

(Fig. 1d). The data show that macrofauna, unlike meiofauna, do not discriminate among sources of organic matter with different fossil-carbon contents and sequester their body carbon in direct proportion to sediment TOC, regardless of the specific quality of the organic matter (for example, C/N is ~ 20 at site A⁹).

The large number of documented hydrocarbon seeps in the southern California continental borderland²⁻⁴ indicates that the extensive biogeochemical cycling of hydrocarbon-derived carbon described here in Isla Vista seep sediments could be a widespread regional, if localized, phenomenon. Rates of important geochemical processes including community oxygen meta-

bolism and sulphate reduction are significantly enhanced in seep sediments⁹. Although sediments at the Isla Vista seep have similar types of organisms as non-seep coastal sediments, the numbers of organisms and the distribution of individual taxa are significantly altered⁵⁻⁷. We suggest that these processes and distributions result, at least in part, from extensive incorporation of fossil carbon into seep sediments and its subsequent use as a carbon and energy source by sedimentary assemblages. Careful extrapolation of the information derived from studies of natural hydrocarbon seepage may also provide information on the fate and effects of chronic hydrocarbon inputs to other nearshore sedimentary environments. □

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Evidence from chronosequence studies for a low carbon-storage potential of soils

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OVER most of the Earth's land surface, the amount of carbon stored in soil organic matter exceeds by a factor of two or three the amount stored in living vegetation. This pool of soil carbon is large (1.5×10^{18} g)^{1,2} and plays a dynamic part in the geochemical carbon cycle. Prentice and Fung³ have suggested that terrestrial vegetation and soils would act as a large sink for atmospheric carbon dioxide if its concentration were twice the present level. Here I use data from chronosequence studies to show that the production of refractory humus substances in soils sequesters only $\sim 0.4 \times 10^{15}$ g C yr⁻¹ from the atmosphere, accounting for just 0.7% of terrestrial net primary production. Moreover, agricultural practices tend, on balance, to cause a release of soil carbon to the atmosphere^{4,5}. Thus if the terrestrial biosphere is indeed to act as a carbon sink under future elevated levels of carbon dioxide, this would be more likely to be the result of changes in the distribution and biomass of terrestrial vegetation than of changes in the accumulation of soil organic matter.

The carbon stored in soil organic matter represents the long-term net balance of photosynthesis and total respiration in terrestrial ecosystems. Radiocarbon dates of humic materials in the lower soil profile are frequently $>1,000$ years⁶. New land surfaces produced by disturbances such as volcanic eruptions or glacial retreat yield youthful soils with little or no organic matter. In these soils, the long-term average rate of accumulation of soil organic matter is easily calculated from measurements of bulk density and the percentage of organic carbon at each depth in the soil profile of depth N . Soil organic carbon (g C m^{-2}) = $\sum_{D=0}^N 10,000 D \rho(D) P(D)$ where $\rho(D)$ and $P(D)$ are the bulk density in g cm^{-3} and the percentage of organic carbon,

respectively, in depth interval D . The long-term average rate of accumulation is obtained by dividing the carbon content of the profile by the estimated or measured age. Interval-specific accumulation rates can be calculated from the difference between two soil profiles in the same chronosequence, divided by the difference in their ages.

During soil development, organic matter often shows an initial period of rapid increase for up to 3,000 years, followed by a lower rate of accumulation that may continue for millennia⁷. At any time, the long-term average rate of accumulation is an overestimate of the current rate of accumulation in most soils (Fig. 1). As most soils are at least 3,000 years old, however, the average long-term rate for 3,000- to 10,000-year-old soils provides an upper limit for estimates of soil carbon accumulation during the Holocene and in possible future climates.

In upland ecosystems, the long-term rate of carbon storage varies from $0.2 \text{ g C m}^{-2} \text{ yr}^{-1}$ in some polar deserts to $>10 \text{ g C m}^{-2} \text{ yr}^{-1}$ in some forests (Table 1). Considering the paucity of studies and the range of environments represented, the rates are surprisingly uniform. The long-term rate for 16 soils that are 3,000 to 10,000 years old, $2.4 \pm 0.70 \text{ g C m}^{-2} \text{ yr}^{-1}$ (1 s.e.) (Table 1), can be used to calculate the maximum potential for the present accretion of soil organic matter.

