chemical lifetime supports the hypothesis[2] of a negative feedback in volcanic-induced climate perturbation, by limiting the amount of particulate H_2SO_4 present in the stratosphere at a given time.

Oxidation of volcanic SO_2 seems to be a very slow process in the stratosphere and the SO_2 chemical lifetime is increased to several months after volcanic eruptions. The volcanic SO_2 could then be oxidized far away from the eruption site and may produce a large depletion of the H_2O_2 reservoir when this material is reinjected into the troposphere.

This peroxide depletion seems to be taking place at least on a regional scale following both high-latitude and low-latitude volcanic events. Because of the very high solubility of peroxide in water, it is very likely that this depletion is parallelled by a depletion of gas-phase peroxide. Considering that the HO_2 radical is the main source of H_2O_2 in the tropospheres[5], this may also affect the concentration or the lifetime of this reactive compound. Because of the importance of the HO_2 radical in ozone chemistry, volcanic eruptions are then likely to produce a short-term variation in both nitrogen and odd-oxygen chemistry[7]. More work will be necessary to quantify the impact on other chemical cycles in the atmosphere. □

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The sensitivity of terrestrial carbon storage to climate change

Katharine C. Prentice & Inez Y. Fung

NASA, Goddard Space Flight Center, Institute for Space Studies, New York, New York 10025, USA

THE high correlation found between atmospheric CO_2 concentration and temperature during the past 160,000 years[1,2] implicates CO_2 as an important driving force behind changes in palaeoclimate. Changes in biological processes and circulation patterns in the oceans are believed to drive glacial-to-interglacial changes in CO_2 and thus in climate. Here we estimate the role of the terrestrial biosphere in controlling atmospheric CO_2 levels during climate perturbations. By considering the coupling between vegetation distributions and climate, we perform simulations to calculate the global geographical distribution of vegetation during the last glacial maximum, 18 kyr ago. The known changes in sea level at this time, together with the simulated climate-driven spatial arrangement of vegetation, result in a mass transfer of carbon from the terrestrial biosphere to the atmosphere ranging from 30×10^9 tonnes (30 Gt, corresponding to 15 p.p.m. CO_2) to −50 Gt (25 p.p.m.). Thus, although the biosphere may have contributed to the decrease in atmospheric CO_2 of 80 p.p.m. known to have occurred at 18 kyr (refs 1, 2), it does not seem to have been a dominant factor. For simulations run with twice the present-day CO_2 levels, strong negative feedbacks appear which remove 235 Gt of carbon (128 p.p.m. CO_2) from the atmosphere.

Bioclimate schemes relate the distribution of climates and the distribution of vegetation types, classifying climates according to their thermal and moisture attributes. The climate classes are named for vegetation types with which they have been associated. This association of vegetation and climate permits the simulation of vegetation distributions for different climates.

In an earlier study[3], the contemporary distribution of global vegetation was simulated by using a refined version of the Holdridge Life Zone Scheme[4] to classify observed 30-year mean annual temperature and precipitation[5]. The climate classes were grouped and named for the vegetation types with which they have greatest geographic coincidence[6]. Unlike earlier bioclimate predictions[7–10], resulting climate classes were named for present-day vegetation types defined by the internationally recognized UNESCO classification scheme[11] rather than for the independent and generalized vegetation groups used by the bioclimate schemes. Quantitative evaluation of the simulated distribution is thus possible because a consistent vegetation terminology is used for both observed and simulated vegetation. At $1° \times 1°$ resolution for the globe, 77% of the simulated landscape, classified as 29 vegetation types, spatially corresponded with observed vegetation. The misclassifications over 23% of the land surface generally involve climatically similar vegetation types which also have similar carbon attributes.

