scale ceramic superlattices. Other possible applications of the technique include mapping out the band structure and doping profile in semiconductor multiple quantum wells and studying the coherency of interfaces in superlattices by directly observing the interfacial registry of atoms in the materials.

REFERENCES AND NOTES

- C. R. Leavens and R. Taylor, Eds., *Interfaces, Quantum Wells, and Superlattices* (North Atlantic Treaty Organization Series B: Physics, vol. 179) (Plenum, New York, 1988).
- 2. L. Esaki, *IEEE J. Quantum Electron.* **QE-33**, 1611 (1986).
- C. Weisbuch and B. Vinter, *Quantum Semicon*ductor Structures (Academic Press, San Diego, CA, 1991).
- R. M. Fleming *et al.*, *J. Appl. Phys.* **51**, 357 (1980).
 D. B. McWhan, in *Synthetic Modulated Structures*, L. L. Chang and B. C. Giessen, Eds. (Academic Press, Orlando, FL, 1985), chap. 2.
- M. Shinn, L. Hultman, S. A. Barnett, *J. Mater. Res.* 7, 901 (1992).
- 7. G. Binnig and H. Rohrer, *IBM J. Res. Dev* **30**, 355 (1986).
- P. K. Hansma and J. Tersoff, J. Appl. Phys 61, R1 (1986).
- 9. J. Simmons, ibid. 34, 1793 (1963).
- S. M. Sze, *Physics of Semiconductor Devices* (Wiley, New York, ed. 2, 1981), chap. 1.
 F. Flores and N. Garcia, *Phys. Rev. B* 30, 2289
- (1984).
- 12. C. G. Slough et al., ibid. 42, 9255 (1990).
- 13. X. L. Wu et al., Science 248, 1211 (1990).
- D. Jeon, J. Kim, M. C. Gallagher, R. F. Willis, *ibid.* 256, 1662 (1992).
- 15. P. Muralt, Surface Sci. 181, 324 (1987).
- F. Osaka, I. Tanaka, T. Kato, Y. Katayama, Jpn. J. Appl. Phys. 27, L1193 (1988).

- 17. T. Kato, F. Osaka, I. Tanaka, *ibid* **28**, 1050 (1989). 18. J. M. Gomez-Rodriguez *et al.*, *Appl. Phys. Lett.*
- **56**, 36 (1990).
- 19. J. A. Dagata *et al.*, *ibid.* **59**, 3288 (1991). 20. J. A. Switzer, M. J. Shane, B. J. Phillips, *Science*
- **247**, 444 (1990).
- A. Guinier, X-ray Diffraction in Crystals, Imperfect Crystals, and Amorphous Bodies, translated by P Lorrain and D. Sainte-Marie Lorrain (Freeman, San Francisco, 1963).
- 22 R. J. Phillips, M. J. Shane, J A. Switzer, *J. Mater. Res.* **4**, 923 (1989)
- 23. J. A Switzer, J. Electrochem. Soc. 133, 722 (1986).
- 24 W. Mindt, ibid. 116, 1076 (1969).
- 25. We obtained STM images with a Nanoscope II (Digital Instruments) using Pt-Ir tips. For all samples, the tunneling current was 0.5 nA and the applied bias was 200 mV with the tip positive. The scan rate was 5 Hz or less, and the scan area was approximately the same as the final image area. Fast Fourier filtering was used to remove white noise from the image. The filtering did not change the shape of the apparent height profiles Samples were imaged with several different tips to ensure that the apparent height profiles did not result from tip sharpness.
- S. A. Chalmers *et al.* [*Appl. Phys. Lett.* 55, 2491 (1989)] have imaged tilted superlattices in the atomic force microscope by selectively etching one of the layers to improve contrast.
- 27 The superlattices were deposited from a stirred solution of 5 mM TINO₃ and 100 mM Pb(NO₃)₂ in 5 M NaOH A prelayer of Pb_{0.74}Tl_{0.26}O_{1.9} with a strong (210) texture was used to control the orientation of the superlattice. The potentials and currents were controlled with a potentiostat-gal-vanostat (EG&G model 273A).
- This work was supported in part by National Science Foundation grant DMR-9202872 and by Office of Naval Research grant N00014-91-J-1499.

