Atmospheric CO$_2$ consumption by continental erosion: present-day controls and implications for the last glacial maximum

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Abstract

The export of carbon from land to sea by rivers represents a major link in the global carbon cycle. For all principal carbon forms, the main factors that control the present-day fluxes at the global scale have been determined in order to establish global budgets and to predict regional fluxes. Dissolved organic carbon fluxes are mainly related to drainage intensity, basin slope, and the amount of carbon stored in soils. Particulate organic carbon fluxes are calculated as a function of sediment yields and of drainage intensity. The consumption of atmospheric/soil CO$_2$ by chemical rock weathering depends mainly on the rock type and on the drainage intensity. Our empirical models yield a total of 0.721 Gt of carbon (Gt C) that is exported from the continents to the oceans each year. From this figure, 0.096 Gt C come from carbonate mineral dissolution and the remaining 0.625 Gt C stem from the atmosphere ($F_{\mathrm{CO_2}}$). Of this atmospheric carbon, 33% is discharged as dissolved organic carbon, 30% as particulate organic carbon, and 37% as bicarbonate ions. Predicted inorganic carbon fluxes were further compared with observed fluxes for a set of 35 major world rivers, and possible additional climatic effects on the consumption of atmospheric CO$_2$ by rock weathering were investigated in these river basins. Finally, we discuss the implications of our results for the river carbon fluxes and the role of continental erosion in the global carbon cycle during the last glacial maximum.

Keywords: atmospheric CO$_2$; continental erosion; organic and inorganic carbon flux; Last Glacial Maximum

1. Introduction

Continental erosion represents a sink for atmospheric carbon. Atmospheric CO$_2$ is consumed both by organic matter formation and chemical rock weathering, and subsequently transferred as dissolved organic carbon (DOC), particulate organic carbon (POC), and dissolved inorganic carbon to the oceans by rivers. The latter occurs mainly in the form of bicarbonate ions (HCO$_3^-$). Processes that withdraw this carbon from the ocean reservoir are the organic matter and carbonate sedimentation, as well as the respiration of organic matter in the water column. Carbonate precipitation and organic matter respiration liberate CO$_2$ to the atmosphere, while the carbon that is incorporated in the sediments becomes part of the lithosphere.

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Previous studies estimated the present-day total of atmospheric carbon that is consumed by continental erosion (F_{CO_2}) to be about 0.7 to 0.8 Gt per year: 0.26 to 0.30 Gt are discharged to the oceans as HCO_3^- ions (Berner et al., 1983; Meybeck, 1987; Probst, 1992; Amiotte-Suchet, 1995), 0.20 to 0.22 as DOC (Meybeck, 1982; Spitzy and Leenheer, 1991; Meybeck, 1993), and 0.18 to 0.23 Gt as POC (Meybeck, 1982; Ittekkot, 1988; Meybeck, 1993). These fluxes can account for about one third of the estimated net oceanic carbon uptake for present-day (Sarmiento and Sundquist, 1992). The role of continental erosion in the long-term or geochemical carbon cycle of the Earth has been discussed, for example, by Garrels and Mackenzie (1971), Walker et al. (1981), Berner et al. (1983), and Berner (1991).

Concerning the erosion of organic carbon, the CO_2 consumption is governed by the photosynthesis reaction, that can be simplified as following:

$$CO_2 + H_2O \rightarrow CH_2O + O_2 \tag{1}$$
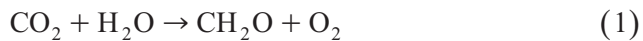
All carbon in the organic matter that is washed to the rivers is of atmospheric origin. Also for riverine HCO_3^- ions resulting from weathering of silicate rocks, all carbon comes from the atmosphere (mainly via soil CO_2), as it can be seen, for example, in the following equation for albite hydrolysis:

$$2NaAlSi_3O_8 + 2CO_2 + 11H_2O \rightarrow Al_2Si_2O_5(OH)_4$$
$$+ 2HCO_3^- + 2Na^+ + 4H_4SiO_4 \tag{2}$$
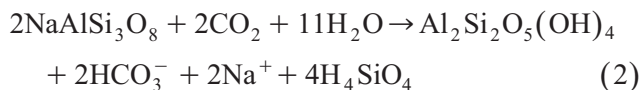
For HCO_3^- ions resulting from weathering of carbonate rocks, however, only half of the carbon originates from atmospheric/soil CO_2, while the other half comes from the carbonate mineral. This can be shown, for example, with the following equation for the calcite dissolution:

$$CaCO_3 + CO_2 + H_2O \rightarrow Ca^{2+} + 2HCO_3^- \tag{3}$$
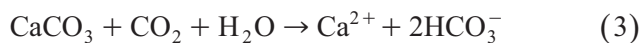
For all of the carbon forms mentioned above, we determined the major factors that control the fluxes at the global scale. Our main purpose is to provide a modelling tool in order to estimate river carbon fluxes for present-day, which may also be applied for different scenarios of climate change. In this paper, we present detailed budgets for the present-day fluxes and discuss then the implications of the controlling factors for the situation during the last glacial maximum (LGM) in order to evaluate the role of continental erosion for changes in the glacial/interglacial carbon cycle.

2. Data and methods

River fluxes of carbon were taken from the literature, and the climatic, biologic, and geomorphologic characteristics were extracted from various ecological databases. We determined first the best possible regression model to describe river carbon fluxes at the global scale, and then extrapolated fluxes on the basis of the determined relationships and the corresponding datasets to the overall continental area. The work is based on two previous studies on fluxes of inorganic carbon (Amiotte-Suchet, 1995; Amiotte-Suchet and Probst, 1995), and on fluxes of organic carbon (Ludwig et al., 1996a). In this study, all fluxes were recalculated with common datasets, and new datasets for continental runoff and organic carbon storage in soils were used.

