growth) and disturbed ((top) Fallbrook-OCD and (bottom) logged) soils. The  $\Delta^{14}\text{C}$  values of organic matter in the upper soil horizons are lower at the disturbed sites than at the natural sites, showing the effects of preferential loss of C from younger C pools.



**Figure 5.** Conceptual two-pool model for soil organic matter. Here  $\tau_{SOM}$  is the turnover time of bulk soil organic matter in the natural soil;  $\tau_{AP}$  and  $\tau_{PP}$  are turnover times of the “active” and “passive” C pools, respectively.  $F_{AP}$  represents the fraction of C in the active pool.  $C_{AP}$  and  $\Delta^{14}C_{AP}$  are the C and  $^{14}C$  contents, respectively, of the active pool; and  $C_{PP}$  and  $\Delta^{14}C_{PP}$  denote the C and  $^{14}C$  contents, respectively, of the passive pool.  $C_{bulk}$  represents the total C content in the bulk soil;  $\tau_{SOM}^*$  is the average turnover time of bulk organic matter in disturbed soil.

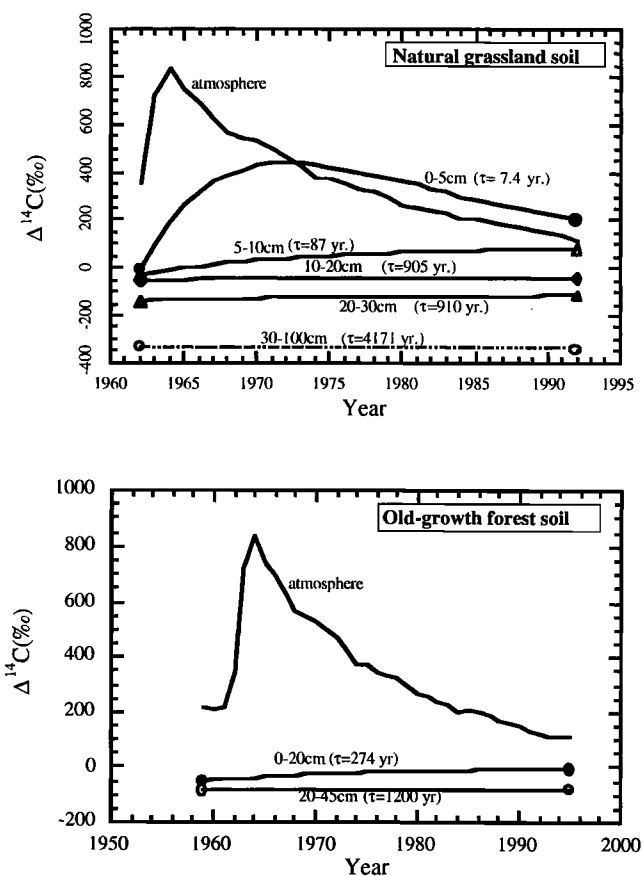
initial sizes and  $^{14}C$  contents of C pools in disturbed soils at the time of land use change were the same as in their natural counterparts. Using these assumptions and applying the time-dependent box model (i.e., (3) and (4)) discussed in section 2.2 to best fit the observed changes in bulk C and  $^{14}C$  content from archived to present soils (Figure 6), we estimated the turnover times of soil C pools (Table 3).

For natural soils, because the prebomb samples of the litter horizon (O) and surface soil horizon (A3) for the forest soils (Table 4) were missing, we were unable to determine the turnover time of organic matter in these two horizons. However, very positive  $\Delta^{14}C$  values of the 1995 litter layer samples (281‰ for the Oi (9–7 cm) and 274‰ for the Oe (7–4 cm) subhorizon, see Table 4) suggest that organic C pools in these horizons have an average turnover time of either ~9 or ~18 years based on Hsieh’s [1993] “active pool” model (Figure 1), assuming that foliage remains on the trees for ~5 years. Our results suggest that the organic C in top soil horizon (upper 5 cm) of natural soils is turning over very fast, with an apparent turnover time of 7.4 years in the natural grassland soil and less than 20 years in the old-growth forest soil (Table 3). The turnover times of organic C in both grassland and old-growth forest soils increase with depth from decades or less to hundreds or even thousands of years at deeper depths. The shorter turnover times of organic C in shallow soil horizons imply that these C pools will respond significantly to perturbations such as changes in land use or climate.

For the agricultural soil, the  $^{14}C$  model failed to fit the observed changes in both C and  $^{14}C$  content of the surface soil horizon unless the assumption of physical removal of top soil is made. In our modeling exercise, we assumed that the top 1 cm of the soil was physically removed during the process of conversion from grassland to agricultural land. The model-

estimated current turnover times for organic C in the agricultural soil are longer than those for the natural grassland soil. For the logged soil, prebomb litter samples are not available for comparison. However, the 1995 sample of the Oi horizon (Table 4) has a  $\Delta^{14}C$  value of 103‰, which would suggest a turnover time of either ~2 or ~90 years, if we assume that this horizon can be represented by an “active” C pool (Figure 1). A turnover time of 2 years would seem too fast for this environment.