20 July 1992; accepted 29 October 1992

Dynamics of Soil Carbon During Deglaciation of the Laurentide Ice Sheet

Jennifer W. Harden, Eric T. Sundquist, Robert F. Stallard, Robert K. Mark

Deglaciation of the Laurentide Ice Sheet in North America was accompanied by sequestration of organic carbon in newly exposed soils. The greatest rate of land exposure occurred around 12,000 to 8,000 years ago, and the greatest increase in the rate of carbon sequestration by soils occurred from 8,000 to 4,000 years ago. Sequestration of carbon in deglaciated peat lands continues today, and a steady state has not been reached. The natural rate of carbon sequestration in soils, however, is small relative to the rate of anthropogenic carbon dioxide production.

T errestrial organic C may be an important source (1, 2) or sink (3) for atmospheric CO_2 . It is also an integral component of C cycling over glacial cycles (4) and may "damp" glacial cycles through release and uptake of atmospheric CO_2 (5). The global mass of atmospheric CO_2 (740 Gt of C, 1 Gt = 10¹⁵ g) is sensitive to changes in the larger pool of terrestrial C (about 2100 Gt of C), of which about two-thirds is soil organic C. Evaluation of soil C sources and sinks is difficult because the dynamics of soil C storage and release are complex and still poorly understood. When the substrate is initially free of C, net C accumulation rates are high but typically decrease manyfold as the soil develops (1, 6–8). However, published inventories of soil C seldom include

SCIENCE • VOL. 258 • 18 DECEMBER 1992

information about soil age and development that is necessary to estimate fluxes.

Rates of soil C accumulation depend on dynamic interactions among landscape, substrate, vegetation, and climate. In this report, we examine the temporal and spatial dynamics of soil C accumulation in the region formerly occupied by the Laurentide Ice Sheet. We combine soil inventories with landscape histories to model soil C sequestration (9) for the region during the last 18,000 years.

We quantified the area of exposure of seven generalized soil units for the past 18,000 years by combining maps of ice retreat and soils (Figs. 1 and 2) in Canada and the northern United States (10-12).

For most soils, rates of net C accumulation are greatest during early stages of soil development on freshly exposed substrates. Studies from diverse climatic and geologic settings (1, 6) show that the rate of this accumulation in mineral soils diminishes after hundreds to thousands of years. Following the example of earlier models of soil organic C (7, 8, 13), we used a reservoir model that incorporates a zero-order input of C and a first-order rate of loss:

$$\frac{dC}{dt} = I - kC \tag{1}$$

where C is the mineral soil C mass per unit area of land surface, I is the rate of input, and k is the first-order rate constant for loss. The solution to this equation is:

$$C = I/k \times (1 - e^{-kt}) = C_e(1 - e^{-kt})$$
(2)

where *Ce* is the equilibrium C inventory for the mineral soil. We used data from representative soil chronosequences (Fig. 3) to fit this equation (14). Chronosequence data were not available for Mollisols, so we used data for Brunisols instead because the C contents in these two mature soils are similar (15, 16). Soils near the Rocky Mountains, Entisols and various other types, were not included because of their heterogeneity. The omission of this 0.4×10^{12} m² area from the total of 8×10^{12} m² likely resulted in an underestimate of sequestered C.

Equations 1 and 2 are not appropriate for organic soils (Histosols or peats). In these soils organic matter accumulates under water-saturated conditions that maintain plant production (input) as decomposition approaches zero. Long-term average rates of C accumulation in peats range from about 10 g m⁻² year⁻¹ (17, 18) to greater than 50 g m⁻² year⁻¹ (18–20), depending on factors that are only partly understood (18). For comparative purposes we used two rates, 10 and 30 g m⁻² year⁻¹, to model C accumulation in peat. Although constant rates for long periods are not documented anywhere (18), the zero-order model likely represents the dramatic differences from