Global distribution of runoff was taken from the UNESCO Atlas of World Water Balance (Korzoun et al., 1977). Isohyets of the maps were digitized and subsequently gridded to a $0.5° \times 0.5°$ longitude/latitude resolution. This yields a global figure of about $41\,800$ km^3 yr^{-1} for the exoreic runoff from the continents (not considering the runoff from the regions under permanent ice cover). The corresponding drainage area is about 106×10^6 km^2. Mean organic carbon content in the soils was extracted from a global dataset developed at the Soil Conservation Service of the United States Department of Agriculture (USDA-SCS). The dataset is, for example, described by Eswaran et al. (1993). Rock types were taken from a global lithological map developed at our CNRS Institute (Amiotte-Suchet, 1995). All other datasets we used are the same as used by Ludwig et al. (1996a).

3. Inorganic carbon fluxes

3.1. Relations with drainage and lithology

The flux of atmospheric/soil CO_2 consumed by rock weathering (F_{CO_2-RW}) is mainly a function of drainage intensity (Q), and of the rock type that is

drained by the surface waters. This has been described in detail elsewhere (Amiotte-Suchet and Probst, 1993a,b, 1995). Empirical relationships were established using data published by Meybeck (1986) concerning runoff and HCO_3^- concentrations of 232 monolithologic watersheds in France. The watersheds were grouped into seven lithologic classes that are representative for the major rock types outcropping on the continents, and linear models between $F_{CO_2\text{-RW}}$ and Q were determined for each of the seven classes. According to Eqs. (2) and (3), $F_{CO_2\text{-RW}}$ was considered to be equal to the HCO_3^- fluxes in waters draining silicate rocks, and equal to half of the HCO_3^- fluxes in waters draining carbonate rocks. Together, these relationships form the Global Erosion Model for atmospheric CO_2 consumption by chemical weathering (GEM-CO_2; Amiotte-Suchet and Probst, 1995).

3.2. Relations with climate

The applicability of the GEM-CO_2 relationships to the scale of large river basins was tested for the Congo and the Amazon under tropical wet climate, and for the Garonne in France under temperate wet climate. Comparison of the GEM-CO_2 results with estimates derived from field measurements revealed a good agreement between both methods (Amiotte-Suchet and Probst, 1993a,b, 1995). Nevertheless, it is not excluded that additional climate effects are possible. The influence of climatic factors, such as temperature, on the consumption of atmospheric CO_2 by rock weathering has been proposed by several authors (e.g., Garrels and Mackenzie, 1971; Berner et al., 1983; Meybeck, 1987; Probst, 1992).

A climatic influence is indicated if one compares observed HCO_3^- fluxes (Table 1) for various major world rivers with the fluxes calculated by GEM-CO_2. This comparison is shown in Fig. 1a. It is interesting to note that the observed values and the values calculated from the model compare well for the tropical wet climate, whereas the model seems to underestimate the fluxes for the other climate types (for a definition of the climate types, see Ludwig et al., 1996a). Because of the climatic heterogeneity which is found in many of the river basins (Ludwig et al., 1996a), we applied the following method to test if there may be a systematic deviation of the

observed fluxes from the GEM-CO_2 fluxes with respect to climate. We calculated an average specific HCO_3^- flux with GEM-CO_2 for all climatic subunits within the basins. Note that all basin grid points that fall into the same class are taken as one subunit, which does not necessarily mean that this subunit is one geographically connected region. The average specific flux for the whole basin (F_{model}) is the sum of the specific values (F_n), multiplied by the percentage that the units occupy in the basin (a_n), divided by 100:

$$F_{model} = (a_1 F_1 + a_2 F_2 + a_3 F_3 + \ldots + a_i F_i) / 100 \tag{4}$$

Specific fluxes are in 10^3 mol km^{-2} yr^{-1}, and climate types (i) correspond to those defined in the legend of Fig. 1a. A multiple regression between the observed fluxes ($F_{observed}$) of all basins and the area weighted model fluxes of all climatic subunits ($a_n F_n$) in these basins can then help to identify the importance of each climate type with regard to the deviation between the observed fluxes and the model fluxes. It leads to the following relationship:

$$F_{observed} = (6.55 a_4 F_4 + 2.91 a_5 F_5 + 1.30 a_7 F_7$$
$$+ 0.85 a_8 F_8) / 100 \tag{5}$$

$n = 31$, $r = 0.95$ with $P < 0.001$ for: $a_4 F_4$, $a_5 F_5$, $a_7 F_7$, $a_8 F_8$.

P is the significance level, r the correlation coefficient, and n the number of rivers considered in the equation. In the regression, we omitted the Danube, Mahandi, Godavari and Magdalena from the rivers in Table 1. With these rivers, the regression has a significant positive intercept. It is interesting to note that the first three rivers have the greatest percentages of cultivated area in their basins (around 50%) compared to all other rivers in Table 1 (see Ludwig et al., 1996a). It is therefore not excluded that they have elevated fluxes because of the use of fertilizers in their basins. Fertilizers can increase natural HCO_3^- fluxes without consuming CO_2 by dissolving carbonate minerals (Amiotte-Suchet, 1995). We do not know why also the Magdalena River does not fit here, but since this river has a low data quality index in Table 1, we did not further follow this question.