For the study reported here, the 29 vegetation types are combined into 14 types to reflect more generalized units of carbon storage. We apply distribution relationships developed between the 14 observed UNESCO types and the observed climatology to perturbed climates. As well as today's climates, climates at 18 kyr BP (before present) and climates for doubled CO_2 levels ($2 \times CO_2$) have been simulated using the general circulation model (GCM) at the Goddard Institute for Space Studies[12]. Temperature and precipitation, simulated at a $4° \times 5°$ resolution, are linearly interpolated to $2\frac{1}{2}°$. Both the $2\frac{1}{2}°$ observed and simulated distributions are then regridded to 1° without further interpolation beyond that necessary for the $\frac{1}{2}°$ registration. Annual temperature and precipitation used here are calculated for each 1° gridbox by adding the difference between simulated perturbed and simulated present climates to observed 30-year mean climatologies[5]. In this way, the interpretation of bioclimate patterns is rooted in observed climatology. Land area changes as 1°-gridded surface areas emerge or submerge with altered sea level. The sea level at 18 kyr is lowered by 130 and 100 m to bracket a best estimate of 121 m (ref. 13), and the sea level for the $2 \times CO_2$ simulation is raised by 2 m, a likely upper estimate[14].

To evaluate the possible impact of the perturbed biosphere on the carbon cycle, we calculate the sizes of the carbon reser-

TABLE 1 Carbon densities and areal extents for 14 vegetation types

Vegetation type	Biomass (kg C m^{-2}) ref. 16	Biomass (kg C m^{-2}) ref. 15	Soil carbon (kg C m^{-2}) ref. 19*	Soil carbon (kg C m^{-2}) refs 17, 18	Present area (10^{12} m^2)	18 kyr BP (−130 m) (% change)	18 kyr BP (−100 m) (% change)	$2\times CO_2$ (+2 m) (% change)
1	15	18	10	10	19	20	17	75
2	5	5	7	8	6	−23	−31	−42
3	3	2	6	12	5	−40	−42	35
4	1	0	6	7	30	2	0	−22
5	0	0	3	1	14	−91	−91	−62
6	3	1	11	20	2	−7	−8	−17
7	3	3	8	29	2	78	76	4
8	10	11	10	10	12	32	31	40
9	9	9	14	18	16	−16	−18	5
10	6	6	12	18	6	−21	−22	−66
11	1	0	17	13	7	2	1	−63
12	0	0	0	0	3	1,223	1,171	0
13	8	13	8	11	2	170	163	−31
14	6	11	7	8	9	−89	−91	21
Global (Gt C)	748	834	1,143	1,313	132	152×10^{12} m^2	148×10^{12} m^2	132×10^{12} m^2

The 14 simulated types are listed in column 1. They are: (1) tropical rain forest, (2) drought-deciduous woodland, (3) savanna, (4) arid grassland and shrubland, (5) desert, (6) mesic grassland, (7) Mediterranean forest and woodland, (8) cold-deciduous broad-leaved forest and woodland, (9) cold-deciduous needle-leaved forest and woodland, (10) evergreen needle-leaved forest and woodland, (11) tundra, (12) polar desert and ice, (13) temperate evergreen seasonal broad-leaved forest, (14) drought-deciduous and drought-seasonal broad-leaved forest. Carbon densities (columns 2–5) are based on published figures. They are constant for all climates. Areal extents of the 14 vegetation types (column 6) were bioclimatically simulated using observed climatology[4]. Simulated changes in areal extents (columns 7–9) are expressed as a percentage of present-day area (excluding Antarctica).

* Soil carbon is estimated directly from climate and averaged over the area of each vegetation type simulated from observed climate.

voirs in vegetation and soils at 18 kyr, for $2\times CO_2$ and for the present climate. Biomass estimates for vegetation types are taken from Whittaker and Likens[15] as well as from Olson *et al.*[16]. Soil carbon densities for their associated soils types are obtained by combining estimates of Ajtay *et al.*[17] and Schlesinger[18]. These carbon densities are geographic extrapolations of field measurements. The distribution of soil carbon densities is also estimated directly from climate using Post *et al.*'s superposition of isolines of soil carbon densities on the Holdridge climate space[19].

Over 75% and 60% of the present vegetated landscape was altered by the 18 kyr BP and $2\times CO_2$ simulated climates respectively. During the 18 kyr BP simulation, land temperatures dropped ~4.4 °C in the Southern Hemisphere and 8.5 °C in the Northern Hemisphere. Although the global landscape was more arid, eastern and southern coasts as well as present arid regions were wetter than today. The 100- and 130-m drop in sea level increased the land area by 12% and 15%. Regions of extreme climates (deserts, tropical rainforests and ice) show greatest sensitivity to climate change.