The model-derived input rates and turnover times of soil organic matter in disturbed soils are very different from their natural counterparts. These differences represent changes in the sizes of the active and passive C pools (Table 3). The relative size of the active pool has been reduced in the disturbed soils in comparison with their natural counterparts for equivalent depth intervals because of a preferential loss of C from the active pools. Our data demonstrate that land use change, either conversion of grassland to agricultural land or logging, has drastically altered the C balance and turnover in soils and that shallow soil horizons are most susceptible to land use change because of shorter turnover times of C pools in these horizons.



**Figure 6.** “Bomb”  $^{14}C$  uptake in natural (top) grassland and (bottom) old growth forest soils. Curves represent the  $^{14}C$  content of the atmosphere and the model-calculated  $^{14}C$  content of the soil at different depth intervals;  $\tau$  represents the average turnover time that best fits the observed C and  $^{14}C$  variations from the pre- to postbomb soils.



**Table 3.** Model-Derived Input Rate and Turnover Time  $\tau$  of Organic Matter in Soils in the Sierra Nevada, California

Depth, <sup>a</sup> cm	Input Rate, $\times 10^6$ mol $\text{cm}^{-3}$ $\text{yr}^{-1}$	Average $\tau$ , <sup>b</sup> years	$\tau_{\text{AP}}$ , <sup>c</sup> years	$\tau_{\text{PP}}$ , <sup>c</sup> years	Relative Size of "Active" Pool <sup>d</sup> , %
<i>Natural Grassland Soil<sup>e</sup></i>					
0-5	272.88	7.4	7.4		100
5-10	9.27	87	87		100
10-20	0.65	905	3	1000	10
20-30	0.78	910	3	1000	9
30-100	0.065	4171			
<i>Agricultural Soil</i>					
0-5	33.1	42.08	3	42.1	1
5-10	84.1	139	3	308	55
10-20	12.55	943	3	1000	6
20-30	10.91	1237	3	1297	5
30-100	0.065	4171			
<i>Old-Growth Forest Soil<sup>f</sup></i>					
5-0		~9 or ~18	~9 or ~18		100
0-20	18	274	3	300	9
20-45	0.9	1200	10	2600	54
<i>Logged Forest Soil</i>					
2-0		~2 or ~90	~2 or ~90		
0-20	139.8	279	3	300	7
20-45	7.3	2363	10	2600	9

<sup>a</sup> The C and  $^{14}\text{C}$  contents for a given depth interval in our modeling exercise are weighed mean average values calculated from the C and  $^{14}\text{C}$  data in Table 4.

<sup>b</sup> Values represent average turnover time for the natural soils and current turnover time for the disturbed soils.

<sup>c</sup> Here  $\tau_{\text{AP}}$  and  $\tau_{\text{PP}}$  represent the turnover times of the active and passive pools, respectively, that were used in the model to best fit the observed changes in C and  $^{14}\text{C}$  content of the soils.

<sup>d</sup> Relative size  $F_{\text{AP}}$  is calculated from  $\tau_{\text{som}} = \tau_{\text{AP}} F_{\text{AP}} + \tau_{\text{PP}} (1 - F_{\text{AP}})$ .

<sup>e</sup> Total input rate is  $0.0014$  mol C  $\text{cm}^{-2}$   $\text{yr}^{-1}$ , ~14% of total  $\text{CO}_2$  flux.

<sup>f</sup> Total input rate for 0 – 45 cm depth interval is  $0.0004$  mol C  $\text{cm}^{-2}$   $\text{yr}^{-1}$ , ~ 4% of total  $\text{CO}_2$  flux in July.

When a soil is at steady state, the C input to a soil is balanced by the C loss from decomposition of organic matter. Soil  $\text{CO}_2$  is produced by root respiration and decomposition of soil organic matter. Most studies suggest that root respiration contributes ~30 - 70% of the total soil  $\text{CO}_2$  flux [Schlesinger, 1977; Raich and Schlesinger, 1992]. The comparison of our model-derived annual C input rates with measured average  $\text{CO}_2$  fluxes at our sites (Tables 2 and 3) suggests that ~14% of the  $\text{CO}_2$  production in the natural grassland soil is from decomposition of organic matter, assuming that the natural grassland soil is at steady state (that is, >80% of  $\text{CO}_2$  is derived from root respiration). For the old-growth forest soil, our model-derived input rates for a 0-45 cm depth interval would suggest that at least 4% of the  $\text{CO}_2$  is produced from microbial decomposition of soil organic matter. Our estimates of  $\text{CO}_2$  production from organic matter decomposition are lower than expected [Schlesinger, 1977; Raich and Schlesinger, 1992]. There are likely several reasons for this: (1) we only measured  $\text{CO}_2$  flux during the day, when soil respiration tends to be higher than at night, and (2) C input and decay in the litter horizons (i.e., the O horizons) and deeper horizons at the forest site could not be determined because of the lack of prebomb samples of these soil horizons. If we assume litter horizon at the old-growth forest site is at steady state, the C

flux from the litter horizon would be ~ 513 gC  $\text{m}^{-2}$   $\text{yr}^{-1}$  (that is, flux equals C content in litter horizon/turnover time) using a turnover time of 18 years. This would suggest that decomposition of organic matter in the upper 50 cm of the soil (including the litter horizon) contributed at least 27% of total  $\text{CO}_2$  flux at the old-growth forest site in the summer and more than 56% of the total  $\text{CO}_2$  flux in the fall. However, if we use a turnover time of 9 years, the C flux from the litter horizon would be ~ 982 gC  $\text{m}^{-2}$   $\text{yr}^{-1}$ . This would suggest that at least 50% of total  $\text{CO}_2$  flux in the summer and about 100% of total  $\text{CO}_2$  flux in the fall were derived from organic matter decomposition at the old-growth forest site.