J. W. Harden and R. K. Mark, U.S. Geological Survey, Menlo Park, CA 94025.

E. T. Sundquist, U.S. Geological Survey, Quissett Campus, Woods Hole, MA 02543.

R. F. Stallard, U.S. Geological Survey, Denver Federal Center, Lakewood, CO 80225.

Fig. 1. Map of modern soil types (11, 12) and ice retreat lines (10). Expansion of each Histosol (peat) area was modeled within the ages indicated based on maximum and minimum basal peat dates [extracted from (18) and other sources indicated]: (A) 9 to 6 ka [samples 15 to 19 in (19)], (B) 6 to 4 ka (estimated), (C) 7 to 5.3 ka [samples 48 to 51 in (21)], (D) 8 to 4 ka [samples 43 to 47 in (21)], (E) 8 to 3 ka [samples 1 to 4 in (19)], and (F) 6 to 4 ka. Canadian organic Fibrisols and some gleysolic soils are shown as Histosols. Podzolic soils are shown as Spodosols. Most Luvisols.



solonetzic, and some regosolic soils are shown as Entisols and other Chernozems shown as Mollisols. Numbers on ice retreat lines in thousands of years ago.

first-order decomposition of other soil types. The two rates encompass a reasonable range of spatial and temporal variation.

For each soil type i, we calculated the total amount of C sequestration at time t with the equation:

$$C_{i}(t) = \sum_{t_{0}=0}^{t} A_{i}(t_{0}) \times C_{i}(t-t_{0}) \quad (3)$$

where $A_i(t_0)$ is the incremental soil area that originated at time t_0 and $C_i(t - t_0)$ is the modeled C content of soil at age $(t - t_0)$. For most soil types, $A_i(t_0)$ was estimated from the times of ice retreat from land that is now covered by soil type *i* (Fig. 2). However, in many peat sections the basal



Fig. 2. Area of individual soil types exposed by ice retreat over the past 18,000 years, calculated from digitized maps shown in Fig. 1. Modern-day areal inventories of soil units are compiled in Table 1.

layer postdates local ice retreat by 2000 to 4000 years, and in some cases peat began forming after drainage of large glacial lakes (10, 18). We used basal dates for the peat lands (18, 19, 21) to calculate the timing of postglacial C sequestration by peats. We assumed that each peat land began to form at its oldest basal date and that the peat expanded at a constant rate to its modern mapped area until its youngest basal date (Fig. 1). The sums of Eq. 3 were determined for 100-year increments from 18,000 years ago (18 ka) to the present (Fig. 4 and Table 1).

There are many assumptions and simplifications in our approach. First, we assumed that recent climatic and environmental influences on modern soils are analogous to factors that influenced soils soon after local ice retreat. A more accurate treatment of

Fig. 3. Inventories of soil organic C in soil chronosequences (1). Each symbol represents one or more soil profiles of a given age; points of a chronosequence are connected with straight lines. Terms for fits of all data to Eq. 2 are shown in Table 1. Curve A, Alaskan Spodosols from Glacier Bay with a soil profile off scale at 27 kg m⁻², 1000 years (27) (scale for soils in Curve A only is in years). Curve B, Michigan Spodosols (26). Curve C, Baffin Island Brunisols (16, 34). Curve D, Baffin Island Cryosols (16, 34, 35). Curve E, Pennsylvania Alfisols (36). Where data were missing, we used a C concentration of 58% by weight for organic matter and 1.5 g cm⁻³ for bulk density of mineral soil. For Baffin Island data, age

this problem would account for ecosystem successions along ice margins, local fluctuations in water tables, and changes in temperature and rainfall. These factors are not well known, but their effects greatly complicate the interpretation of soil chronosequence data. Second, the chronosequence data are from well-drained soils, which typically store and generate less organic C than do soils in other settings (22). Third, we modeled soil C as a single homogeneous pool for each soil type. However, soil C is a heterogeneous mixture that includes both labile soil C (such as roots, litter, and soluble organics) and generally more abundant refractory components (such as claybound compounds with slow rates of replacement or turnover) (7, 8, 23). We restricted our study to the mineral soil horizons only. The relative contributions of labile and refractory C are uncertain, and few chronosequence data are available for several soil types. We did not separately consider soils over permafrost (24) and those in active erosional and depositional landscapes because of the paucity of data and maps. However, these areas are likely significant sites of C accumulation (20, 25). Most of the above assumptions probably lead to underestimates in our inventories and sequestration rates.