In this study, the biospheric carbon changes only because the distribution of vegetation types changes. The dynamics involved in both the change in the distribution of vegetation and changes in carbon density in response to changing climate are not considered. This analysis is a comparison of static snapshots of potential vegetation existing for today's, for the 18 kyr BP and for a $2\times CO_2$, climate.

Table 1 lists carbon densities for the 14 vegetation types (columns 2–5) as well as their present land areas (column 6) and percentage change in their areas associated with perturbed climates. (Numbers in the discussion refer to Olson *et al.*[16] and Post *et al.*[19] unless otherwise stated.) During the 18 kyr BP (−130 m) simulation, the advance of ice sheets claimed 25×10^{12} m^2 of present land, significantly reducing the extent of boreal vegetation. The areas of needle-leaved boreal forests (types 9 and 10) were reduced by 17% despite a 5% expansion on newly emerged land. The descent of the ice sheets and other effects of unfavourable climate as well as sea-level changes reduce the carbon stored in the 18 kyr BP boreal forests (type 9 and 10) by 80 Gt (30 Gt as biomass carbon and 50 Gt as soil carbon).

During the 18 kyr BP simulation, a large conversion of carbon occurred in the present land area of the equatorial region and in the Southern Hemisphere. Temperate and tropical broad-leaved forests (types 7, 8, 1, and 13) expanded into the cooler and wetter subtropical regions displacing subtropical broad-leaved forests and savannas (types 2, 14 and 3) which exist in climates having more strongly seasonal precipitation. The areal expansion of the less drought-seasonal forests (types 7, 8, 1 and 13) on existing and new land increases their carbon storage by 256 Gt, which is about equally divided between biomass and soil and also equally divided between new and existing land. The areas of the more drought-seasonal vegetation (types 2, 14 and 6) were reduced by 57% shrinking their carbon pool by 136 Gt (57 Gt carbon in biomass and 79 Gt as soil carbon).

Because of increased precipitation in arid regions during the 18 kyr BP simulations the area of deserts is reduced by 91%. Although 18 kyr BP desert decreases by 12×10^{12} m^2, this large area reduction represents only a 45 Gt decrease in the desert carbon pool—almost entirely as soil carbon.

Although the simulated biosphere responds dramatically to 18 kyr BP changes in climate, large regional changes in carbon reservoirs and changes in soil and biomass carbon offset one another in the global carbon budget. Global estimates of changes in carbon reservoirs from present to perturbed climates are listed in Table 2. Estimates of carbon-reservoir changes in Table 2 are broken into those effected by changes in both climate and sea level and those associated with changes in climate only. For 18 kyr BP climate with a 130 m drop in sea level, the 6% (43 Gt) gain in biomass carbon is offset by the 4% (43 Gt) decline in the soil carbon pool. Net change for the amount of carbon held in the 18 kyr BP (−130 m) biosphere is approximately zero. Like the changes in the biomass and soil reservoir, the changes in the global carbon reservoir associated with climate are also compensated by the effects of sea-level change on the biosphere. The direct effects of unfavourable 18 kyr BP climate reduce biomass carbon by 7% and soil carbon by 13%. This 200-Gt reduction of terrestrial biospheric carbon is offset by the development of vegetation and soil on new land. The 20×10^{12} m^2 of newly emerged land (Table 1) increases the carbon in the present terrestrial biomass by 13% and soil carbon by 9%. The amount and location of emerged land can account for a large increase in biospheric carbon. In this experiment, the biospheric

TABLE 2 Global terrestrial carbon reservoir changes (%) relative to today

Climate	Reservoir	Present-day carbon reservoir (Gt C)	Percent change in carbon reservoir: change in sea level (m) −130	−100	0*	+2
18 kyr BP	biomass					
	ref. 16	748	6	3	−7	—
	ref. 15	834	5	2	−8	—
	soil carbon					
	ref. 19	1,143	−4	−6	−13	—
	refs 17, 18	1,313	−1	−3	−10	—
$2\times CO_2$	biomass					
	ref. 16	748	—	—	31	31
	ref. 15	834	—	—	35	35
	soil carbon					
	ref. 19	1,143	—	—	1	1
	refs 17, 18	1,313	—	—	3	3

* Reservoir changes brought about by direct climate effects only

carbon supported on new land is as great as the changes in the biospheric carbon resulting directly from climate.

