Tellus (1999), 51B, 156–164
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TELLUS
ISSN 0280–6495

Soil erosion and atmospheric CO_2 during the last glacial maximum: the rôle of riverine organic matter fluxes

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(Manuscript received 11 March 1998; in final form 10 July 1998)

ABSTRACT

Atmospheric CO_2 is consumed both by organic matter formation and chemical rock weathering, and subsequently discharged as dissolved organic carbon, particulate organic carbon, and dissolved inorganic carbon to the oceans by rivers. In the long term, varying the ratio of the amount of atmospheric CO_2 consumed by continental erosion and the amount of CO_2 released during carbonate precipitation and organic matter respiration in the oceans can change the CO_2 content in the atmosphere. The purpose of this paper is to determine whether riverine organic carbon fluxes during the last glacial maximum (LGM) may have been different from today in order to assess the potential impact on atmospheric CO_2. Previous studies mainly focused on the role of the river fluxes of inorganic carbon in this respect, but none of them examined possible variations in the fluxes of organic carbon, although the erosion of organic carbon actually represents the bulk of the atmospheric CO_2 consumption by continental erosion. We therefore applied a global carbon erosion model to a LGM scenario in order to determine the riverine fluxes of organic matter during that time. The climatic conditions during the LGM were reconstructed using a computer simulation with a general circulation model. It is found that during the LGM the riverine organic carbon input into the oceans was at least $\sim 10\%$ lower than today. Most of the reduction of the total organic matter fluxes is due to the reduction of the fluxes of dissolved organic carbon. The fluxes of particulate organic carbon remained almost unchanged. The oceanic response to the lower carbon input was estimated on the basis of a present-day steady state budget for organic river carbon in the oceans, and implies that the reduction of the river fluxes were more than counterbalanced by lower burial rates due to the smaller shelf area during the LGM. This suggests that both the lower river carbon input and the relatively greater share of this carbon being subjected to oceanic respiration, acted as a negative feedback to the low atmospheric CO_2 content during the LGM.

1. Introduction

Varying the rates of continental erosion has been widely postulated as a major control of atmospheric CO_2 during the Earth's history on a multimillion-year time scale (Walker et al., 1981; Berner et al., 1983; Raymo et al., 1988; Berner, 1991, 1994; François and Walker, 1992; Raymo and Ruddiman, 1992; Raymo, 1994). More recently, it has been discussed that this may be also the case on shorter time scales, in particular for the glacial/interglacial climate cycle during Quaternary times (Gibbs and Kump, 1994; Sharp et al., 1995; Munhoven and François, 1996). The general idea is that during cold periods such as the last glacial maximum (LGM), weathering

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fluxes on the continents and the concomitant consumption of atmospheric CO_2 may have been considerably greater than today, either due to a shift in weathering intensity as a consequence of climate change (Munhoven and François, 1996), or due to high weathering fluxes related to deglacial meltwater discharge pulses (Gibbs and Kump, 1994; Sharp. et al., 1995). Because of the close coupling of atmospheric CO_2 and global climate as documented in ice cores (Barnola et al., 1989), this implies that continental erosion could be at least partly responsible for the global cooling during the LGM.

Most of the studies investigating the consumption of atmospheric CO_2 by continental erosion during the LGM have only considered the chemical weathering of rocks. Silicate weathering is especially important here because it can act as a net sink of atmospheric carbon. None of these studies, however, examined possible variations in the fluxes of organic carbon. Also, the erosion of organic matter represents a permanent transfer of carbon from the atmosphere to the oceans by rivers, both in the form of dissolved organic carbon (DOC), and in the form of particulate organic carbon (POC). In the present-day carbon cycle, the river fluxes of organic carbon are quantitatively even more important than the fluxes of inorganic carbon (Meybeck, 1982; Ludwig et al., 1996b; 1998). Omitting them in global budget considerations leaves an uncertainty of more than half of the river carbon, which is too much for any reliable estimate of the atmospheric CO_2 consumption by continental erosion during the LGM.

In this study, we respond to this data need. A global erosion model that has been developed to simulate the present-day river fluxes of organic carbon (Ludwig et al., 1996a) was applied to a LGM scenario in order to determine the amount of atmospheric CO_2 consumed by the erosion of organic matter during that time. The study is complementary to a previous study in which we determined the amount of atmospheric CO_2 consumption by chemical rock weathering during the LGM (Ludwig et al., in press).