Several of these problems are illustrated by a comparison of calculations based on two different Spodosol chronosequences (Fig. 4 and Table 1). Older chronosequences (26) tend to exhibit low rates of C turnover, most likely because they contain high proportions of refractory C. Young chronosequences (27) probably include high proportions of labile organic matter, which imply more rapid turnover and response of sequestration rates to land-exposure rates. To examine the effect of turnover on sequestration we compared the deglaciation model using two different Spodosols. One used an older chronosequence from Michigan that has low inputs of C and the other used a young chronosequence



ranges of 100 to 200 years were assigned as 100 years, 700 to 1,000 years as 1,000 years, 8,000 to 10,000 years as 10,000 years, middle Foxe as 60,000 years, and early Foxe as 100,000 years.

SCIENCE • VOL. 258 • 18 DECEMBER 1992

REPORTS

from southeast Alaska that has high inputs. Both chronosequences have equilibrium C contents (C_e) that are comparable to an average of those in Spodosols of northern latitudes (15). The conspicuous differences in sequestration (Fig. 4) emphasize both the large uncertainties in our calculations and the need for a better understanding of soil C dynamics on local to regional scales.

To test the overall simulation, we compared (Table 1) our integrated totals of soil C with independent modern inventories in the deglaciated region. Although the uncertainties are large, the range of totals calculated from our model compares favorably with the ranges of independent inventories. On the basis of the data in Fig. 5, we estimate that 0.005 to 0.015 Gt of C per year was sequestered in the area of the Laurentide Ice Sheet during the Middle Holocene and that 0.015 to 0.035 Gt of C per year was sequestered during preindustrial times.

Extrapolation of these calculations to all deglaciated regions would require the additional assumption that deglaciation of the Laurentide Ice Sheet is representative. If Laurentide ages, distributions, and accumulation rates were similar to those in the remaining 80% of deglaciated land (28), then about 0.075 to 0.18 Gt of C per year would have been sequestered on deglaciated terrain globally during preindustrial times. In comparison, two global estimates of pre-industrial C accumulation in peats are 0.04 (29) and 0.076 (25) Gt per year.

Far broader areas of the land surface were profoundly altered during deglaciation. For example, large areas of nonglaciated land in the conterminous United States are covered by Late Pleistocene and Holocene deposits (30). These deposits represent erosional and depositional events and changes in terrestrial C budgets that are more com-



Fig. 4. Sensitivity of C sequestration to decomposition dynamics. Rates of C sequestration are from model rates of exposure of Spodosols (Fig. 2) and rates of organic-matter accumulation as determined in two contrasting soil-chronosequence studies at Glacier Bay, Alaska (27), and Michigan (26), respectively. Raw data are in Fig. 3 and model coefficients are in Table 1. Rapid turnover rates (low values of *k*) determined for Glacier Bay enable close coupling of C sequestration with exposure of land. When turnover rates are low, as in Michigan, C accumulation lags behind land exposure.

plex than simply a readjustment of ecosystems to a change in climate. However, one must consider the predominance of peats in current soil C sequestration and global inventories of peats. From these measures, it appears that natural postglacial soil C sequestration cannot account for the terrestrial sink of 2 to 5 Gt per year of C that is required to balance the modern annual anthropogenic CO₂ budget (3, 31).

Our results indicate that peats have dominated soil C sequestration since the Late Holocene, although they occupy only 12% of the land area. Although ice retreat was most rapid between 12 and 8 ka (Fig. 2), soil C sequestration increased most rapidly between 8 and 4 ka because of the time



Fig. 5. Model sequestration rates of C for the soil types shown in Figs. 1 and 2 for rates in Table 1. For the lower estimate we used rates of $10 \text{ g}^{-2} \text{ year}^{-1}$ of C for peats and the chronose-quence of (*26*) for Spodosols. For the upper estimate we used rates of $30 \text{ g}^{-2} \text{ year}^{-1}$ of C for peats and the chronose-quence of (*26*) for Spodosols. For the upper estimate we used rates of $30 \text{ g}^{-2} \text{ year}^{-1}$ of C for peats and the chronose-quence of (*27*) for Spodosols.

lag between ice retreat and peat formation (Fig. 5). The rapid increase in sequestration generally coincided with the end of the rapid increase in atmospheric CO_2 after deglaciation, as recorded in ice cores (32). The deglacial CO_2 increase likely resulted from interactions between the climate system and the marine C cycle (33). The deglacial CO_2 increase may have ended partly as a result of C sequestration by peats (28).

