

CAN CLIMATE VARIABILITY CONTRIBUTE TO THE "MISSING" CO₂ SINK?

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Abstract. The contemporary carbon budget for the atmosphere requires a large "missing" carbon sink to balance anthropogenic carbon inputs. We investigated climatic effects on carbon exchanges between the atmosphere and the undisturbed biosphere and assessed the possible contribution of climate variability to the carbon sink. Empirical models and global temperature and precipitation data sets were used in the study. It was found that climate perturbations during 1940-1988 caused considerable variations in plant productivity and soil respiration. The different sensitivities of the fluxes to climate perturbations led to a significant carbon accumulation in the biosphere. The cumulative carbon sink for the period 1950-1984 ($\sim 20 \pm 5$ GtC or 10^{12} kg C) was predominantly located in mid-latitudes in the northern hemisphere (30° - 60° N) and could amount to half of the missing CO₂ sink as derived from deconvolution analyses. Our results indicate that climate variations have unequal impacts on biospheric carbon fluxes from different ecosystems and imply that caution must be exercised in generalizing in situ observations to the globe.

1. INTRODUCTION

Balancing the current budget of atmospheric CO₂ requires a large terrestrial sink (referred to as the "missing" sink) of anthropogenic emissions [Keeling

et al., 1989; Tans et al., 1990; Watson et al., 1990, 1992; Sarmiento and Sundquist, 1992; Sundquist, 1993]. The sink is estimated to be 1.4 - 4.7 GtC yr⁻¹ (1 GtC = 10^{12} kg C) [Tans et al., 1990; Sarmiento and Sundquist, 1992]. The wide range stems from uncertainties in the estimates of CO₂ emission from land use modification and of the ocean sink. Because the missing carbon is small relative to the biospheric carbon inventory, direct evidence for the missing sink is difficult to establish.

It has been hypothesized that growth stimulation of natural ecosystems by elevated CO₂ levels in the atmosphere and by enhanced nitrogen deposition from the atmosphere has contributed to the missing sink [Bacastow and Keeling, 1973; Keeling et al., 1989; Harrison et al., 1993]. Also, the effects of agricultural fertilizers and forest regrowth may have been underestimated in the calculation of carbon release from land use modification. While CO₂ fertilization on individual leaves or plants has been demonstrated in controlled experiments, nutrient limitation and community competition for resources may severely diminish ecosystem response in nature [Bazzaz and Fajar, 1992]. In particular, Mooney et al. [1991] argue that tropical rainforests and tundra ecosystems have low potentials for CO₂ fertilization.

On small scales, it is well known that temperature and precipitation are the dominant controls on plant photosynthesis and soil respiration [see Singh and Gupta, 1977; Larcher, 1980; Schlesinger, 1991]. Globally, the impact of climate variations on ecosystem functioning is suggested by the co-variation of global mean surface air temperature anomalies and atmospheric CO₂ anomalies, defined as the residual after the secular trend and mean annual cycle of CO₂ concentration has been removed

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from the Mauna Loa CO₂ record [Keeling et al., 1989]. The instrumental temperature record of the past 200 years has revealed regional temperature variations as well as interannual and interdecadal fluctuations [Folland et al., 1990, 1992; Hansen and Lebedeff, 1987, 1988]. The "dust bowls" of the 1930s attest to a widespread drought accompanying the warm anomalies of the decade. *in USA.*

In this study, we ask if these climate perturbations could have affected biospheric functioning to an extent that has impacted the atmospheric CO₂ budget. Global data sets of monthly temperature and precipitation are used to force simple empirical equations relating ecosystem net primary production (NPP) and soil respiration to climate. We focus on the period 1940–1988, when there was relatively abundant data in the 100-year precipitation record. The goal of the study is not precise numerical determination of the net carbon flux, but an identification of the ecosystems whose carbon processing has been perturbed by climate variations of the past 50 years. By calculating the net biospheric carbon flux of the last 4–5 decades, we wish to provide an overlooked dimension to the assessment of a biospheric carbon sink during the period.

2. CLIMATE DATA

Monthly surface air temperature anomalies (ΔT) for the past 200 years have been compiled, quality-checked, and are available in gridded formats for the globe [Hansen and Lebedeff, 1987, 1988]. We have supplemented the ΔT data set with a companion data set of monthly precipitation anomalies (ΔP) for 1900–1988 based on the precipitation data set of Eischeid et al. [1991].

The precipitation data set includes 5328 stations over land surface of the globe and about 4.2 million monthly records for the period. In addition to the quality checks by Eischeid et al. [1991], we further rejected precipitation anomalies beyond 3 standard deviations (about 1.5% records were eliminated) and screened the data based on spatial coherence of the anomaly fields. On the bases of spatial correlation analysis of ΔP , an interpolating radius of 300 km was established for a distance-weighted scheme to extrapolate the anomalies onto a 2.5° latitude by 2.5° longitude global grid, which has about 2600 terrestrial grid points.

The anomalies were added to a mean temperature and precipitation data set with a similar reference period [Shea, 1986] to yield global distribution of the monthly mean temperature and precipitation for 1900–1988.