2. Data and methods

2.1. Modelling of the river carbon fluxes

Our approach to predict the river fluxes of organic carbon during the LGM includes both the fluxes of dissolved and particulate organic carbon (in the following F_{DOC} and F_{POC}, respectively). It is based on a number of empirical models relating the river carbon fluxes to their major controlling factors for present-day (Ludwig et al., 1996a). Together, these relationships form the Global Erosion Model for the atmospheric CO_2 consumption by the erosion of organic carbon — GEM-C_{org} (Ludwig et al., 1998; Probst et al., in press).

For DOC, a multiple regression model including runoff intensity, basin slope, and the amount of carbon stored in the soils is the most significant model to predict the fluxes on a global scale:

$$sF_{DOC} = 0.0044 \times Q - 8.49 \times \text{slope} + 0.0581 \times \text{soilC}. \quad (1)$$

sF_{DOC} is the specific DOC flux in t km^{-2} yr^{-1}, Q the runoff intensity in mm, slope the mean grid point slope in radian; and soilC is the mean carbon content in the soils in kg/m^3 (data from the United States Department of Agriculture–Soil Conservation Service (USDA–SCS), pers. com.).

For POC, sediment fluxes are the dominant controlling parameter. POC fluxes generally increase with increasing sediment fluxes (F_{TSS}), but the percentage of organic carbon in the total suspended solids clearly decreases with increasing sediment concentrations. This relationship was mathematically fitted, allowing calculation of the percentage of organic carbon in the total suspended solids as a function of sediment concentration:

$$\text{POC\%} = -0.160 \times \log(c\text{TSS}^3) + 2.83 \times \log(c\text{TSS}^2) - 13.6 \times \log(c\text{TSS}) + 20.3. \quad (2)$$

POC% is the percentage of organic carbon in the total suspended solids (TSS), and cTSS is the TSS concentration in mg/l. Sediment yields (sediment fluxes divided by basin area) can be calculated as a function of hydroclimatic, geomorphological, and lithological parameters, that is runoff intensity, basin slope, an index characterizing rock hardness, and an index characterizing rainfall variability over the year. For all climate types together, the following equation is the most significant model to describe the specific sediment fluxes (sF_{TSS}) globally:

$$sF_{TSS} = 0.020 \times (Q \times \text{slope} \times \text{four}). \quad (3)$$

sF_{TSS} is in t km^{-2} yr^{-1} for F_{TSS}, and Q and slope are the same as for eq. (1). four is in mm and calculated as the sum of the square of mean monthly precipitation over mean annual precipitation for all 12 months of the year (Ludwig et al., 1996a). The best correlated parameter combination varies to some extent when the regressions are established for specific climate types only, but it is always a combination of the above given parameters that yields greatest correlation coefficients. More details of our methods to determine river sediment yields can be found in Ludwig and Probst (1996, 1998). Finally, POC fluxes to the oceans can be calculated as a function of sediment fluxes:

$$F_{POC} = \text{POC}\% \times F_{TSS}/100. \quad (4)$$

2.2. *Climatic data sets for the LGM*

Changes of the continental area during the last glacial maximum related to the sea level fall, as well as the appropriate LGM ice-coverage of the continents, were taken from the reconstructions of Peltier (1994). For the simulation of the Earth's climate prevailing during the LGM, we used the monthly temperature and precipitation climatologies calculated by the ECHAM2 general circulation model (GCM). More details on the ECHAM2 model are given in Lautenschlager and Herterich (1990), or in Lautenschlager (1991). From the ECHAM2 output files, runoff intensity could be derived by taking the difference between the precipitation and evaporation fields, but this value turns out to be negative many times, which makes this method less suitable for the determination of runoff during the LGM. For this reason, we also derived this parameter by an empirical method. In this modelling (Ludwig, in press; Ludwig et al., in press), precipitation and temperature, which is uniquely used to derive potential evaporation, are the main parameters determining runoff. Morphology, however, also represents a non-negligible controlling factor. Given the same temperature and precipitation values, a steep morphology increases the runoff ratio (precipitation divided by runoff) compared to a more flat morphology.