Deglaciation is accompanied by profound changes in terrestrial C cycling, changes that may persist long after the period of most dramatic climate change. Our calculations demonstrate that shifts between glacial and interglacial modes of terrestrial C cycling affected rates as well as inventories and that the terrestrial C cycle of preindustrial times may not have been at steady state. The lack of sufficient data that link soil dynamics to landscape dynamics seriously limits a clear understanding of the

Soil type	Area (m ² × 10 ¹²)	Average C inventory (15) (kg m ⁻² of C)	Parameter in Eq. 1			Model C sequestered	Modern
			<i>C</i> _ (kg m ^{−2} of C)	<i>I</i> (kg m ⁻² year ⁻¹ of C)	1/ <i>k</i> (years)	after deglaciation (Gt)	of soil C* (Gt)
Spodosols	2.59	16	11 (<i>27</i>) 12 (<i>26</i>)†	0.055 0.0018	200 6540	28 24	41
Cryosols	1.65	?	4.6 (34, 35, 16)	0.0043	1100	7.7	?
Histosols (peats)	1.03	_	_	0.010	_	68	170‡
				0.030	_	210	
Mollisols	0.989	9.1	11§ (<i>16, 34</i>)	0.0020	5500	9.7	9.0
Brunisols	0.672	?	11 (<i>16, 34</i>)	0.0020	5500	6.3	?
Alfisols and Inceptisols	0.473	6.6	4.7 (36)	0.00027	17,000	1.3	3.1
Entisols and other	0.705	—	—	_	—	_	—

Table 1. Dynamics and model sequestration of soil organic C for different soil types. C, carbon inventory; C_e , carbon inventory at equilibrium; k, decomposition coefficient; l, input rate; 1/k, turnover time.

*Area \times average C inventory for soil type. as 340 Gt peat; at 50%, C = 170 Gt (40). tWe used C inventories of Franzmeier and Whiteside and updated ages given by Barrett and Schaetzl (26). \$No chronosequence data available. We used data for Brunisols, based on similar inventories. relation between terrestrial C and atmospheric CO₂. Assessment of modern terrestrial inventories alone will not provide sufficient information to predict potential future sinks and sources for atmospheric CO_2 . One must also base such predictions on time-series data and process studies to determine the effects of landscape alterations, including those caused by human activities (2).