Similar changes in biospheric carbon reservoirs occurred in the 18 kyr BP (−100 m) simulation but the changes were affected by the smaller increase in land area associated with the 100 m drop in sea level. Biomass carbon storage increases by 3% (25 Gt) relative to the present reservoir whereas soils lose 6% (65 Gt). The terrestrial biosphere in this experiment loses 40 Gt of carbon.

The land climate for the $2\times CO_2$ simulation is an average of 5.4 °C warmer and is 17.5 mm d^{-1} wetter than present climate. Areas of tropical rainforests increase by 75% (Table 1) increasing their biomass and soil carbon reservoirs by 215 Gt and 130 Gt of carbon respectively. Desert (type 5) and semi-desert (type 4) regions are reduced by 60% and 20%, reducing the carbon stored in arid lands by 75 Gt. Cold-deciduous broad-leaved (type 8) forests expand northward in the Northern Hemisphere in the warmer, wetter $2\times CO_2$ climate adding 50 Gt carbon to their soil reservoirs and 60 Gt to their biomass. Areal decrease of boreal evergreen needleleaved forests (type 10) and tundra (type 11) reduces their soil carbon pool by 145 Gt.

The global terrestrial biosphere responds to the $2\times CO_2$ climate with a 31% increase in biomass a 1% increase in soil carbon. This 235-Gt carbon expansion of the terrestrial biosphere is due entirely to the change in climate (the sea level rises 2 m).

Although the spatial rearrangement of vegetation in response to perturbed climates is significant, the magnitude and direction of the net change in carbon storage remain unclear. Many caveats must accompany the resulting global carbon budgets. (1) At present, only 77% of the observed vegetated landscape can be replicated because we do not know how to express climate to define best the structural/functional attributes of vegetation. In addition, the secondary effects of soils, topography, humans and other environmental factors on the vegetation distribution are not accounted for in this study. Because misclassification of vegetation most often occurs between climatically similar vegetation types having similar carbon-storage attributes, the study's results are not highly sensitive to small errors in classification. Results are most sensitive to misclassifications in areas having a strong gradient in carbon density (for example, along a north–south transect of soil through ice, tundra and boreal forests). (2) We assume relationships now existing between climate classes and vegetation types hold for perturbed climates. In fact vegetation types existing under perturbed climates may differ in structure and function from those existing today. We also assume that vegetation is in equilibrium with climate — a condition which is never entirely met in a constantly changing climate. The assumption of steady state becomes less correct for the more slowly changing soil-carbon reservoir over what is believed to be the relatively short time period of an approaching $2\times CO_2$ climate. (3) The estimation and extrapolation of carbon densities for the geographically extensive and diverse vegetation types requires more field measurements than are presently available. Sensitivity of global terrestrial carbon storage to small errors in the carbon-storage description (Table 1) increases with the area of the vegetation type. Such errors are likely to increase with the density of carbon storage, with increasing percentage of storage in the soil (where carbon density is more difficult to measure), with diversity of carbon storage within vegetation type (making extrapolation of field measurements difficult) and inversely with the frequency of the field sampling of carbon density. The discrepancies between the minimum terrestrial carbon storage[16,19] shown in Table 2 and the maximum values[15,17,18] for a single-climate scenario are generally more than double the storage differences between climate scenarios. (4) Uncertainties and errors in GCM simulations, especially of precipitation affect the distribution of vegetation and terrestrial carbon. (5) At the GCM level of generalization and structural homogenization of vegetation types, issues of CO_2 fertilization and the functional differences of C_3 and C_4 plants cannot be addressed. (6) The 100-m and 130-m drops in sea level in the 18 kyr BP simulations are responsible for a 40-Gt difference in the amount of carbon stored in the biosphere. (7) Estimates of the changes in the net global biospheric carbon are obtained by summing large numbers often having opposite signs. Because there is a large margin of error in each component number, an accurate net change in the global terrestrial biosphere is difficult to obtain. Some generalizations, however, are possible.