Our empirical modelling hence neglects other possible effects on runoff such as changes in the water use efficiency of plants due to the different CO_2 content in the atmosphere during the LGM. This effect has been examined by a number of authors, but there is no general consensus how important it may be at the global scale. Aston (1984), for example, suggests that streamflow may increase by 40 to 90% when doubling atmospheric CO_2, whereas Skiles and Hanson (1994), in a process-based study of arid and semi-arid watersheds in the US, report that changes in rainfall dominate over second order effects, such as stomatal closure, in runoff production. Nevertheless, one has to mention here that assuming a positive feedback of higher atmospheric CO_2 concentrations on continental runoff would mean that the runoff values we calculated for the LGM may rather be upper limits. The real values may then have been lower.

Finally, one has also to say that, as this is also a common practice in other modelling studies (Esser and Lautenschlager, 1994; Gibbs and Kump, 1994), we obtained the final LGM climatologies by subtracting the ECHAM2 derived anomalies from the corresponding data sets for the present-day distributions, which are based upon observations. For a few data points, it could happen that this method yields negative values. In these cases, the grid point values were set to 0, although cutting off negative values overestimates to some extent the global LGM values with regard to the ECHAM2 simulation. In the LGM data sets we created, however, this effect is globally only of small importance. Only for sediment yields, this procedure results more often in negative values than for the other data sets, which is related to the high spatial variability of this parameter. Setting here negative values to 0 leads to an overestimation of the global F_{TSS} value of nearly 6%, whereas this is only around 1% both for the precipitation and runoff data set we created. More information of our reconstruction of the climatic conditions during the LGM can be found in Ludwig et al. (in press).

3. Results and discussion

3.1. *River fluxes at present-day*

When GEM-C_{org} is applied to the total ice-free and exoreic continental area on the basis of global data sets for the corresponding controlling factors (see Ludwig et al., 1996b), it yields a total amount of 0.36 gigatons of atmospheric carbon per year (GtC/yr) that is discharged as riverine organic matter to the oceans (in the following F_{CO_2} is for total values, and sF_{CO_2} for specific values). The corresponding fluxes of water and of sediments are 41750 km^3/yr and 16 Gt/yr, respectively. About 57% of F_{CO_2} is due to dissolved organic carbon, and about 43% due to particulate organic carbon. This accounts for about 61% of the total atmospheric CO_2 consumption by continental erosion (including chemical rock weathering) at present-day (Ludwig et al., 1996b).

3.2. *River fluxes during the LGM*

In order to calculate the corresponding fluxes for the last glacial maximum, we assumed that both the morphological characteristics of the continents and the distribution of organic carbon in the soils were the same as for today. We therefore only changed in our simulation the data set for runoff intensity derived from the reconstructed LGM climatologies (see above). One problem is that the morphological characteristics and the soil carbon content are not known for the grid elements which became land area during the LGM as a result of the lower sea (in the following: "new" continental grid points). Consequently, GEM-C_{org} cannot be applied to these grid elements, and it is thus not possible to calculate one definitive value for total F_{CO_2} during the LGM. We therefore used the following method to estimate the global CO_2 consumption by soil erosion. For each climate type, for each continent, and for each of the continental areas which are drained to the different ocean basins, the average F_{CO_2} and runoff values were calculated only on the basis of the grid points that belong to the exoreic and ice-free continental area both for present-day and for the LGM (in the following: "old" continental grid points). At the same time, runoff for the new grid points was estimated by a triangular interpolation of the runoff ratios between the old continental grid points over the oceans and applying them to the LGM precipitation values. The total F_{CO_2} fluxes within each category can then be established according to:

$$F_{CO_2} = (sF_{CO_2\text{-old}} \times \text{area}_{\text{old}}) + (sF_{CO_2\text{-old}} \times \text{area}_{\text{new}} \times Q_{\text{new}}/Q_{\text{old}}). \quad (5)$$

The units are TgC/yr for F_{CO_2}, t km^{-2} yr^{-1} for sF_{CO_2}, $10^6 \times km^2$ for area, and mm for Q. Finally, the arithmetic mean of the global F_{CO_2} values in the three categories was selected as best estimate for the global atmospheric CO_2 consumption by organic matter erosion during the LGM (Table 1). Hence, the above method assumes that sF_{CO_2} on the new grid points linearly increase or decrease according to the variations of runoff intensity compared to the old continental grid points. This procedure can be justified by the dominant role of continental runoff as controlling factor for F_{CO_2} (see above).