The coefficients in Eq. (5) indicate that in the dry temperate ($i = 4$) and the tundra and taiga climates ($i = 5$), GEM-CO$_2$ underestimates considerably the HCO$_3^-$ fluxes. In the temperate wet climate ($i = 7$), the underestimation is less (about 30%), while in the tropical wet climate ($i = 8$) the fluxes seem to be slightly overestimated (about 15%). The tropical dry climate was not significant in the regression (it tends to have regression coefficients between those of a_5F_5 and of a_7F_7). Also the other climate types which are partly represented in some of the basins were not found to be significant, which can be

Table 1
Bicarbonate fluxes of major world rivers

	Area, 10^6 km^2	Q, km^3 yr^{-1}	HCO$_3^-$, 10^3 mol km^{-2} yr^{-1}	DQ	Source
Amazon	5.903	6300	462	2	(1)
Zaire	3.704	1382	823	3	(2)
Mississippi	3.246	490	309	3	(3)
Ob	3.109	416	236	1	(4)
Parana	2.868	541	119	2	(5)
Yenisei	2.567	587	279	2	(6)
Lena	2.465	532	237	1	(7)
Amur	1.926	342	124	1	(4)
Nil	1.874	90	122	2	(8)
Changjiang	1.822	930	921	3	(9)
Ganges/Brahmaputra	1.656	1220	1116	3	(10)
Mackenzie	1.615	270	291	1	(11)
Niger	1.540	200	72	2	(12)
Zambesi	1.413	100	29	2	(4)
Murray	1.131	12	16	3	(13)
St. Lawrence	1.114	450	600	2	(14)
Orinoco	1.026	1100	219	2	(15)
Indus	0.912	240	654	2	(16)
Mekong	0.864	470	526	3	(10)
Yukon	0.843	210	467	3	(3)
Huanghe	0.823	59	235	2	(9)
Danube	0.773	200	801	2	(10)
Orange	0.716	11	27	3	(10)
Colorado	0.708	20	67	3	(3)
Columbia	0.664	186	374	3	(3)
Si Kiang	0.464	300	1411	1	(17)
Limpopo	0.344	6	37	1	(7)
North Dvina	0.329	105	798	1	(7)
Godavari	0.311	100	474	1	(18)
Magdalena	0.285	240	703	1	(10)
Fraser	0.248	98	390	1	(4)
Yana	0.243	33	71	1	(7)
Mahandi	0.190	94	483	3	(19)
Rio Negro (Argentinia)	0.175	29	189	1	(20)
Hungho	0.159	120	975	1	(21)

Basin area was calculated in this study. Q was taken from Global Runoff Data Center Koblenz (1991), or, if not available, from Milliman et al. (1995). DQ is a data quality index: (1) poor, (2) sufficient, (3) good. Sources: (1) Probst et al. (1994); (2) Probst et al. (1992); (3) U.S. Geological Survey (annual); (4) Meybeck (1987); (5) Kempe (1982); (6) Gitelson et al. (1988); (7) Livingstone (1963); (8) Kempe (1983); (9) Wei-Bin et al. (1983); (10) Meybeck (1979); (11) Reeder et al. (1972); (12) Martins (1983); (13) Herczeg et al. (1993); (14) Cossa and Tremblay (1983); (15) Paolini et al. (1987); (16) Arain (1987); (17) Qunying et al. (1987); (18) Biksham and Subramanian (1988); (19) Subramanian (1979); (20) Depetris (1980); (21) Ming-Hui et al. (1982).
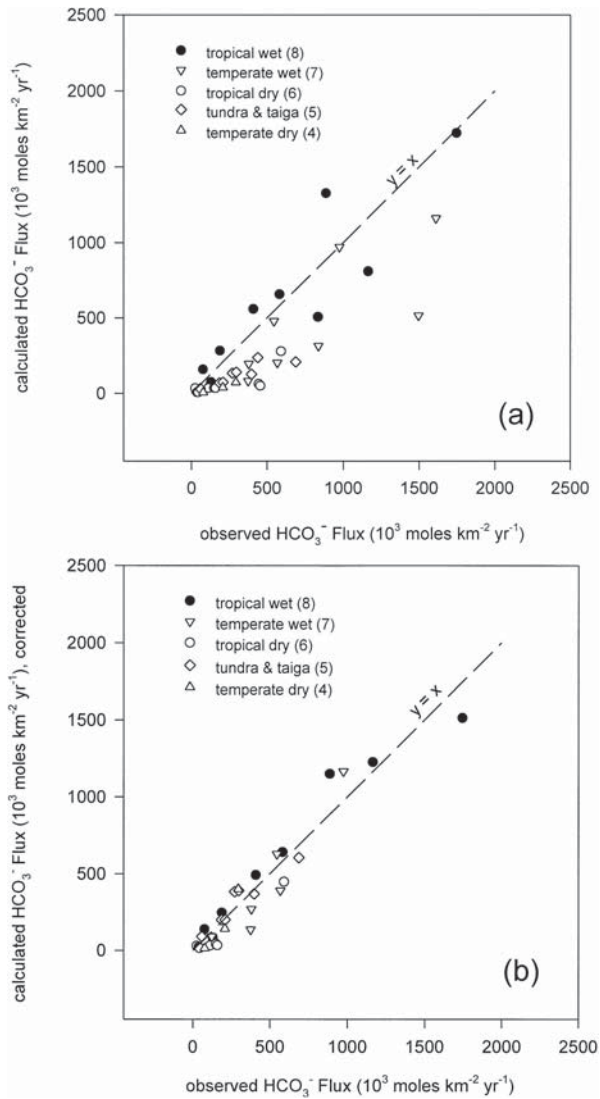
Fig. 1. (a) Comparison of the observed HCO_3^- fluxes from Table 1 with the HCO_3^- fluxes calculated with GEM-CO_2. The rivers are grouped according to their average climatic situation. (b) is as (a), but adjusted with the climatic correction factors determined in Eq. (5).

explained with the small areas they occupy in the basins (ice-free polar climate type) and their low flux contribution in the basins (desert climate type).