In our study, both biomass and soil carbon pools during the 18 kyr BP (−130 m) simulation decreased by 200 Gt carbon on the present land area, or ~10% of the total present land reservoir. The decrease is countered by the substantial increase of carbon stored on the newly emerged land. The magnitude of this compensation is difficult to ascertain because of the uncertainties in the sea-level change and lack of understanding of the biosphere on the new landscape. By assuming a sea-level drop of 130 m and applying today's terrestrial carbon relationships, we obtain an increase in carbon storage of 200 Gt on new land, exactly compensating for the loss owing to unfavourable climate. Although our best estimate of the mass transfer of carbon between the biosphere and the atmosphere for the 18 kyr BP simulation is approximately zero, Tables 1 and 2 show possible biospheric changes ranging from an increase of 30 Gt carbon for −130 m (refs 16–18) to a decrease of 50 Gt carbon for the −100-m experiment[15,19]. Although biospheric changes do not dominate the 82-p.p.m. (165 Gt carbon) difference in atmospheric CO_2 between the Holocene and 18 kyr BP found in ice cores, their contribution may not be negligible because of uncertainties in calculating the components of large compensating factors in the biosphere–atmosphere carbon flux.

In our $2\times CO_2$ simulation, the change in sea level (+2 m) has a negligible effect on global land area. In this experiment, a terrestrial biosphere in equilibrium with a $2\times CO_2$ climate has an increased carbon storage ranging from 338 Gt (refs 15, 17, 18) to 245 Gt (refs 16, 19), or 169 p.p.m. to 122 p.p.m. This represents a potentially strong negative feedback on the atmospheric CO_2 increase.

In both cold and warm climates, the redistribution of vegetation results in carbon transfers from soil to biomass and from boreal to tropical regions. The present climate seems to be unique in its capacity to support such extensive carbon-rich boreal soils and such extensive carbon-poor arid lands.

These sensitivity experiments explore the influence the terrestrial biosphere might have had in the glacial–interglacial changes of atmospheric CO_2 as well as the possible carbon-storage role in a $2\times CO_2$ climate. We hope that this research can provide a framework to analyse and extrapolate terrestrial response to global climate changes. □

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Temperature measurements during initiation and growth of a black smoker chimney

M. K. Tivey*, L. O. Olson†, V. W. Miller† & R. D. Light†

* Woods Hole Oceanographic Institution, Woods Hole, Massachusetts 02543, USA
† Applied Physics Laboratory, University of Washington, Seattle, Washington 98105, USA

BLACK smoker chimneys are hollow spires formed by mineral deposition at sea-floor hydrothermal vent sites. They grow in two stages: formation of a sulphate-dominated wall (stage I), followed by precipitation of sulphide minerals on the inner side, and in pore spaces, of the wall (stage II)[1,2]. Here we present the results of an *in situ* 46-day experiment which allows direct observation of stage I growth processes by monitoring wall and fluid temperatures. The position and thickness of stage I chimney walls can change on short timescales, and are controlled by the dynamics of fluid flow both during[3] and after initial wall emplacement. The temperature in the main flow remained stable at 353±2 °C throughout the experiment, including the period during which the chimney wall was constructed. Six thermocouples, however, recorded maximum temperatures between 365 and 405 °C, which are significantly higher than most exit temperatures documented previously[4]. These latter observations bear on the question of the maximum temperatures attainable in sea-floor hydrothermal systems (refs 4–9, J. R. Delaney, manuscript in preparation).

In July 1988 an instrument equipped with an array of thermocouples housed within a test collar was deployed in the Endeavour hydrothermal vent field (47° 57′ N) on the Juan de Fuca Ridge[10,11]. The collar (Fig. 1*a*) was set over the opening of a black smoker vent and remained in place for 46 days. Temperatures (Fig. 2) were recorded every 2 min by 10 thermocouples inside the collar, 2 thermocouples outside the collar, and a platinum resistance temperature device (RTD) on the instrument platform, which was moored 1 m from the collar. The presence of the collar and thermocouples probably influenced fluid flow and mineral deposition to some extent, but some disturbance is unavoidable.