Table 1 shows that both for DOC and for POC, the total river flux to the oceans was lower during the LGM than today (by about 12% and 3%, respectively). This mainly reflects the global reduction in continental runoff as a result of climate change, although also the effective erodible continental area (exoreic and ice-free) was slightly reduced during the LGM (Fig. 1, Table 1). Due to a steeper temperature gradient towards the poles, the humid climate zones of the mid and high latitudes were considerably smaller than today. Note from Table 1 that the areas of the temperate wet and the tundra and taiga climates were reduced by about 34% and 22%, respectively. On the other hand, the area increased slightly for the tropical wet climate due to the gain of continental area in the low latitudes. Runoff intensity especially decreased on the old continental grid points, whereas it was considerably greater on the new continental grid points than on the rest of the continents (for a discussion, see also Ludwig et al., in press). Nevertheless, all in all, climate was on average more arid during the last glacial maximum than today.

In our simulation, the reduction in F_{DOC} is greater than the reduction in F_{POC}. This is because the overall sediment flux to the oceans is even slightly increased during the LGM compared to the present-day. Dry regions generally show a high susceptibility to mechanical erosion (Ludwig and Probst, 1998), which implies that a global shift from humid to more dry conditions does not necessarily result in lower sediment fluxes. However, one has to say here that our estimate for F_{TSS} during the LGM may also be somewhat too high because of methodological reasons (see above), so that the global riverine sediment flux during the last glacial maximum may have been at about the same value as for today. Another uncertainty concerning our global F_{TSS} value is related to the fact that our modelling cannot

Table 1. *Fluxes of water, sediments, DOC and POC to the oceans during the last glacial maximum, as well as changes of these fluxes compared to present day*

	Area[(1)]		Runoff		F_{TSS}		F_{DOC}		F_{POC}	
Grouping according to	(10^6 km^2)	changes (%)	(km^3)	changes (%)	(Tg/yr)	changes (%)	(TgC/yr)	changes (%)	(TgC/yr)	changes (%)
Climate types [(2)]										
polar (no ice)	6.10	57	2213	190	194	545	12.7	310	5.1	408
tundra and taiga	17.99	−23	4483	−35	1102	69	26.9	−41	15.4	2
temperate dry	10.04	4	996	37	1233	−49	3.4	16	7.5	−40
temperate wet	11.15	−34	5992	−23	2902	−12	22.3	−37	27.2	−17
tropical dry	20.77	−5	3135	1	3175	−30	16.7	5	20.3	−27
tropical wet	25.99	4	21174	−5	6665	31	98.5	−3	69.3	1
desert	7.51	26	228	245	1034	2486	1.4	362	4.1	1253
total	99.55	−6	38221	−8	16304	2	181.9	−11	148.7	−6
Continents										
Africa	19.17	5	4010	−3	689	−29	19.2	−4	9.6	−10
Europe	7.16	−25	2211	−28	1131	35	10.4	−37	9.9	−6
North America	12.07	−48	4668	−35	1858	−41	20.5	−48	17.8	−36
South America	18.65	5	10228	−8	2795	−5	48.2	−7	29.9	−13
Asia	36.73	13	16184	6	10477	32	77.4	6	83.4	16
Australia	5.77	29	920	19	333	59	4.9	25	3.9	30
Antarctis	—	—	—	—	—	—	—	—	—	
total	99.55	−6	38221	−8	17284	8	180.6	−12	154.4	−2
Ocean basins[(3)]										
Arctic Ocean	11.90	−30	1469	−55	176	−23	12.4	−50	3.7	−40
North Atlantic	19.37	−29	11328	−16	3002	−17	54.5	−23	33.6	−14
South Atlantic	17.86	5	4695	−7	665	28	24.3	−6	10.5	11
Pacific	24.00	14	13296	−2	8177	10	57.2	0	67.8	−1
Indian Ocean	19.33	16	6457	25	4434	25	26.1	22	33.0	17
Mediterranean	7.09	5	977	−10	769	8	4.4	−5	5.6	−9
below 60°S	—	—	—	—	—		—	—	—	—
total	99.55	−6	38221	−8	17224	7	178.8	−13	154.1	−2
best estimate					16937	6	180.4	−12	152.4	−3

[(1)] Only exoreic and without ice cover.
[(2)] Climate types are defined only on the basis of temperature and precipitation data (Ludwig et al., 1996a).
[(3)] This is the continental area that is drained to the different ocean basins.

account for sediment retention in river basins (Ludwig and Probst, 1998). Due to the lower sea level during the LGM, it is possible that the overall sediment delivery to the oceans may have been greater than today, not because mechanical erosion on the continents was greater, but simply because some of the sediments that have previously been stored in the drainage basins were then exported to the oceans.