An annual production of $2.4 \text{ g C m}^{-2} \text{ yr}^{-1}$ of humic substances implies a total storage of $0.32 \times 10^{15} \text{ g C yr}^{-1}$ in upland ecosystems under natural vegetation ($133 \times 10^6 \text{ km}^2$; ref. 5). The actual net storage of organic carbon in upland soils is probably less than $0.32 \times 10^{15} \text{ g C yr}^{-1}$, because this value assumes that all upland soils are only 3,000-10,000 years old. Following a similar approach, Armentano and Menges⁸ found that before recent disturbance, soils in wetlands, bogs and peatlands ($3.5 \times 10^6 \text{ km}^2$) were a sink for up to $0.08 \times 10^{15} \text{ g C yr}^{-1}$. Their value may be too low, given recent higher estimates of global wetland area⁹. Nevertheless, assuming these estimates are reasonable, the total storage of soil organic matter ($0.40 \times 10^{15} \text{ g C yr}^{-1}$) is $\sim 0.7\%$ of recent estimates of net primary production on land ($60 \times 10^{15} \text{ g C yr}^{-1}$; ref. 10), but 20% of that storage occurs in wetland ecosystems which cover only 2% of the land surface. A similar approach shows that the net storage of carbon in desert-soil carbonate is also a minor flux in the global carbon cycle ($<0.023 \times 10^{15} \text{ g C yr}^{-1}$; ref. 11).

TABLE 1 Long-term rates of accumulation of organic carbon in Holocene-age soils

Ecosystem type	Vegetation in terminal state	Soil origin	Accumulation interval (yr)	Long-term rate of accumulation ($\text{g C m}^{-2} \text{ yr}^{-1}$)	Ref.
Tundra	Polar desert	Glacial retreat	8,000	0.2	32
	Polar desert	Glacial retreat	9,000	0.2	33
	Polar desert	Glacial retreat	2,600	2.4	34
	Sedge moss	Glacial retreat	1,000	2.4	33
	Sedge moss	Glacial retreat	9,000	1.1	33
	Sedge moss	Glacial retreat	8,700	5.7	34
Boreal forest	Spruce	Glacial retreat	3,500*	11.7	35
	Spruce-fir	Glacial retreat	5,435	0.8	16
	Spruce-fir	Glacial retreat	2,740	2.2	17
Temperate forest	Broadleaf evergreen	Volcanic ash	1,277	12.0	36
	Coniferous	Volcanic mudflow	1,200	10.0	27
	Deciduous	Alluvium	1,955	5.1	37
	Deciduous	Dunes	10,000	0.7	38
	<i>Podocarpus</i>	Dunes	10,000	2.1	28
	<i>Angophora</i>	Dunes	4,200	1.7	39
	<i>Eucalyptus</i>	Dunes	6,500	1.4	40
	<i>Eucalyptus</i>	Dunes	5,500	2.1	41
	Low forest	Glacial deposits	9,000	2.5	42
	<i>Metrosideros</i>	Volcanic ash	3,500	2.5	43
Tropical forest	Rain forest	Volcanic ash	8,620	2.3	44
Temperate grassland	<i>Chionochloa</i>	Glacial deposits	9,000	2.2	42
Temperate desert	Grassland	Alluvium	3,040	0.8	45

Unless actual data were included, calculated accumulation rates assume soil bulk density of 1.5 g cm^{-3} and that soil organic matter contains 50% organic carbon by mass. Rates do not include accumulations of undecomposed organic matter on the soil surface.

* Data corrected for 8 kg C m^{-2} in forest floor and a new estimate for soil age⁴⁶.