3. CARBON FLUX MODELS

The net flux of carbon between the atmosphere and the biosphere represents the imbalance between photosynthetic carbon uptake by plants and

respiratory carbon release by soils and plants. A number of models simulating the two processes on various scales have been developed [e.g., Meentemeyer, 1978; Janeczek et al., 1989; Running and Gower, 1991; Rastetter et al., 1991; McGuire et al., 1992]. To estimate the global scale annually averaged net CO₂ flux caused by temperature and precipitation fluctuations during last 50 years, we employed the Miami NPP model as modified by Friedlingstein et al. [1992] and the soil respiration models of Raich and Schlesinger [1992], both of which do not require the large number of ecosystem-specific parameters that are usually needed in the more sophisticated ecosystem models. The empirical models are approximation of the relationships between current carbon fluxes and the current climate (temperature and precipitation). They are not necessarily valid under a changed climate. However, since climate variations in the last 50–100 years are only small perturbations (at least in terms of temperature and precipitation changes), we assume that these models are valid, to first order, representations of the altered carbon fluxes of the period.

The original Miami model [Lieth, 1975] had been evaluated against other observations [O'Neill and DeAngelis, 1981] and found to generate realistic patterns of global NPP [Box, 1975]. When compared with a NPP map compiled from observations by Fung et al. [1983], Friedlingstein et al. [1992] found that the model tends to overestimate NPP in the tropics and underestimate it in the high latitudes. They modified the Miami model to fit the global NPP distribution given by Fung et al. [1983]. Our calculations also suggested that the modified NPP model can produce NPP ranges for major ecosystems closer to observations than the original Miami model. Thus Friedlingstein et al.'s [1992] NPP model is used in this study:

$$\text{NPP} = \min(\text{NPPt}, \text{NPPp}) \quad (1)$$

where

$$\begin{aligned} \text{NPPt} &= 69.1875T_b, T_b \leq 8^\circ\text{C} \\ &= 1350/(1 + \exp(1.315 - 0.119T_b)), T_b > 8^\circ\text{C} \\ \text{NPPp} &= 1125(1 - \exp(-0.000664P)), \text{ in the tropics} \\ &= 1350(1 - \exp(-0.000664P)), \text{ in other regions} \\ \text{NPP} &\text{ annual net primary production (g C m}^{-2}\text{ yr}^{-1}) \\ T_b &\text{ biotemperature (}^\circ\text{C, mean annual surface air} \\ &\text{ temperature obtained by substituting negative} \\ &\text{ monthly mean temperatures with zeros)} \\ P &\text{ annual precipitation (mm).} \end{aligned}$$

Raich and Schlesinger [1992] compiled available data of soil respiration (about 170 observations all over the world) and correlated them with annual temperature and precipitation. Strong correlations with r^2 up to 0.5 were found. We chose one of the correlation equations proposed in the paper for our soil respiration calculations:

$$SR = (9.26 + 0.0127P)T + 289 \quad (2)$$

where SR is annual soil respiration in g C m⁻² yr⁻¹, and T is the annual mean surface air temperature (°C). Equation (2) includes both temperature and precipitation and is one of the regressions yielding the highest correlation possible.

Annual mean NPP and soil respiration were computed on the 2.5° × 2.5° global grid for 1940–1988. Since the focus here is the variability of the functioning of the biosphere due to climate perturbation, the effects of structural changes (e.g., due to land use) on NPP and SR are not considered.

By nature of the measurements, soil respiration is the sum of heterotrophic respiration (HR) and root respiration, and is generally greater than NPP. Here we assume that HR is locally proportional to SR and use the equilibrium balance of a steady state period between HR and NPP to determine the proportionality constant at each grid cell. At each grid point, the net carbon flux from the biosphere (soils and biomass) to the atmosphere may be expressed as

$$NET(t) = HR(t) + \Delta(t) - NPP(t) \quad (3)$$

where $\Delta(t)$ denotes the consumption of NPP(t) by herbivores and fires in year t. During the steady state period, $HR_o + \Delta_o = NPP_o$, where o denotes the averaged values of the period. If we assume that HR is proportional to SR, then equation (3) can be rewritten as

$$NET(t) = [(NPP_o - \Delta_o)/SR_o] SR(t) + \Delta(t) - NPP(t) \\ \approx [(NPP_o/SR_o) SR(t) - NPP(t)] + [1 - SR(t)/SR_o] \Delta_o \quad (4)$$

where $\Delta(t) \approx \Delta_o$ is used. Since $[1 - SR(t)/SR_o]$ is generally less than 10% and $\Delta_o \leq 10\%$ of NPP(t) [Schlesinger, 1991], the last term in equation (4) is about an order of magnitude smaller than the first term and is negligible here. Thus the net flux is calculated as

$$NET(t) = (NPP_o/SR_o) SR(t) - NPP(t) \quad (5)$$

Positive values of NET(t) denote fluxes from the biosphere to the atmosphere. The NPP to SR ratio (NPP_o/SR_o) is about 0.2–0.4 for desert, 0.6 for tropical rainforest and savanna, 0.7–1.0 for temperate forests, and 1.2–1.5 for boreal forest and tundra. These NPP/SR ranges are generally consistent with observations [Raich and Schlesinger, 1992].