Finally also the global F_{DOC} value we calculated may be the subject of discussion, since we used in our LGM simulation the present-day distribution of organic carbon in the soils. There is some evidence that the biospheric soil carbon pool during the LGM may have been considerably smaller than at present, although some controversy exists about the extent of the reduction (for a review, see for example Adams and Faure, 1996). According to GEM-C_{org}, less carbon in the soils would also reduce the global DOC flux to the oceans (eq. (1)). The soil carbon data set we used (see above) results in a global value of about

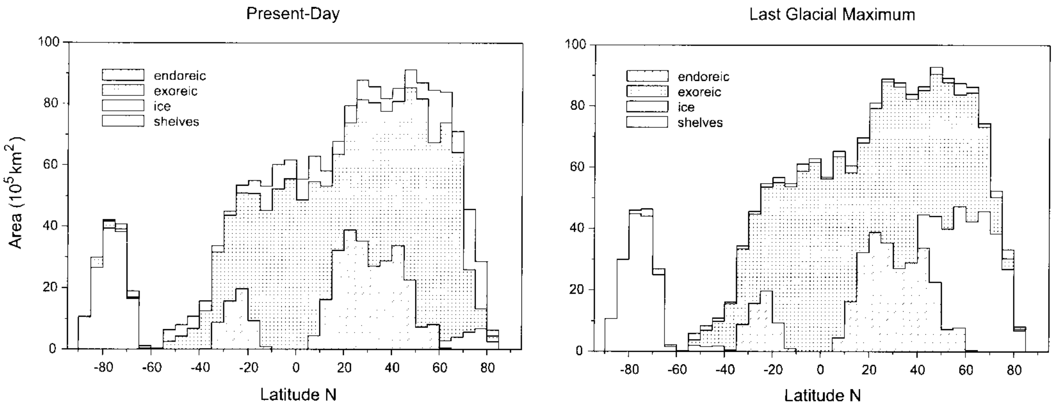

Fig. 1. Latitudional distribution of continental area during the glacial maximum. Extension of the shelves was calculated according to Peltier (1994) by taking all grid points between 0 and −200 m water depth. It was assumed that the endoreic regions (the parts of the continents that are not drained to the oceans) were the same as today.

1470 GtC for present-day. Great amounts of this carbon are actually stored in the soils of the high latitudes, which have been partly covered by the ice sheets during the LGM. This effect alone reduces the global soil carbon pool with about 390 GtC in our simulation. Adams et al. (1990), however, estimated the global soil carbon pool at the LGM to be only 625 GtC, which would result in an additional reduction with about the same value as being lost by the ice sheets (their global estimate for present-day soil carbon is 1395 GtC and hence close to our value). When we uniformly apply this additional reduction to our soil carbon data set in our simulation yields a lowering of global F_{DOC} during the LGM by about 19% instead of 12% (as given in Table 1).

In summary, we estimate that during the LGM the global reduction in the atmospheric CO_2 consumption related to organic matter erosion (DOC and POC) may have been at least as great as 10%.

3.3. *Fate of the river carbon in the oceans*

An important question for the evaluation of the role of the fluvial carbon in the global carbon cycle is the fate of this carbon once it has entered the oceanic system. Assuming a steady state, i.e., assuming that the river input is balanced by the removal of carbon from the oceans, Smith and Hollibaugh (1993) proposed a global budget for riverine organic carbon in the oceans. They concluded from a review of data in the literature that at present-day, approximately one third of the total organic carbon input by rivers is buried in sediments, whereas two thirds of it returns to the atmosphere via respiration in the oceans. No distinction is made in their budgets between the fate of POC and DOC because much of the POC in the water column and in the upper sediment layers may be converted to DOC. About equal amounts of the fluvial organic carbon becomes involved in the coastal organic matter cycling and in the offshore ocean organic matter cycling, but the ratio of burial to respiration in the coastal zone is about 1.5, while this is only about 0.1 in the open sea. The shelves hence account for most of the overall burial of fluvial organic matter by far (80% of total burial), whereas the bulk of the respiration takes place in the open oceans (70% of total respiration).