At first sight, one may be surprised because the empirical relationships forming GEM-CO_2 were established in the temperate wet climate, but when applied to the scale of large basins, they fit best in the tropical wet climate. This may be explained by the basic relationships of the model, which reflect

uniquely the situation in small watersheds with a mean size of 8 km^2 (Meybeck, 1986). At the scale of a large river basin, these watersheds represent more the headwater regions, where ground water contributions to the fluxes are normally small with regard to the lower course of the river. The relationships should therefore mainly reflect the composition of typical surface or subsurface waters. Note that the deviation of the observed fluxes with respect to the GEM-CO_2 fluxes generally becomes greater for dryer and colder climates, and the coefficients in Eq. (5) decrease in the order of the increasing specific drainage intensities we calculate for each climate type. This is indicating that the deviation of the fluxes may be at least partly related to the average residence time of the water in the basins, and thus to the extent of the water–rock interactions. In the tropical wet climate, the average residence time of the water is naturally short because of the great and permanent water flux in the basins, and the chemical composition of the river waters probably reflects closest the composition of surface waters with only relative short contact with the outcropping lithology.

However, there may also be other effects related to climate that can influence river bicarbonate fluxes. For example, in the tropical climate, deep lateritic formations depleted in alterable minerals are typical soil formations, which may generally reduce chemical erosion and, thus, also specific HCO_3^- fluxes. Generally, the role of the soils for weathering processes is an important question in this context. Since it can be supposed that much of the CO_2 involved in the weathering reactions may originate from the soils via the biological respiration of organic matter, the CO_2 consumption by rock weathering should also be enhanced by elevated pCO_2 values (partial pressure of CO_2) in the soils. Very little is known about the spatial and temporal variability of this parameter at the global scale. It is possible that in certain regions, the pCO_2 values in the soils may be especially elevated in winter when the soils are covered by a snow cover, preventing the CO_2 to degas to the atmosphere. Also this could explain why in Eq. (5) great regression coefficients are found for the colder climate types. In Fig. 1b, we corrected the GEM-CO_2 fluxes of Fig. 1a for each climate by a simple multiplication with the coefficients determined in Eq. (5). For the climates that were not found to be

significant in the regressions, the fluxes remained unchanged. The corrected fluxes compare now better with the observed ones.

4. Organic carbon fluxes

In the following, we discuss the main relationships that control organic carbon fluxes at the global scale. Analogous to GEM-CO_2, we will call these relationships the Global Erosion Model for organic carbon (GEM-C_{org}). The flux of atmospheric CO_2 consumed by the erosion of organic matter is thus $F_{CO_2\text{-OM}}$.

4.1. Dissolved organic carbon

Multiple regression analyses based on the average DOC fluxes (F_{DOC}) observed for various world rivers (many of the data are summarized, for example, in the work of Degens et al., 1991) and the environmental characteristics of their river basins indicate that drainage (Q), the steepness of morphology, and the amount of carbon stored in soils (soil C) are the main factors that control DOC fluxes globally (Ludwig et al., 1996a). Based on the datasets used in this study, the following equation is the best model to estimate DOC fluxes:

$$F_{DOC} = 0.0044\,Q - 8.49\,\text{slope} + 0.0581\,\text{soil C}$$

$$r = 0.90,\ P < 0.0001,\ n = 29 \tag{6}$$

All parameters are significant at least with $P < 0.01$. F_{DOC} is given in t km^{-2} yr^{-1}, Q in mm yr^{-1}, slope in radian, and soil C in kg m^{-3}. DOC fluxes become greater with increasing drainage intensities, flatter morphologies, and larger carbon reservoirs in the soils. This suggests that soils are globally the major contributors to riverine DOC, which corresponds to the general assumption in most of the studies on this topic. However, it is only recently that scientists have become aware that basin morphology may also play an important role in controlling organic carbon fluxes (e.g., Clair et al., 1994; Ludwig et al., 1996a).

4.2. Particulate organic carbon

Among all factors that may control the export of particulate organic carbon on a global scale, it is the total suspended sediment (TSS) flux that shows the most significant relationship with the POC flux (F_{POC}). The POC percentage in the riverine suspended solids generally decreases with increasing sediment concentrations following a nonlinear relationship that has been described by Ludwig et al. (1996a).

A reliable extrapolation of POC fluxes over the total continents requires thus the creation of an additional dataset for sediment fluxes. For this study, we created a global dataset on TSS fluxes by an extrapolation of observed river data, coupled to an empirical relationship between sediment fluxes and a morphoclimatic index (for details, see Ludwig et al., 1996a; Ludwig and Probst, 1996, 1998). It yields a quantity of 18.9 Gt yr^{-1} for the total sediment flux to the oceans, what is in good agreement with the recent estimate of Milliman and Syvitski (1992) who assumed that the total sediment flux may be about 20 Gt yr^{-1}. From this dataset, we derived then global F_{POC}.