We chose the 30 years from 1920 to 1949 as the steady state period for three practical reasons: (1) global temperature and precipitation records varied relatively little during the period; (2) the paucity of precipitation observations during the period would have necessitated the assumption of zero precipitation anomalies (and hence near-zero

anomalies of NPP and SR in our calculations) over much of the globe; (3) this is the period prior to significant human disturbance. Obviously, the interpretation of a net flux calculated in this way will be critically linked to the definition of a steady state biosphere.

To facilitate the discussion of NPP and respiration changes, we regrouped Matthews' [1983] vegetation types into 14 ecosystems of synthesized biome groups (Table 1).

4. RESULTS AND DISCUSSIONS

Figure 1 shows temperature and precipitation anomalies from 1940 to 1988 for the 14 ecosystems. It is interesting to note that while high latitude ecosystems, including evergreen needleleaved forest/woodland (types 9 and 10) and tundra (type 12), have experienced the largest temperature changes during the last 5 decades, most of the interannual to interdecadal precipitation variations have occurred in the tropical and subtropical ecosystems (types 1, 2, 3, 4 and 5), which is consistent with the strong El Niño/Southern Oscillation (ENSO) influence in precipitation field in the region [Ropelewski and Halpert, 1987, 1989]. It is clear that different ecosystems experienced different patterns of temperature and precipitation variations. However, common features are evident in Figure 1. First, the rapid temperature rise since the middle 1970s evident in the global temperature record [Hansen and Lebedeff, 1987, 1988] is experienced by all ecosystems. Second, most of the high carbon-capacity ecosystems, including tropical rain forest, temperate forests and grasslands, received excess rainfall in the middle 1950s and the middle 1970s while the 1960s and the 1980s were relatively dry decades.

The changes in NPP and soil respiration induced by the above temperature and precipitation variations are shown in Figure 2. It can be seen that NPP and soil respiration covary as a function of time. However NPP anomalies are more sensitive to climate perturbations, especially to precipitation perturbations, than soil respiration anomalies in all ecosystems except tropical rain forest (type 1), where soil respiration variations exceed NPP changes. Precipitation anomalies controlled both NPP and soil respiration changes in tropical and temperate ecosystems (types 1–5, 8, 11 and 13) while temperature was the dominant factor controlling soil respiration patterns in cold-deciduous and evergreen needleleaved forests (types 6–7 and 9–10). The enhanced plant growth and soil decomposition induced by the excess rainfall in the middle 1950s and the middle 1970s are evident in Figure 2 in most of the ecosystems. The rapid warming since the middle 1970s apparently caused an increasing trend in soil respiration of high latitude ecosystems (types 7, 9, 10 and 12), while its effects are not so clear in other ecosystems. Globally averaged increases in

TABLE 1. Ecosystem Types Grouped According to Their NPP and Structure

Type Index	Area 10 ⁶ km ²	Temperature ± S.D. °C*	Precipitation ± S.D. mm yr ⁻¹ *	NPP ± S.D. g C m ⁻² yr ⁻¹ †	Vegetation Type Description‡
1	12.8	25.2 ± 2.1	2096.9 ± 528.4	811.8 ± 94.5	Tropical evergreen rainforest
2	9.3	22.8 ± 3.1	1451.5 ± 387.8	739.5 ± 116.1	Tropical/subtropical/temperate seasonal forest
3	8.1	23.2 ± 3.2	1027.7 ± 377.4	548.0 ± 149.0	Trop/subtropical
4	14.2	18.6 ± 9.0	935.3 ± 519.5	518.3 ± 187.8	Arid forest/woodland
5	3.0	19.4 ± 4.0	716.4 ± 308.8	436.0 ± 129.4	Tall grassland
6	7.8	7.4 ± 7.3	759.1 ± 315.8	512.7 ± 162.5	Evergreen broadleaved Sclerophyllous forest/woodland
7	5.5	0.8 ± 8.5	591.2 ± 261.9	433.0 ± 157.8	Cold-deciduous forest, with evergreens
8	19.6	13.8 ± 8.4	472.1 ± 325.7	310.5 ± 157.2	Cold-deciduous forest, without evergreens
9	9.6	0.9 ± 4.3	578.4 ± 241.2	363.6 ± 91.2	Arid grassland
10	3.4	1.0 ± 9.1	564.1 ± 317.7	309.2 ± 127.5	Evergreen needleleaved forest
11	11.0	16.5 ± 6.5	331.3 ± 245.8	224.2 ± 131.1	Evergreen needleleaved woodland/shrubland
12	10.0	-9.6 ± 5.0	333.4 ± 172.4	192.2 ± 82.9	Xeromorphic/dwarf shrubland
13	15.9	16.7 ± 13.7	153.0 ± 178.1	97.5 ± 110.0	Tundra/cold deciduous woodland
14	2.7	10.6 ± 4.9	801.9 ± 297.2	424.5 ± 115.0	Desert
14	2.7	10.6 ± 4.9	801.9 ± 297.2	424.5 ± 115.0	Other§
Global mean		13.3 ± 6.9	764.3 ± 327.1	408.5 ± 255.0	
Total	132.8		54.2 GtC yr ⁻¹		

* The annual temperature and precipitation were computed from the 1950–1979 climatology of Shea [1986]. S.D., standard deviation.
† The NPP was computed using a modified Miami Model [Friedlingstein et al., 1992] and Shea's [1986] climatology.
‡ Matthews' [1983] vegetation map was used in the grouping.
§ This includes the minor vegetation types which could not fit into the above major types.