On a unit-area basis, the annual production of refractory humus substances in upland soils is greater than the deposition of carbon in deep-ocean sediments ($0.44 \text{ g C m}^{-2} \text{ yr}^{-1}$; ref. 12), but ocean sediments cover a greater area and are likely to accumulate and persist for longer periods of time. Organic matter does not accumulate in upland soils forever; eventually a steady-state profile content is achieved. In many forests the soil profile contains $\sim 10\text{--}20 \text{ kg C m}^{-2}$ (refs 1, 2), implying that $>6,000$ years are required to establish a stable distribution of carbon in the soil. At this point the production of humic compounds, in excess of decomposition, must equal the removal of these substances by erosion. Thus, estimates of the transport of organic carbon in rivers provide an upper limit for the production of refractory humic substances in the soil (some river transport is derived from fresh plant material that enters rivers directly). Recent estimates of the global transport of organic carbon in rivers ($0.4 \times 10^{15} \text{ g C yr}^{-1}$; ref. 13) are consistent with the estimate of an annual production of $0.40 \times 10^{15} \text{ g C yr}^{-1}$ of soil humus substances in steady-state conditions. Some of the organic carbon carried by rivers is degraded in the sea¹² and the remainder is deposited in sediments of estuaries and on the continental shelf¹⁴.

At the maximum extent of Pleistocene glaciation, $29.5 \times$

10^6 km^2 of the present land area was covered with ice¹⁵. This area now contains roughly $400 \times 10^{15} \text{ g C}$ or $\sim 25\%$ of the carbon contained in all soils of the world¹. The pool of carbon in deglaciated soils is consistent with accumulation rates of $<2.4 \text{ g C m}^{-2} \text{ yr}^{-1}$ for the last 10,000 years^{16,17}. The annual rate of storage in glaciated ecosystems ($0.04 \times 10^{15} \text{ g C yr}^{-1}$) is too small to have affected atmospheric CO_2 significantly during the Holocene³. It is difficult to see how this rate of storage might have changed enough during the past century to account for the $2\text{--}3.4 \times 10^{15} \text{ g C yr}^{-1}$ that is currently missing from the atmospheric inventory of CO_2 ¹⁸. Moreover, if the global climate becomes warmer, most soils are likely to become net sources of atmospheric CO_2 , because higher soil temperatures increase decomposition rates^{19–21}. Shifts in the distribution and extent of terrestrial vegetation will also produce changes in the storage of soil organic carbon. The areal extent of vegetation predicted by Emanuel *et al.*²² and the soil carbon contents of Post *et al.*² suggest a loss of $45.5 \times 10^{15} \text{ g C}$ from the soil carbon pool when the atmospheric CO_2 concentration is double the present value. In a similar analysis, Prentice and Fung³ suggest a slight increase in the soil carbon pool in the ecosystems of a warmer Earth.

Most soils show net losses of organic matter when they are converted to agricultural use⁵ but soil carbon contents increase

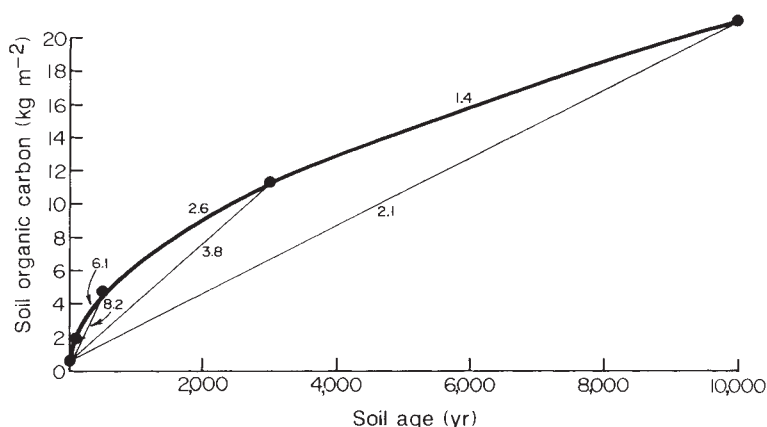


FIG. 1 The rate of carbon accumulation in New Zealand sand dunes²⁸. For any soil age, the calculated long-term rate of accumulation overestimates the actual rate of accumulation during the most recent interval.

when these lands are allowed to return to native vegetation^{23,24}. Soil organic matter may also be enhanced under some types of improved agricultural management²⁵. Between 1950 and 1959, $27 \times 10^4 \text{ km}^2$ of farmland was abandoned in the eastern United States²⁶. Using $30 \text{ g C m}^{-2} \text{ yr}^{-1}$ as an estimate of the rate of accumulation of soil organic matter in 40- to 50-year-old soils^{24,27,28} results in a storage of $0.008 \times 10^{15} \text{ g C yr}^{-1}$ in these soils—an insignificant flux in the global carbon cycle. Similarly, Delcourt and Harris²⁹ report that forests in the southeastern United States have served as a sink for $0.07 \times 10^{15} \text{ g C yr}^{-1}$ in biomass and soils in recent years. Over most other areas of the world, agricultural lands have expanded in recent years³⁰, offering no sink for atmospheric CO_2 .