5. Global and regional budgets for present-day erosion fluxes

Fig. 2 summarises our approach to predict the atmospheric CO_2 consumption by continental erosion at the global scale. We calculate that continental weathering represents a sink for atmospheric CO_2 of 0.625 Gt C/yr. A total of 0.205 Gt can be attributed to DOC, 0.187 Gt to POC, and 0.233 Gt to HCO_3^- coming from the atmosphere. HCO_3^- that originates from carbonate dissolution makes an additional flux (F_{carb}) of 0.096 Gt C/yr. Fig. 3a–c show the corresponding maps of F_{DOC}, F_{POC}, and $F_{CO_2\text{-RW}}$, respectively. The fluxes agree well with other literature estimates (see Section 1), confirming the applicability of our method. Only HCO_3^- fluxes are somewhat lower. If one corrects them for climatic effects according to Eq. (5), they become about 34% greater. In Table 2, carbon fluxes are further regionalized with respect to major climates, different continents, and different ocean basins. Also here, possible climatic effects on inorganic carbon fluxes are listed. Since the corrected figures indicate possible trends rather than to display precise figures, we discuss and present here principally the non-corrected GEM-CO_2
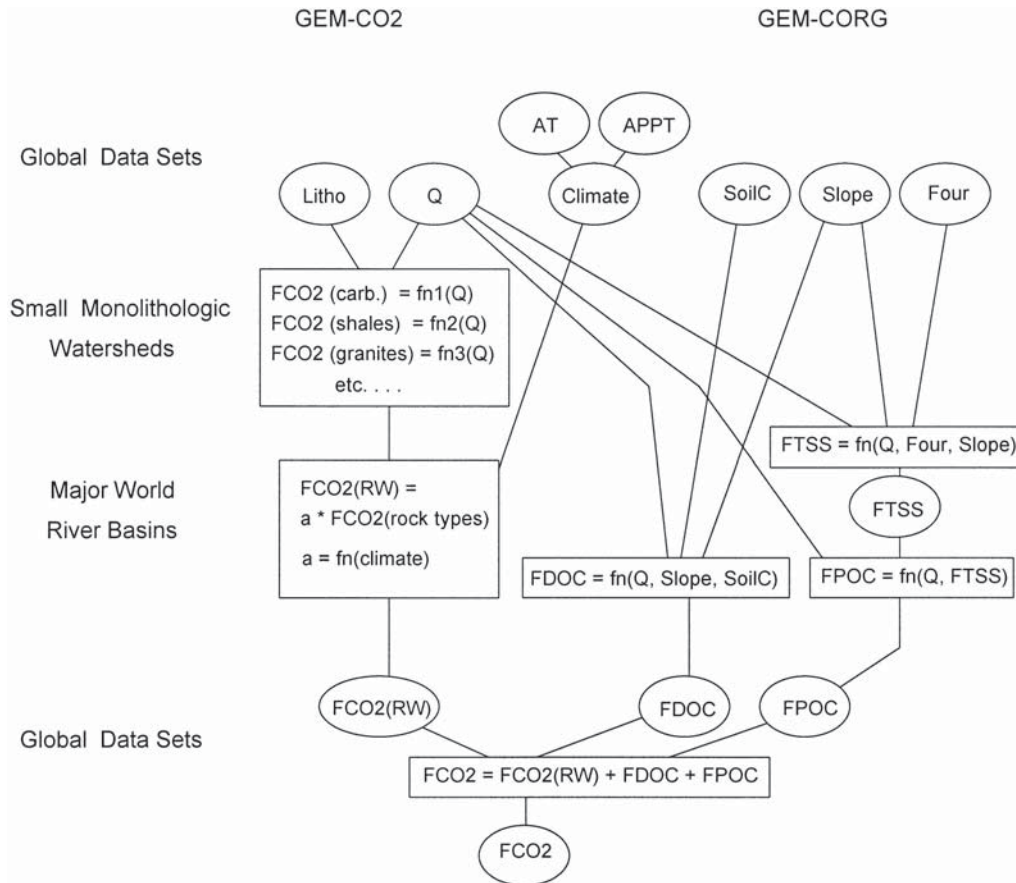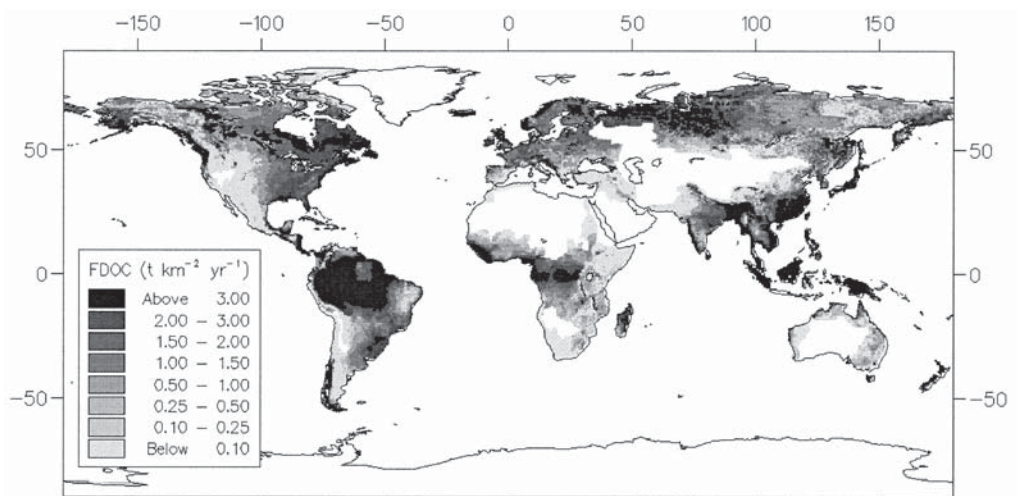
Fig. 2. Flow diagram of the approach applied in this study in order to predict the consumption of atmospheric CO_2 by continental erosion (F_{CO_2}). AT, mean annual temperature; APPT, annual precipitation total; Litho, global lithology; Q drainage intensity; Soil C, organic soil carbon; Four, index to characterize variability of precipitation over the year (Ludwig et al., 1996a); FTSS, sediment flux; $F_{CO_2(RW)}$ flux of CO_2 consumed by rock weathering; F_{DOC}, flux of dissolved organic carbon; F_{POC}, flux of particulate organic carbon. Circles represent global datasets while rectangles represent basic empirical relationships.