NPP exceeded that in soil respiration in the middle 1950s and the middle 1970s, indicating that the biosphere as a whole was a sink to atmospheric CO₂ during these periods. Although the changes of carbon fluxes in Figure 2 may be small compared to the biospheric carbon budget, they are significant for the atmospheric carbon cycle.

The latitudinal distribution and timing of our calculated net carbon flux are shown in Plate 1, where the net flux was aggregated for each 2.5° latitude zone. The high latitudes (north of 70°N and south of 40°S) are virtually negligible due to the small ice-free land areas there. Plate 1 shows that the northern middle latitudes (30°–60°N) and,

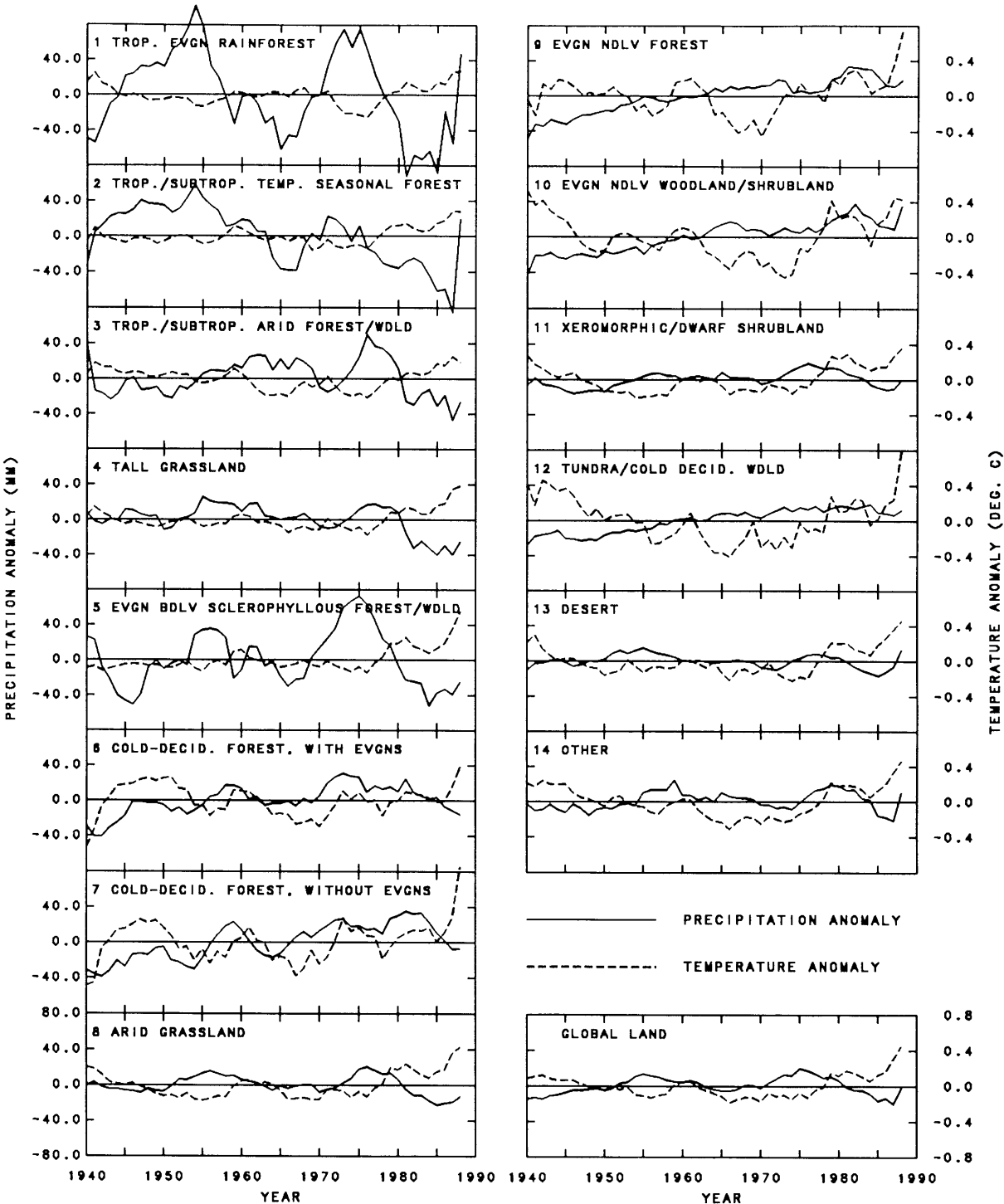


Fig. 1. Precipitation (solid line) and temperature (dashed line) anomalies (5-year running mean) during 1940-1988 for the ecosystems listed in Table 1.