These findings have several implications for understanding recent perturbations of the global carbon cycle that are leading to an increase in atmospheric CO_2 . First, the amount of carbon that can be stored in soils is likely to be only a small fraction of the annual release from fossil fuels ($5.3 \times 10^{15} \text{ g C yr}^{-1}$; ref. 18), even assuming a substantial increase in net primary production as a result of higher atmospheric CO_2 . If the terrestrial biosphere acts as a sink for atmospheric CO_2 , the organic carbon must be found in live-plant biomass and fresh, undecomposed plant litter on the soil surface^{3,31}. Second, recent estimates of the loss of soil organic matter when natural lands are cultivated indicate an annual net release of $0.8 \times 10^{15} \text{ g C}$ (refs 5, 30). Under human influence, the loss of humic materials from cultivated lands is greatly in excess of the rate of formation of humus in undisturbed lands, transforming the soil pool from a small sink to a large source of atmospheric CO_2 . □

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Evidence for chaotic fault interactions in the seismicity of the San Andreas fault and Nankai trough

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INTERACTIONS between fault segments are one of the main sources of complexity in the seismicity of active tectonic regions. Such interactions are likely to influence the spatial and temporal patterns of earthquakes. Here we examine the dynamical behaviour introduced by fault interactions using a simple spring-loaded, slider-block model with velocity-weakening friction. The model consists of two slider blocks coupled to each other and to a constant-velocity driver by elastic springs. For an asymmetric system in which the frictional forces on the two blocks are not equal, the solutions exhibit chaotic behaviour. The system's behaviour over a range of parameter values seems to be generally analogous to that of weakly coupled segments of an active fault. We see similarities between our model simulations and observed patterns of seismicity on the south central San Andreas fault in California and in the Nankai trough along the coast of south-western Japan.

Earthquakes in a zone of crustal deformation are a complex phenomenon. A standard procedure for determining whether a complex system exhibits chaotic behaviour is to study a low-order analogue model. A widely recognized analogue model for seismic faulting is a spring-loaded slider-block system (see refs 1–5, for example). This deterministic mechanical model incorporates the essential elements of the earthquake mechanism, including elastic storage of energy and friction-controlled stick and slip behaviour.

Dynamical instabilities associated with complicated friction laws are well known from studies using single-block models^{6–10}. Under certain conditions, the motion of a system subject to rate- and state-dependent friction laws is characterized by chaotic oscillations⁹. Analyses of the dynamics of a two-block model with spatial symmetry show spatially asymmetric events but no chaotic behaviour¹¹. The introduction of asymmetries, however, results in chaotic behaviour¹². Simulations of homogeneous many-slider systems approximating a continuous fault have shown size distributions that are consistent with the Gutenberg–Richter relation¹³.

Here we model interactions between two weakly coupled fault segments. Our two-block model is illustrated in Fig. 1 and the applicable equations of motion are

$$m_1 \ddot{y}_1 + (k_1 + k_c)y_1 - k_c y_2 = F_1 \quad (1)$$

$$m_2 \ddot{y}_2 + (k_2 + k_c)y_2 - k_c y_1 = F_2 \quad (2)$$

where F_1 and F_2 are the friction forces exerted on masses m_1 and m_2 , which are coupled to each other by a spring with spring constant k_c . The masses are coupled to the slider block by springs of constants k_1 and k_2 , which are extended by distances y_1 and y_2 ($\ddot{y}_i = d^2 y_i / dt^2$). We chose a velocity-weakening friction law of the form

$$F = \frac{F_0}{1 + |\dot{y}|/v_f} \quad (3)$$

where v_f is a reference velocity. The frictional forces decrease monotonically as the sliding velocity increases. Velocity-weakening phenomena have been widely observed in laboratory experiments (refs 14, 15, for example) and this simplified form of velocity-weakening has been used in previous model studies¹³.

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