During the last glacial maximum, the extension of the shelves was considerably smaller than today (Fig. 1). Taking an average depth of the shelves of 200m, we calculate this reduction to be about 60%. This implies also, that the burial of the riverine organic matter in the oceans may have been lower during the LGM. For a first order estimate, one may linearly reduce the share of the river carbon being involved in the coastal organic matter cycling according to the shelf area reduction, and assume that the above mentioned ratios of burial to respiration both in the coastal and in the open ocean waters may also hold for the LGM. A reduction of the riverine input of organic carbon to the oceans by 10% as predicted in this study would then reduce the value for the organic matter burial in the oceans by about a factor of two (from 116 to 55 TgC/yr) whereas it would increase the value for the oceanic respiration by about 10% (from 247 to 272 TgC/yr). As mentioned above, no distinction is made here between the fate of DOC and POC, but our finding that the riverine supply of POC during the LGM may have been as great as that today is in good agreement with the idea of a lower share of riverine organic matter burial as a consequence of the reduced shelf area. It is POC especially that has the potential of being buried.

Of course, there are no means to verify whether the oceans have been (or are actually) in a steady state with regard to the river input of organic carbon. One may argue that due to the lower sea temperatures during the LGM also the oxidation of DOC in the oceans should have been slowed down, so that both lower burial rates on the shelves and lower respiration rates on the open ocean should have resulted in an accumulation of the organic matter in the oceans. This could at least partly explain why there is actually a discrepancy between estimates of the biospheric carbon pool during the LGM based on palynological, pedological, and sedimentological evidence (Adams et al., 1990) and those based on the oceanic $\delta^{13}C$ record (Crowley, 1995). The former method reveals a global reduction of the terrestrial biosphere about twice as great as the second method (1350 GtC in comparison to 610 GtC). Carbon isotopes, however, can only record changes of biospheric carbon when a part of this carbon has been oxidized and transferred to the atmospheric/oceanic system, but not when it has been transferred into the oceanic pool of organic carbon via the erosion and river transport pathways. Assuming, for example, that instead of an increase of oceanic respiration by about 10% as calculated above, respiration was reduced by about 10% due to lower respiration coefficients as a result of the lower temperatures could increase the organic matter pool in the oceans by 740 GtC (which is the difference between the above given values) within a relative short time interval of about 15000 yrs.

4. Conclusions

On the basis of a global erosion model for the riverine fluxes of organic carbon to the oceans and a climatic computer simulation with a GCM we determined the global organic matter erosion during the last glacial maximum to be at least $\sim 10\%$ lower than today. At the same time, burial of this carbon in the oceans may have been reduced because of the much smaller shelf area during the LGM, subjecting a greater share of the riverine organic carbon to slow respiration in the open oceans than at present-day. Especially riverine DOC fluxes were reduced, whereas POC fluxes remained about the same as for today. This rather supports the general idea of a negative coupling of the atmospheric CO_2 consumption by continental soil erosion and global temperature on Earth for glacial/interglacial times, although there is still little known about the cycling and fate of riverine organic matter in the oceans, and more information is needed to confirm or to reject such a hypothesis.

Naturally, our findings are strongly dependent on the reliability of the used GCM outputs. Such computer simulations are still at an early stage of development, and there has been some discussion in the literature whether the prescribed sea surface temperatures normally used to constrain the LGM simulations are realistic or not (Adams and Faure, 1996). Another point is that our approach underlies the assumption that variations in the carbon reservoirs in glacial/interglacial time scales can be understood as a continuous shift between steady state conditions corresponding to the average LGM and present-day climates, respectively. This may be misleading, since it cannot account for exceptional conditions such as the breakdown of the ice sheets at the end of the LGM, which may have had a considerable impact on the erosion fluxes and hence also on the carbon budgets. The values we calculate for the river fluxes of organic carbon during the last glacial maximum therefore present rather general trends than precise figures. It becomes nevertheless evident from our simulation that variations in the erosion of soil organic matter may also have the potential to influence atmospheric CO_2 in glacial/interglacial climate cycles, a fact that has been overlooked in most of the previous studies which were only looking at chemical rock weathering and riverine fluxes of dissolved inorganic carbon.

5. Acknowledgments

This work was supported by the Commission of the European Communities within the program

European Study of Carbon in the Ocean, in the Biosphere, and in the Atmosphere (ESCOBA/ Biosphere; contract no. ENV4-CT95–0111). We are especially grateful to Dr. P. Reich of the United States Department of Agriculture Soil Conservation Service for the global soil carbon data set. This paper is a contribution to the working group Weathering as a Carbon Sink of the INQUA commission on The Global Carbon Cycle and to the IGCP Project no. 404 Terrestrial Carbon in the Past 125 Ky.

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