REFERENCES AND NOTES

- W. H. Schlesinger, *Nature* **348**, 232 (1990).
 R. A. Houghton, J. E. Hobbie, J. M. Melillo, *Ecol.* Monoar. 53, 235 (1983)
- 3. P. P. Tans, I. Y. Fung, T. Takahashi, Science 247, 1431 (1990).
- N. J. Shackleton, in The Fate of Fossil Fuel CO2 in the Oceans, N. R. Andersen and A. Malahoff, Eds. (Plenum, New York, 1977), pp. 401-427.
- J. N. Adams, H. Faure, L. Faure-Denard, J. M. 5 McGlade, F. I. Woodward, Nature 348, 711 (1990).
- 6. H. Jenny, The Soil Resource: Origin and Behavior (Springer-Verlag, New York, 1980).
- 7 D. S. Jenkinson, Philos. Trans. R. Soc. London 329, 361 (1990)
- W. J. Parton, D. S. Schimel, C. V. Cole, D. S. 8 Ojima, Soil Sci. Soc. Am. J. 51, 1173 (1987).
- We use the term accumulation for site-scale data with units of grams per square meter per year and the term sequestration for areally extrapolated data with units of grams per year. 10. A. S. Dyke and V. K. Prest, *Geog. Phys. Quat.* **41**,
- 237 (1987).
- National Átlas of Canada (Canada Department of Energy and Resources: Surveys and Mapping Branch, Ottawa, 1973).
- Soil Conservation Service, in National Atlas of the 12 United States (U.S. Geological Survey, Reston, VA, 1987), pp. 42–43.
- 13. H. Jenny, S. P. Gessel, F. T. Bingham, Soil Sci. 68, 419 (1949).
- 14. A soil chronosequence is a series of soil profiles from sites that have similar factors such as land use, climate, vegetation, slope, and original material but differ in age of the landform (37). Each chronosequence in our study was fit to a curve of the form of Eqs. 1 and 2 with the use of the Marquardt-Levenberg nonlinear regression implementation of Sigma Plot Version 4.0 (Jandel Scientific, Corte Madera, CA, 1990). Chronose-quences were selected where geologic ages of deposits represented age of geomorphic surface and inception of soil development. Chronosequence data for soils other than peats apply to mineral soil below organic leaf-litter layers. 15. J. M. Kimble, H. Eswaran, T. Cook, in International
- Congress of Soil Science (International Congress of Soil Science, Kyoto, 1990), pp. 248-253.
- 16. P. W. Birkeland, Arct. Alp. Res. 10, 733 (1978)
- K. R. Everett, ibid. 3, 1 (1971). 17
- C. Tarnocai, in Global Change, Canadian Wet-18. lands Study, H. I. Schiff and L. A. Barrie, Eds. (Canadian Institute for Research in Atmospheric Chemistry, York University, North York, U.K., 1988), pp. 21–25.
- 19. L. Ovenden, Quat. Res. 33, 377 (1990).
- 20. W. D. Billings, *Quat. Sci. Rev.* **6**, 165 (1987). 21. S. C. Zoltai and D. H. Vitt, *Quat. Res.* **33**, 231 (1990)
- D. Schimel, M. A. Stillwell, R. G. Woodmansee, 22. Ecology 66, 276 (1985)
- 23. S. Trumbore, *Ecol. Appl.*, in press. 24. J. Brown, *Soil Sci. Soc. Am. Proc.* **31**, 686 (1967).
- E. Gorham, Ecol. Appl. 1, 182 (1991). 25.
- D. P. Franzmeier and E. P. Whiteside, Mich. Agric Exp. Stn. Q. Bull. 46, 1 (1963); L. R. Barrett and R. J. Schaetzl, Can. J. Soil Sci., in press.
- 27. F. C. Ugolini, in Soil Development and Ecological Succession in a Deglaciated Area of Muir Inlet,

Southeast Alaska, R. P. Goldthwaite, F. Loewe, F. C. Ugolini, H. F. Decker, D. M. DeLong, Eds. (Ohio State University, Columbus, 1974), pp. 29-56.

- 28. The timing of peat land expansion in northern Asia, around 5 ka or later (25), and the size of peat land in north central Asia (38) were comparable to those of the Laurentide region. The Laurentide Ice Sheet represents about 20% of deglaciated land (39, 40). Peats are the most important C sink. The calculation implies about 3 \times 10¹² m² of peat land in deglaciated regions outside of the Laurentide region. Total world peat inventories are 3×10^{12} to 4×10^{12} m² (25), so our extrapolation overestimates C sequestration by deglacial peats. The exclusion from our inventories of soils formed on continuous permafrost helps to compensate for this overestimate. Carbon is sequestered in many of these soils because of the lack of drainage and limited decomposition. Field studies and better maps would help improve this extrapolation.
- 29. W. H. Schlesinger, Biogeochemistry, An Analysis of Global Change (Academic Press, San Diego, CA. 1992). Preindustrial rates include those of peats and low rates for accumulation for non-peat soils
- 30. S. C. Porter, in Americans Before Columbus: Ice Age Origins, R. C. Carlisle, Ed. (Department of Anthropology, Pittsburgh University, Pittsburgh, PA, 1988), pp. 1-24.