output. Climatic effects are only included if it is mentioned in the text. We consider, however, the corrected values to represent the possible bias of the GEM-CO$_2$ output with respect to the real fluxes at the global scale.
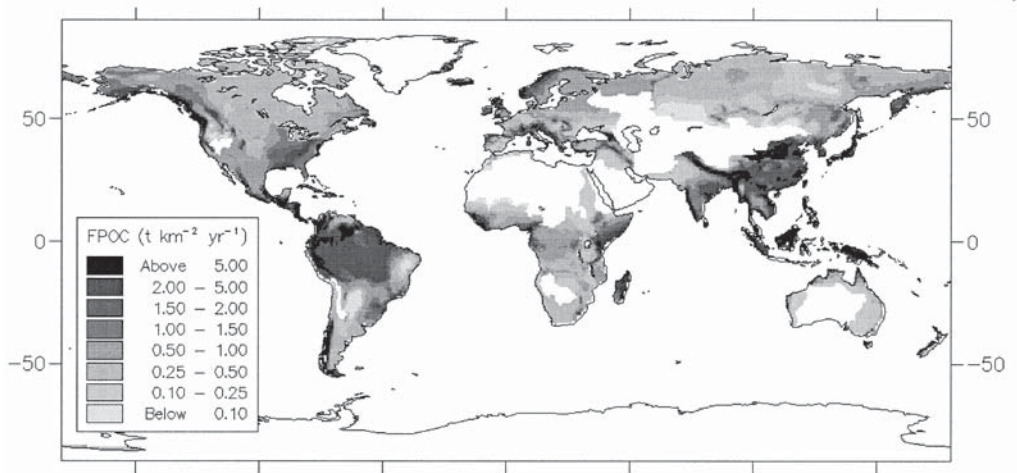
Table 2 shows that for all carbon forms, the tropical wet climate is the most important climate for the consumption of atmospheric CO_2 by continental erosion. This points out the important role of drainage intensity for the carbon fluxes. 54.9% of total F_{CO_2-RW}, 53.9% of total F_{POC}, and 49.6% of total F_{DOC} are discharged from the wet tropics to the oceans. Note that the corresponding runoff accounts for 50% of the global runoff.

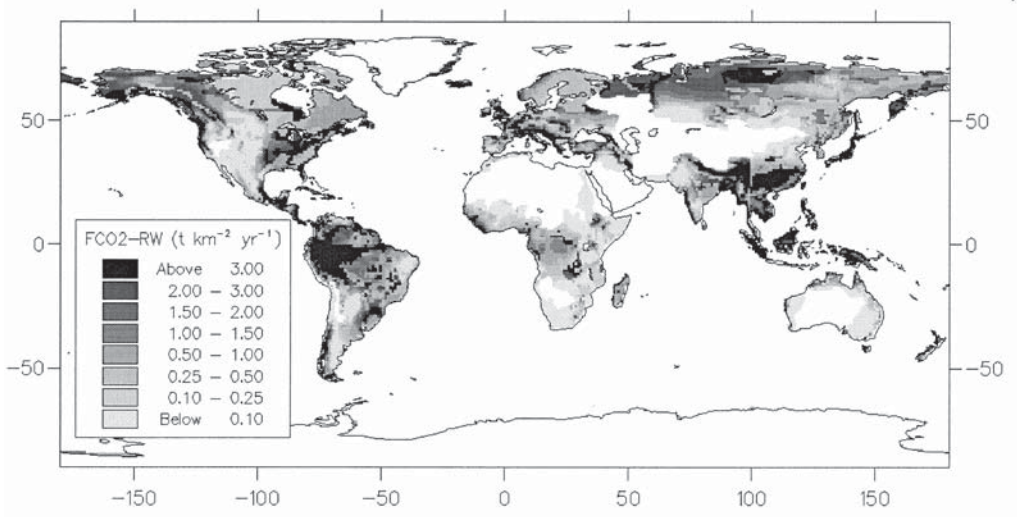The influence of lithology on F_{CO_2-RW} becomes evident at the continental and regional scale. For

example, in Africa the specific F_{CO_2-RW} is about five times smaller than in South America, while the specific drainage intensity is only three times greater in South America. In Africa, plutonic and metamorphic rocks as well as sandstones are abundant over large areas. Because these rock types consume small amounts of atmospheric CO_2, this continent has low specific F_{CO_2-RW} values. On the other hand, by far the greatest values for the specific F_{CO_2-RW} are observed in the south and south-east of Asia, where large carbonate outcrops coincide with a great drainage intensity. Among the major rock types, carbonates consume the greatest amount of atmospheric CO_2 (Amiotte-Suchet and Probst, 1993a). This makes Asia to be the continent with the highest specific F_{CO_2-RW}. We calculate that 22.7% of global

FDOC (t km^{-2} yr^{-1})

Above   3.00
2.00 — 3.00
1.50 — 2.00
1.00 — 1.50
0.50 — 1.00
0.25 — 0.50
0.10 — 0.25
Below   0.10

a)

FPOC (t km^{-2} yr^{-1})

Above   5.00
2.00 — 5.00
1.50 — 2.00
1.00 — 1.50
0.50 — 1.00
0.25 — 0.50
0.10 — 0.25
Below   0.10

b)

FCO2-RW (t km^{-2} yr^{-1})

Above   3.00
2.00 — 3.00
1.50 — 2.00
1.00 — 1.50
0.50 — 1.00
0.25 — 0.50
0.10 — 0.25
Below   0.10

c)