#### 4. Conclusions

Our study shows that soil temperature and moisture are important, but not the sole, factors affecting  $\text{CO}_2$  production in soils. In a Mediterranean climate, soil  $\text{CO}_2$  flux is positively correlated with soil moisture but negatively correlated with soil temperature. This negative correlation between soil  $\text{CO}_2$  flux and soil temperature is in contrast to what has been observed in other ecosystems. Conversion of grassland to agriculture land or logging has resulted in significant decreases in total soil C (~26% and 30% reduction,

**Table 4.** Carbon and  $^{14}\text{C}$  Content of Soils in Central Sierra Nevada, California

Soil Horizon	Depth, cm	C, g m <sup>-2</sup> cm <sup>-1</sup>	$\Delta^{14}\text{C}$ , ‰	$\delta^{13}\text{C}$ , ‰
<i>Fallbrook 1992</i>				
Litter			98.3±6.7	-27.6
A1	0-4	302	208.9±8.7	-27.8
A2-1	4-8	105	106.8±8.1	-26.7
A2-2	8-12	116	39.9±6.7	-26.8
A3	12-27	68	-72.6±6	
Bw1	27-58	31	-326.3±4.3	
Bw2	58 +	35	-382.5±4	
Total <sup>a</sup>		5839		
<i>Fallbrook-OCD 1994</i>				
A	0-5	112	-17±8.1	-25.3
B1	5-19	70	-54.8±6.9	-25.5
B2	19-38	55	-92.1±5.9	
B3	38-69	23	-246.4±5.5	-24.8
B4	69-99	27	-390.2±5	-24.5
B5	99-108	25	-316.6±4.5	-24.5
Total (0-108cm)		4304		
<i>Fallbrook 1962<sup>b</sup></i>				
	0-10.2	155	-15	
	10.2-22.9	64	-51	
	22.9-48.3	36	-207	
	48.3-76.2	32	-389	
	76.2-111.7	18	-390	
Total <sup>c</sup>		4840		
<i>Windy, Old Growth Forest 1995<sup>d</sup></i>				
Oi	9-7 (5-3)	2155	281.7±8.2	-25.2
Oe	7-4 (3-0)	1640	274.9±7.3	-24.8
Oa	4-0 (0-4)	860	21.8±5.9	-24.5
A1	0-10 (4-14)	489	-14.2±6.8	
A2	10-19 (14-23)	262	-51.3±6.2	
AB	19-33 (23-37)	150	-80.8±6.4	
Bw1	33-60 (37-64)	121	-104.2±6.2	
Bw2	60-94 (64-98)	59.6	-123.6±6.1	
Cr	94-128 (98-132)	4.5	-165.5±7.2	
Total <sup>c</sup>		27480		
<i>Windy-Cut, Logged 1995</i>				
Oi	2-0	1226	102.5±7.1	
AO	0-6	995	2.8±6.5	
Oeb	6-14	722	-23.3±6.8	
A	14-23	142	-60.2±6.3	
AB	23-33	156	-77.8±6.3	
BA	33-86	40	-77.7±5.9	
CB	86+	12	-122.1±4.6	
Total <sup>c</sup>		19265		
<i>Windy 1959</i>				
O	5.08-0	---	---	
A1	0-17.8	753	-52.7±6	
A2	17.8-45.7	199	-85.5±5.8	
A3	45.7-73.7	---	---	
BC	73.7-127	20.8	-161.0±5.3	

<sup>a</sup> Total is based on 0 – 108 cm.<sup>b</sup> Data are from *Trumbore et al.* [1996].<sup>c</sup> Total is based on 0 – 111.7 cm.<sup>d</sup> Depth intervals in parentheses were used in calculating turnover times.<sup>e</sup> Total is based on 0 – 128 cm.

respectively) and  $^{14}\text{C}$  content as a result of changes in C input rates and turnover times in soils. Most of the C loss is from the upper horizons. In subsoil horizons of the agricultural and logged soils examined here, organic matter is relatively enriched in  $^{14}\text{C}$ , suggesting increased C inputs in these horizons. Turnover times of organic C pools increase with depth from decades or less to hundreds or even thousands of years. In natural soils, most of the C in the surface soil horizon resides in active C pools with turnover times of decades or less. In disturbed soils, the current turnover times are apparently increased if compared to their natural counterparts owing to a preferential loss of C from active pools. At deeper depths, the organic C is dominated by C pools with turnover times greater than 1000 years.

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