- 31. J. L. Sarmiento and E. Sundquist, Nature 356, 589 (1992)
- A. Neftel, H. Oeschger, J. Schwander, R. Stauffer, R. Zumbrunn, *ibid*. **295**, 220 (1982). 32.
- 33. W. S. Broecker and T. H. Peng, Tracers in the Sea (Lamont Doherty Geological Observatory, Palisades, NY, 1982).
- J. G. Bockheim, Arct. Alp. Res. 11, 289 (1979). 34
- L. J. Evans and B. H. Cameron, Can. J. Soil Sci. 35 59, 203 (1979).
- E. J. Ciolkosz, in Quaternary Deposits and Soils of the Central Susquehanna Valley of Pennsylvania, D. Marchand, E. J. Ciolkosz, M. E. Bucek, G. H. Crowl, Eds. (Agronomy Department, Pennsylva-nia State University, University Park, 1978).
- H. Jenny, Factors of Soil Formation (McGraw-Hill, 37. New York, 1941).
- 38 Food and Agriculture Organization, United Nations Educational, Scientific, and Cultural Organization, Soil Map of the World (1978).
- 39 K. C. Prentice and I. Y. Fung, Nature 346, 48 (1990).
- G. H. Denton and T. Hughes, The Last Great Ice 40. Sheets (Wiley-Interscience, New York, 1981).
- 41. L. E. Jackson, in Quaternary Geology of Greenland and Canada, R. J. Fulton, Ed. (Geological Survey of Canada, Ottawa, 1989), pp. 667-697.
- We thank P. Showalter for work on Geologic 42 Informations Systems and W. Schlesinger, S Trumbore, and C. Tarnocai for reviews

8 July 1992; accepted 2 November 1992

Evidence from the Lamarck Granodiorite for Rapid Late Cretaceous Crust Formation in California

Drew S. Coleman,*† Thomas P. Frost, Allen F. Glazner

Strontium and neodymium isotopic data for rocks from the voluminous 90-million-year-old Lamarck intrusive suite in the Sierra Nevada batholith, California, show little variation across a compositional range from gabbro to granite. Data for three different gabbro intrusions within the suite are identical within analytical error and are consistent with derivation from an enriched mantle source. Recognition of local involvement of enriched mantle during generation of the Sierran batholith modifies estimates of crustal growth rates in the United States. These data indicate that parts of the Sierra Nevada batholith may consist almost entirely of juvenile crust added during Cretaceous magmatism.

Present models for the origin of the Sierra Nevada batholith developed from the strontium isotope studies of Kistler and Peterman (1) and indicate that the isotope systematics of the batholith can be reproduced through mixing between depleted mantle and Proterozoic upper crust (2-5), or by mixing of crustally contaminated mafic magmas and siliceous melts of Proterozoic lower crust (6). These models assume that the regular east/west change in the isotopic composition of granitic rocks in the batholith reflects an eastward increase in the contribution of old continen-

tTo whom correspondence should be addressed.

SCIENCE • VOL. 258 • 18 DECEMBER 1992

tal crust. In particular, the ${}^{87}\text{Sr}/{}^{86}\text{Sr} = 0.706$ isopleth (1), or the ${}^{87}\text{Sr}/{}^{86}\text{Sr} = 0.708$, $\varepsilon_{\text{Nd}} =$ -7 isopleths (3, 7), commonly are interpreted to reveal the approximate western limit of the continental crust under the United States. DePaolo et al. (8) reached similar conclusions regarding the origin of the batholith using Nd model ages for granitic rocks.

All of these models assume a depleted composition for the mantle end member. However, isotope studies in the Basin and Range province of the western United States suggest that parts of eastern California, Nevada, and Utah were underlain by enriched lithospheric mantle during Cenozoic (9–13) and Mesozoic (14) magmatism. Limited data from mantle xenoliths and alkalic lavas also suggest the presence of an enriched mantle component as far west as the central Sierra Nevada [Fig. 1 (15, 16)]. If an enriched mantle was involved in Sierra Nevada batholith formation, estimates of crust for-

D. S. Coleman and A. F. Glazner, Department of Geology, University of North Carolina, Chapel Hill, NC 27599

T. P. Frost, U.S. Geological Survey, West 904 Riverside Avenue Spokane, WA 99201

^{*}Present address: Massachusetts Institute of Technology, Department of Earth, Atmospheric, and Planetary Sciences, 80 Ames Street 54-1016, Cambridge, MA 02139