Table 2
Regional distribution of river carbon fluxes to the oceans

	Area, 10^3 km^2	(a) F_{DOC}, 10^{12} g yr^{-1}	(b) F_{POC}, 10^{12} g yr^{-1}	(c) $F_{CO_2\text{-RW}}$, 10^{12} g yr^{-1}	(d) F_{carb}, 10^{12} g yr^{-1}	% of (c), consumed on carbonates	Effects on (c + d) according to Eq. (5), %
Polar (without ice)	3892	3.1	1.7	3.4	1.5	45.3	0.0
Tundra and Taiga	23 232	45.9	18.9	33.5	9.2	27.5	191.0
Temperate dry	9635	2.9	10.9	4.4	2.1	46.9	555.0
Temperate wet	16 918	35.4	33.6	48.4	26.3	54.4	30.0
Tropical dry	21 790	16.0	18.2	15.1	5.8	38.4	0.0
Tropical wet	24 919	101.7	100.9	128.1	51.0	39.8	−15.0
Desert	5940	0.3	2.9	0.4	0.2	59.5	0.0
Total	106 326	205.2	187.1	233.3	96.2	41.2	34.2
Africa	18 288	20.0	16.3	11.7	5.2	44.5	−1.6
Europe	9564	16.6	11.0	18.8	7.8	41.3	90.2
North America	23 020	39.4	30.3	41.1	18.4	44.7	68.0
South America	17 732	51.8	44.0	53.6	8.5	15.8	−5.1
Asia	32 518	73.0	82.4	105.8	56.3	53.2	32.9
Australia	4476	3.9	2.7	2.2	0.1	2.1	28.1
Antarctis	728	0.6	0.5	0.1	0.0	0.0	−0.7
Total	106 326	205.2	187.1	233.3	96.2	41.2	34.2
Arctic Ocean	16 982	24.9	7.0	20.9	6.7	32.2	176.0
North Atlantic	27 300	71.0	47.8	72.9	24.2	33.2	20.5
South Atlantic	16 959	25.7	15.1	15.8	5.2	32.8	5.2
Pacific	21 025	57.0	75.2	80.7	34.1	42.3	21.6
Indian Ocean	16 594	21.4	36.1	33.4	21.0	63.0	28.9
Mediterranean	6739	4.6	5.4	9.5	4.9	51.4	74.6
Below 60° south	728	0.6	0.5	0.1	0.0	0.1	1.2
Total	106 326	205.2	187.1	233.3	96.2	41.2	34.2

Column 7 gives the percentage of $F_{CO_2\text{-RW}}$ consumed over carbonate rocks. Column 8 shows the increase or decrease (in %) of the fluxes in columns 4 and 5 after correction of possible climatic effects according to Eq. (5).

$F_{CO_2\text{-RW}}$ is discharged to the oceans from the part of Asia between 75° to 135°N longitude and 10° to 40° latitude, although this region covers only about 9.4% of the total exoreic continental area.

At the same time, the south and south-east of Asia is of similar importance for flux of particulate organic carbon because of great sediment fluxes typical for this part of the world. 23.3% of global F_{POC} is discharged from this region to the oceans. For F_{DOC}, this is only 13%. Neglecting the morphological effect, which tends to average out over large scales, dissolved organic carbon fluxes are principally a function of the variation of drainage intensity together with the variation of the soil carbon pool. Thus, for a constant Q, fluxes increase towards the northern regions with generally greater amounts of carbon in the soils. Therefore, in the tundra and taiga climate F_{DOC} is more important than the fluxes of other carbon forms (Ludwig et al., 1996a).

It is important to note that after a correction for climatic effects, the consumption of atmospheric CO_2 by rock weathering north of 30°N becomes almost twice as great (Fig. 4). In this case, F_{CO_2-RW} in the tropical wet climate clearly falls below the fluxes of

Fig. 3. Estimated river carbon fluxes (t km^{-2} yr^{-1}) to the oceans. Endoreic basins and glaciated regions are omitted: (a) F_{DOC}, (b) F_{POC}, and (c) $F_{CO_2\text{-RW}}$.
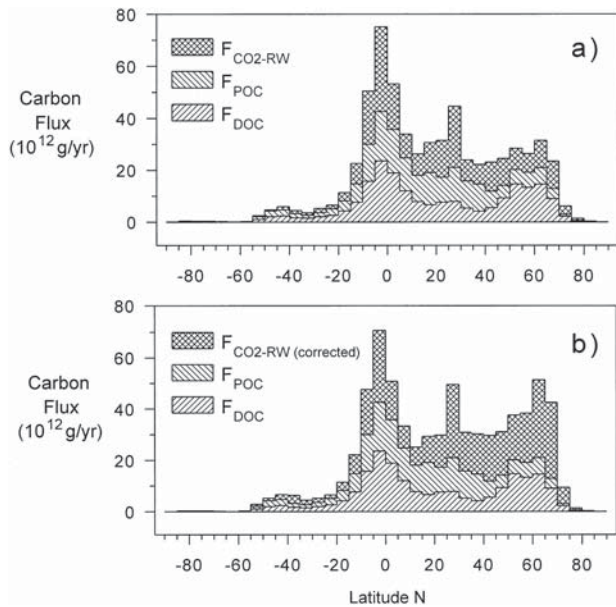
Fig. 4. Latitudinal distribution of river carbon fluxes to the oceans, before (a) and after (b) correction for possible climatic effects for $F_{CO_2\text{-}RW}$ according to Eq. (5).

the cooler northern climates (temperate wet and tundra and taiga). Without correction, 68.1% of F_{DOC}, 69.4% of F_{POC}, and 69.6% of F_{CO_2-RW} are consumed over the Northern Hemisphere. With correction, F_{CO_2-RW} increases to 76.9%, indicating that the total amount of atmospheric CO_2 consumption north of the equator may total about 0.50 Gt C/yr. At the same time, the fluxes of the dissolved carbon forms ($F_{DOC} + F_{CO_2-RW}$) become much more important towards the north compared to F_{POC}.

6. Implications for the role of continental erosion during the last glacial maximum

It has been proposed that increased erosion fluxes may have at least partly contributed to the low atmospheric CO_2 concentrations during the last glacial maximum (e.g., Munhoven and François, 1994). In the geochemical carbon cycle, the atmospheric CO_2 consumption resulting from carbonate weathering is normally balanced in relative short time scales by carbonate sedimentation in the oceans, where all CO_2 is released back to the ocean/atmosphere system (Eq. (3), from the right to the left). This is not the case for CO_2 consumed by silicate

weathering or by organic matter erosion, where the carbon that is lost by carbonate and organic matter sedimentation only returns to the ocean/atmosphere system via metamorphism/volcanism. A part of the organic carbon returns also to the atmosphere by the slow oxidation of old sediment carbon (kerogen) in sedimentary rocks (e.g., Kramer, 1994). Because the latter processes can vary considerably over geological time scales, this may result in large perturbations of atmospheric CO_2.