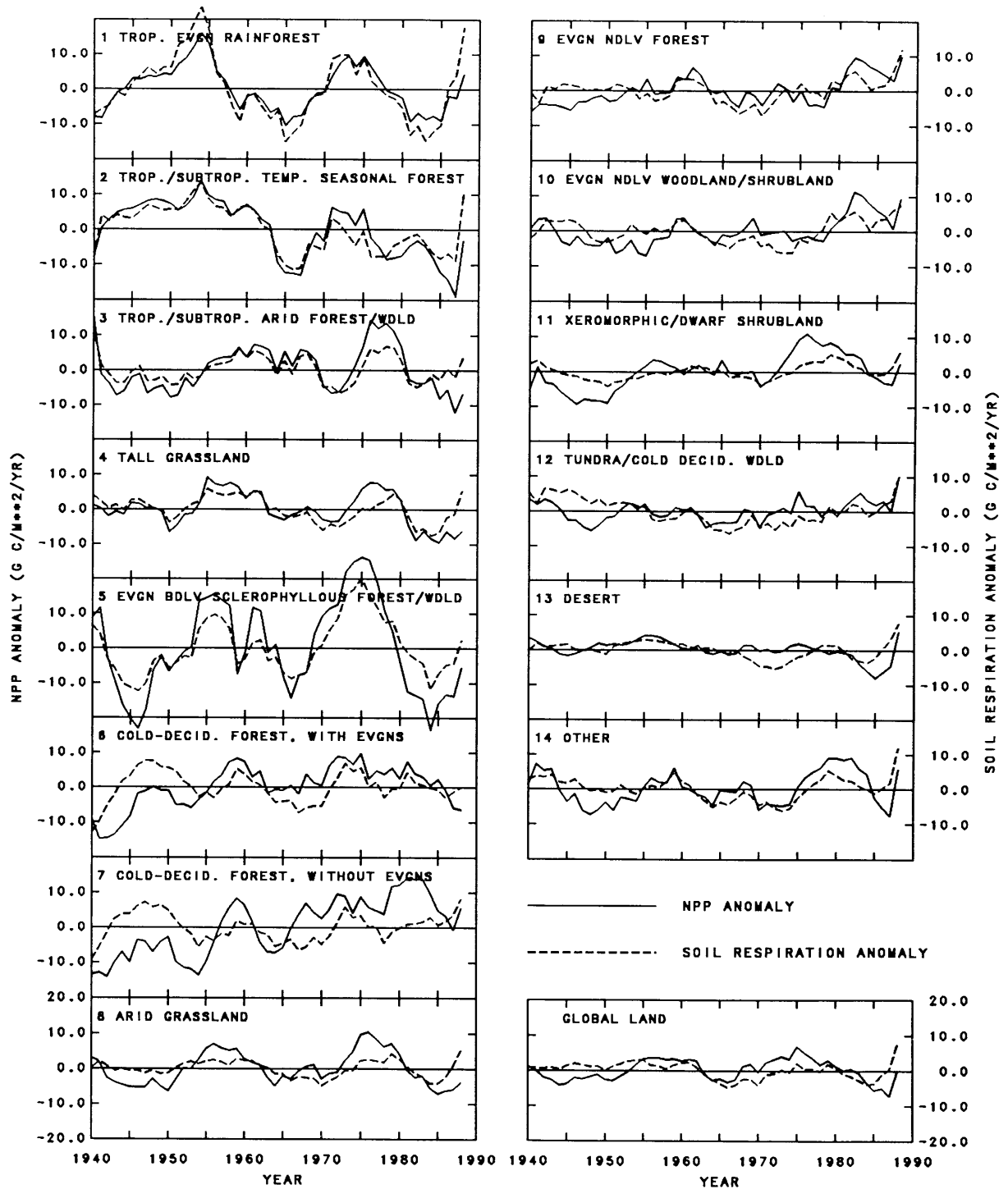


Fig. 2. Model calculated NPP (solid line) and soil respiration (dashed line) anomalies (5-year running mean) during 1940-1988 for the ecosystems listed in Table 1.