During the first 52 hours of the experiment, before chimney growth, the thermal structure of the hydrothermal plume was documented. Temperatures at TC02 (thermocouple 2), TC03 and TC09 increased rapidly to, and remained stable at, 353± 2 °C. Temperatures increased more slowly to maxima of 353±

TABLE 1 Pressure difference across chimney wall

Diameter (m)	Volume flow rate ($m^3 s^{-1}$)	Pressure difference (Pa)
0.016	0.00039	0
	0.00069	6,700
0.02	0.00039	−2,200
	0.00069	0
0.025	0.00039	−2,900
	0.00069	−2,200

Pressure differences calculated using equation (2), using $y=1$ m and $f=0.04$ (ref. 3); densities for 2 °C sea water and 350 °C hydrothermal fluid are 1,003 and 674 kg m^{-3} (ref. 12), respectively.

2 °C at TC05, TC06 and TC11, 365 °C at TC08, 375 °C at TC10, 381 °C at TC07, and 405 °C at TC04. The three highest temperatures are on the boiling curve (375 °C) or in the two-phase region for a seawater-salinity fluid at 220 bars[6,12]. Excursions to ⩾375 °C lasted for 90 min at TC04, 42 min at TC07 and 4 min at TC10, and were generally coincident but not simultaneous. In 1984 temperatures of 370–>400 °C and of >400 °C were measured at this site (Endeavour hydrothermal field; refs 6, 7, J. R. Delaney, manuscript in preparation) but their reliability has been disputed because these values were not reproduced on subsequent visits (temperatures were lower, but the same individual vents were not occupied) (ref. 4, J. R. Delaney, manuscript in preparation). In 1988 a temperature of 375 °C was measured 2 km to the north (ref. 5, J. R. Delaney, manuscript in preparation). Nearly all other published exit temperatures are <350±5 °C (ref. 4), including the temperature of the main flow measured in this experiment. Experimental reproduction of hydrothermal fluid compositions, however, has only been accomplished at temperatures >385 °C (refs 8, 9). Our high-temperature measurements seem to be reliable because they were all made adjacent to one another (Fig. 1*a*), and all thermocouples recorded temperatures within 2–4 °C of one another (within the error limits, ±2 °C) during deployment and recovery, and seem to have performed consistently throughout all other portions of the experiment. On the other hand, thermocouple TC02, only 4 cm from TC04, recorded a stable temperature of 353±2 °C for the duration of the experiment, including the period during which TC04 recorded 405 °C, and the thermocouples that measured anomalously high values did not record stable temperatures or temperatures similar to one another (Fig. 1*b*). The simplest explanation that accounts for all measurements is that there were two sources, one that was very stable at 353 °C, and a second, hotter source which was responsible for the >355 °C temperatures (J. R. Delaney, manuscript in preparation).

The anomalously high temperatures were recorded only during a short time following emplacement of the collar. Within 7 hours, temperatures at TC04, TC07, TC08 and TC10 became relatively stable at 369, 361, 358 and 357 °C respectively, and drifted to 363, 356, 354 and 354 °C over the next 45 hours. During this time all other thermocouples in the collar measured stable temperatures of 353±2 °C (Fig. 1*c*, Fig. 2*a*). The initial rapid increases to 353±2 °C at TC02, TC03 and TC09 indicate that hot fluid was directed against one side of the test collar (Fig. 1). Progressive increases in temperatures at the other thermocouples to >350 °C probably reflect decreases in leakage of sea water into the collar as mineral precipitation at the collar base sealed openings.

During the third day of the experiment, temperatures at all thermocouples except TC02 and TC03 began to change rapidly (Fig. 2*b*), indicating the sudden onset of seawater entrainment over the top of the collar (Fig. 1*d*). The cause of entrainment is unclear but possibilities include development of an instability in the turbulent plume, or a sudden decrease in flow rate. The