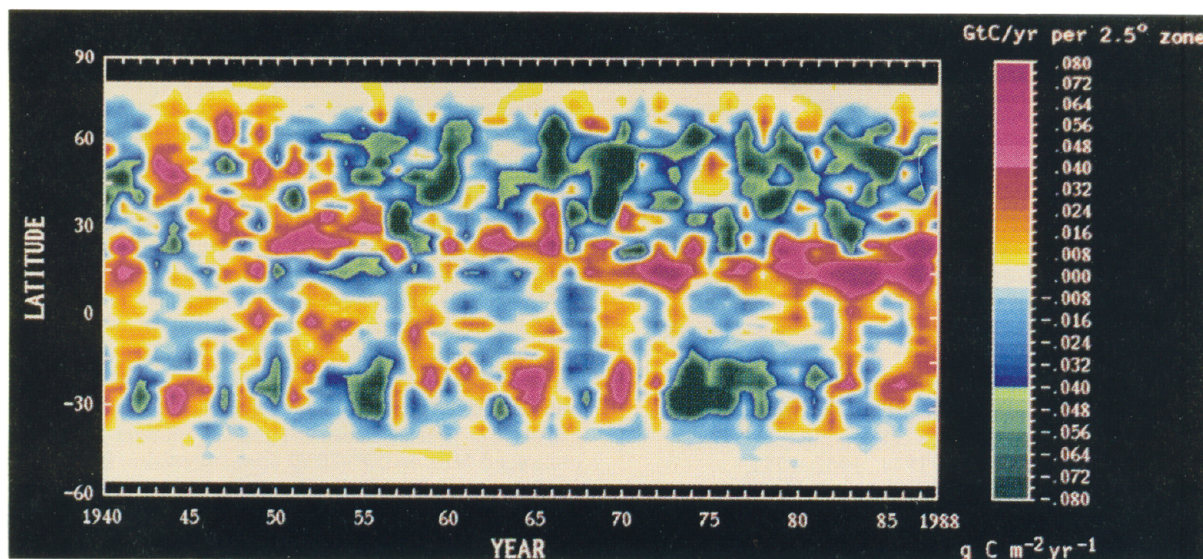


Plate 1. Zonally aggregated net carbon flux from the biosphere (for positive values) induced by climate variations during 1940–1988. Negative fluxes are from the atmosphere into the biosphere. Values are in unit of GtC yr^{-1} per each 2.5° latitude zone.

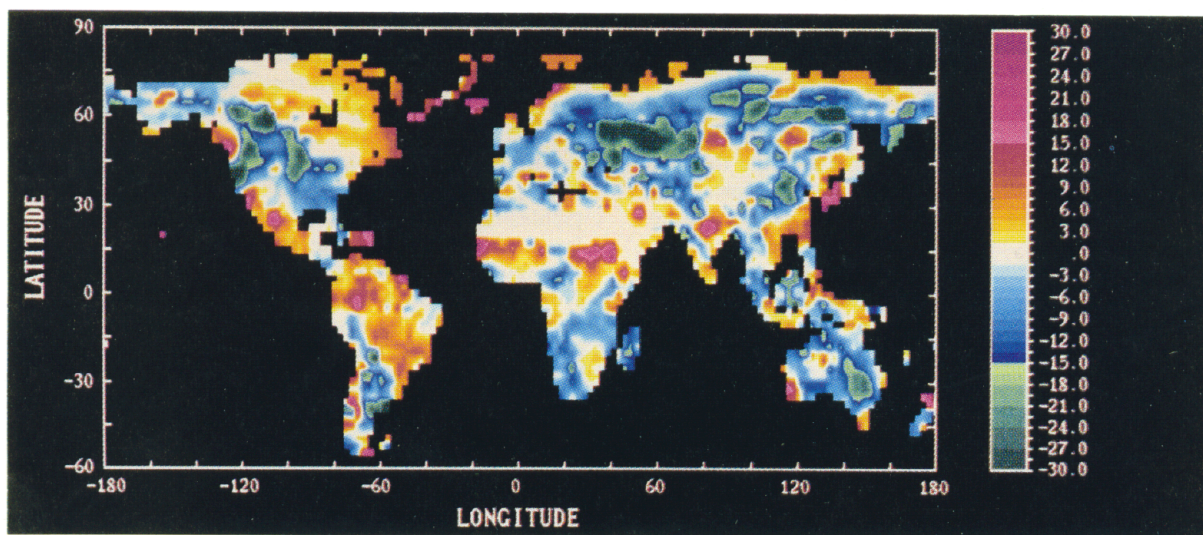


Plate 2. Geographic distribution of the mean net carbon flux ($\text{g C m}^{-2} \text{yr}^{-1}$) from the biosphere (for positive values) induced by climate changes during 1940–1988.

secondarily, the $15^\circ\text{--}40^\circ\text{S}$ zones are important regions in terms of net carbon flux. It may be surprising that the tropical and subtropical regions ($15^\circ\text{S--}25^\circ\text{N}$), where land area is second only to the northern middle latitudes, did not have much net carbon accumulation over the period 1940–1988 even though the largest precipitation changes were found in the region. This is because the NPP and soil respiration models have similar sensitivities at high temperatures and precipitation, while the NPP model is more sensitive to climate changes at low

temperatures. Thus the large precipitation variations at low latitudes ($15^\circ\text{S--}25^\circ\text{N}$) caused compensating changes in NPP and soil respiration (see Figure 2) and resulted in insignificant changes in net carbon flux. For the period 1980–1987, the averaged northern mid-latitude carbon sink is estimated to be $\sim 0.2 \text{ GtC yr}^{-1}$. The effects of CO₂ and nitrogen fertilization, which are not included in this calculation, would further enhance the magnitude of the sink. This supports the inference of three-dimensional atmospheric tracer model studies

[Keeling et al., 1989; Tans et al., 1990], which all require, for this period, a large carbon sink in the northern middle to high latitudes in order to maintain the observed atmospheric CO₂ gradient.

Plate 1 shows that most of the northern middle latitudes have been a persistent sink of atmospheric CO₂ since the early 1950s, partially as a result of the abundant precipitation in the region during the last 3–4 decades. The excess rainfall in the middle 1950s and the middle 1970s in the southern subtropics (10°S–30°S) was the major cause for the large biospheric carbon sinks in these two periods, which are clearly shown in Figure 3. Plate 1 also shows that in low latitudes there were significant ENSO-associated interannual variations of the net carbon flux, with more positive net carbon fluxes during warm El Niño years (1940/1941, 1952/1953, 1957/1958, 1965, 1969/1970, 1972, 1967/1977, 1982/1983, and 1987) and more negative net carbon fluxes during cold La Niña years (1941/1942, 1949/1950, 1955/1956, 1967/1968, 1970/1971, 1973/1974, 1980/1981, and 1985/1986). This supports the hypothesis proposed by Keeling et al. [1989] and Elliott et al. [1991] that climate effects on the terrestrial biosphere seem to be a likely source of much of the interannual variation in atmospheric CO₂, which rises during warm events and falls during cold events [Keeling et al., 1989].

Plate 2 shows the geographic pattern of the net carbon flux averaged for the period 1940–1988. Again a middle latitude sink is evident while the Amazon appears to be one of the major carbon source regions. The important point to be made here is that climate effects on the biospheric carbon fluxes may be totally different from one region to another and a meaningful global picture can not be obtained from a limited number of regional studies.

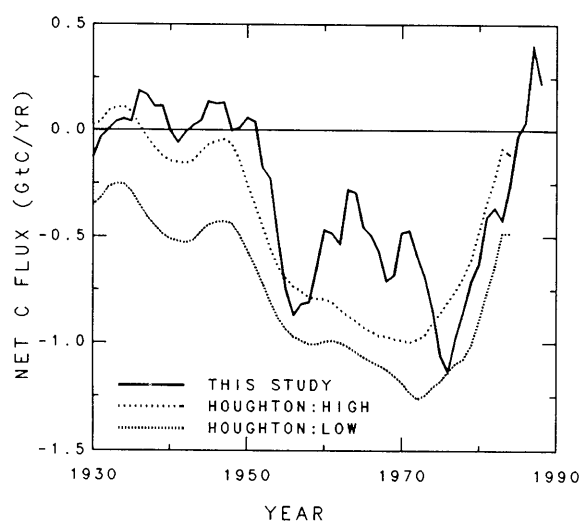


Fig. 3. Comparison of the climate-induced net carbon flux with the missing CO₂ sink as derived from deconvolution analysis by Houghton [1989].

This is important for interpreting the data from in situ observations.

The calculated NPP variations are qualitatively consistent with a recent estimate of the biomass change in European forests during last three decades [Kauppi et al., 1992]. Their analysis, which was based on available forest surveys, shows that the growing stock of European forests, which primarily consist of cold deciduous and evergreen needleleaved forests located between 50° and 70°N, increased from 1950 to 1980 with some decrease in the 1960s. This agrees well with the calculated NPP changes between 50° and 70°N (not shown here) as well as the NPP change patterns of evergreen needleleaved and cold deciduous forests (types 9–10 and 6–7, cf. Figure 2). Our estimate for the the climate-enhanced NPP in 50°–70°N zones between 1970 and 1988 is about 0.03 GtC yr⁻¹, which is about one third of Kauppi et al.'s [1992] estimate of 0.085–0.12 GtC yr⁻¹ between 1971 and 1990. Kauppi et al. [1992] suggested that the nitrogen fertilization from the increased fertilizer use and air pollution may be responsible for the biomass increase in European forests. We argue that the interdecadal variations of climate also plays an important role in regulating the growth of the forests.

Figure 3 suggests that the global undisturbed vegetation and soils have sequestered carbon from the atmosphere at a global mean rate ranging from 3.8 g C m⁻² yr⁻¹ (for the 1960s) to 8.0 g C m⁻² yr⁻¹ (for the middle 1970s), which is within the range of the long-term accumulation rates of soil carbon [Schlesinger, 1990]. Local net carbon flux may be as large as 30 g C m⁻² yr⁻¹, as shown in Plate 2.

Figure 3 shows that the calculated net carbon flux from the terrestrial biosphere is also qualitatively consistent with the missing CO₂ sink from deconvolution analysis [Houghton, 1989]. However, the calculated carbon sink is about 0.5 GtC yr⁻¹ smaller than the deconvolution flux during the 1960s. The difference might stem from the inaccuracy in the model formulation and the data we used, or from other effects (e.g., CO₂ and N fertilization and forest regrowth in temperate latitudes) that might induce additional absorption of carbon by the biosphere.

Figure 3 also shows that the climate-induced biospheric carbon flux may be able to account for at least half of the missing CO₂ for 1950–1984 as estimated by Houghton [1989]. According to Houghton [1989], the total missing biotic carbon flux after deforestation and land use effects were accounted for is about 30–35 GtC for the period 1950–1984. Our calculation suggests that about 20 ± 5 GtC was absorbed by natural ecosystems due to climate variations in the same period. If the two estimates are correct, they then indicate that other processes (e.g., CO₂ and N fertilization and forest regrowth) in the biosphere could induce a carbon sink of about 13 ± 8 GtC or 0.37 ± 0.23 GtC yr⁻¹ during the period.

The above analysis is affected by several kinds of possible errors. First, the simplicity of the model formulation and assumptions used in our carbon flux calculation (see section 3) introduces uncertainties in the calculated carbon fluxes. In particular, the net flux calculated here is critically linked to the definition of a steady state biosphere. Our calculation showed that the net flux would be about 0.6 GtC yr⁻¹ more positive if 1950–1979, instead of 1920–1949, is chosen as the steady state period. With the documented changes in atmospheric composition in recent decades and the suggestions of a greenhouse signal in the recent climate record, we believe that the earlier period may be a closer approximation to biospheric equilibria. More sophisticated ecological models of global scale, in which NPP and soil respiration can be calculated in a self-consistent manner, are needed for further investigation. However, it is unlikely that the use of more sophisticated models could dramatically alter the patterns of NPP and soil respiration variations shown in Figure 2, primarily because temperature and precipitation dominate global NPP and soil respiration so strongly [Lieth, 1975; Raich and Schlesinger, 1992] that the addition of other secondary factors is unlikely to change the basic patterns.

Second, the data, especially precipitation, used in the calculation have instrumental errors. Using a method similar to that of Hansen and Lebedeff [1987] and results of Groisman et al. [1991], Eischeid et al. [1991] and Hulme [1992], we estimated that the error in our precipitation data is probably less than 20–50 mm (~10%) for high latitudes and less than 10–20 mm in other regions. These errors are considerably smaller than the interdecadal variation of precipitation (see Figure 1). Our derived long-term changes in global land precipitation are consistent with the results of other analyses [Diaz et al., 1989; Vinnikov et al., 1990; Eischeid et al., 1991]. The large error in precipitation data at high latitudes (due to snowfall) should not affect the calculated NPP significantly because temperature often is the limiting factor in these latitudes.

Third, Houghton's [1989] two curves of missing carbon flux were obtained by subtracting his high (low) land use flux from Siegenthaler and Oeschger's [1987] high (low) non fossil carbon flux. Although the resulting two curves may bound the missing carbon sink, the possible range of the sink should be much larger. The point to be made here is that due to the large uncertainties in the estimated non fossil and land use carbon fluxes, an additional significant carbon sink might still exist.

It is difficult to give a precise range of the uncertainty in our estimate of the global climate-induced net carbon sink (~20 GtC) for 1950–1984. Our analysis suggests an uncertainty of ~10% for the precipitation data. The error in the temperature data set is about 5% [Hansen and Lebedeff, 1987]. If we assume that Equations (1) and (2) have an error bar

of ~10%, then the uncertainty for the net carbon sink should be around 25% (or 5 GtC).

SUMMARY

The effects of climate variation on natural biospheric carbon uptake and release fluxes during 1940–1988 were investigated using empirical models and global surface temperature and precipitation data sets. It was found that precipitation changes were the major source of the variations in NPP and soil respiration in tropical and subtropical ecosystems whereas temperature fluctuations controlled the variations in soil respiration in cold-deciduous and evergreen needleleaved forests. The excess rainfall in the middle 1950s and the middle 1970s caused an increase in plant productivity as well as soil respiration in most of the ecosystems while global warming since the middle 1970s induced a rising trend in soil respiration in high latitude ecosystems. These climate-induced changes in biospheric carbon fluxes are significant perturbations to the atmospheric carbon budget.

Our results show that various climate changes in different regions can have unequal impacts on biospheric carbon fluxes. This is important in planning and interpreting in situ studies.

Assuming that the undisturbed biosphere was roughly in steady state during the period 1920–1949, our calculation suggests that the climate variations induced a biospheric CO₂ sink after 1950, which is generally consistent with the results from deconvolution analyses. However, the drier conditions in the 1960s resulted in a weaker biospheric carbon sink in the period than deconvolution results. The sink was predominantly located in the northern middle latitudes (30–60°N). Our results suggest that the biosphere sequestered about 20 ± 5 GtC from the atmosphere due to climate change during 1950–1984, which is about half of the missing carbon sink as suggested by deconvolution analyses for the period.

It should be emphasized that our biospheric carbon fluxes were calculated from simple empirical models using data having errors. Thus the conclusions drawn from them, especially those related to net carbon fluxes, are only suggestive. However, it seems likely that the climate changes in the last 5 decades have induced considerable variations in biospheric carbon fluxes, which might have resulted in a significant biospheric carbon sink. Also, the northern middle latitudes seem to be the dominant region for the carbon sink.

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