The most efficient way to increase erosion fluxes would be an increase of the continental area together with an increasing drainage intensity. Assuming that weathering is negligible under the extended ice sheets, the effective erodible continental area does not change much from LGM to present-day because the lost of area caused by ice extensions is more or less compensated by a greater exposure of shelf area when the sea level recedes (Fairbanks, 1989). An important point is, however, whether the assumption that there is no weathering taking place underneath ice sheets is correct or not. When analysing melt waters from a Swiss glacier, Sharp et al. (1995) have observed higher CO_2 consumption rates underneath the ice-cover than in a non-glaciated catchment. If this high consumption rate holds also for extended parts of the glacial ice sheets, the area actually subject to chemical weathering could have been considerably larger at LGM than it is today. Extensive chemical weathering could have occurred at least at the margins of ice sheets during glacial periods, where melt waters were in contact both with atmospheric CO_2 and large amounts of fine grained glacial debris. This may have especially affected silicate erosion, because shales, which show among the silicate rocks the greatest specific CO_2 consumption (Amiotte-Suchet and Probst, 1993a), are widely outcropping in the northern latitudes above 40°N.

Little is known about drainage intensity during LGM. Reconstruction of vegetation distribution based on palynological, pedological, and sedimentological evidence reveals globally a much greater aridity during LGM than today (Adams et al., 1990), indicating that this may also have had a lower river discharge as a consequence (Starkel, 1988). This is in contradiction with results from climate simulations done with General Circulation Models (GCM) under LGM boundary conditions. Such simulations

predict generally a too low reduction of land precipitation compared to the predicted decrease of land temperatures and, thus, decrease of evaporation (Lautenschlager and Herterich, 1990), leading to an unchanged or even greater amount of water running off the continents (e.g., Gibbs and Kump, 1994; Munhoven and Probst, 1995). The question to what amount the continental runoff may have been different during LGM compared to present-day must therefore be left open, but answering it is crucial in order to estimate possible changes in the global carbon cycle related to continental erosion, as we have shown in the previous sections of this paper.

The next important factor is global lithology. Even without changing the total amount of runoff, one may end up in an increased net CO_2 consumption if climate zones with high drainage intensity shift considerably and encompass more and more highly erodible silicate rocks. Since one can suppose that also for the LGM climate, drainage intensity should have been greatest in the low latitudes, the changes in the lithological distribution related to the emerging continental shelves may have been more important than the changes in lithology due to the ice sheet extension. Assuming that continental shelves consist largely of carbonates, it is possible that $F_{CO_2\text{-RW}}$ may have been somewhat enhanced compared to present-day on the whole, but there is no evidence for much greater fluxes from silicate weathering related to changes in lithology. To confirm such a hypothesis, however, a detailed knowledge of the lithological character of the shelf sediments is needed.

One has to look at possible changes in the fluxes of organic carbon. F_{DOC} is related to the amount of carbon stored in the soils, indicating lower fluxes during the LGM compared to present (not considering here, of course, the open question about a possible increase in the amount of total runoff). Taking, for example, the estimate of Adams et al. (1990) that the global soil carbon pool during LGM was reduced by about 770 Gt, we result with Eq. (6) in a global reduction of F_{DOC} of about 19% under the assumption that Q and slope were the same as the present-day values. Since also POC can be supposed to originate mainly from soil carbon (Ludwig et al., 1996a), the effect for F_{POC} should go in the same direction.

Finally, one should also to mention here that it is not only the amount of carbon discharged by rivers but also the oceanic response to this carbon input that determines the role of continental erosion in the glacial/interglacial carbon cycle. For the present-day cycle of organic carbon in the oceans, Smith and Hollibaugh (1993) found that about equal amounts of the riverine input were involved in the coastal cycle and in the open ocean cycle, but the ratio of burial to respiration in the coastal zone may be about 6 to 4, while it is only about 1 to 9 in the open ocean. If this can be extrapolated to LGM, one may conclude that due to the reduced shelf area, greater amounts of riverine organic matter may have reached the open ocean, leading to much lower sedimentation and greater respiration rates in the ocean compared with today. Even if the river inputs were lower during LGM, the amount of carbon that returned to the atmosphere may not have changed or it became even greater than for present day.

In this context, there is also another point of interest: Franzén (1994) proposed that peat deposits of late Eem-interglacial may have been as great as 250–550 Gt C in the regions that were affected by the ice sheets during LGM. Because actually there is no evidence for the existence of large peat deposits at LGM, he supposed that nearly all of these deposits have been destroyed and washed out by melt waters during the retreat of the ice masses at the end of LGM. This should have led to a considerable peak in organic carbon erosion, and a large amount of this carbon may have ended up in the oceans. Oceanic respiration of this carbon could have accounted for at least a part of the increase of the atmospheric and terrestrial carbon pools during the Holocene.

7. Conclusions

With the present study, we propose a modelling tool to predict the consumption of atmospheric CO_2 by continental erosion at the global scale. This approach includes both the CO_2 consumption resulting from rock weathering and from the erosion of organic matter. At present-day, continental erosion represents a carbon transfer of about 0.6 to 0.7 Gt C/yr from the atmosphere to the oceans. High levels of

CO_2 consumption in the Northern Hemisphere coincide with the net sink of atmospheric CO_2 on the continents that has been postulated by Tans et al. (1989) because of the inconsistency of the observed latitudinal CO_2 gradient in the atmosphere with the transport fields generated by a general circulation model. About 0.5 Gt C/yr are consumed north of the equator, accounting for about 15 to 25% of the sink calculated by Tans et al. (1989). An important question in order to evaluate the role of these fluxes in the global carbon cycle is the oceanic response to the river carbon input. It is especially not clear how much of this carbon is trapped in the estuaries and coastal zones, either by sedimentation or by returning to the atmosphere through respiration processes (Kempe, 1995; Ludwig et al., 1996b).

Our approach may be further applied to predict the response of continental erosion in the glacial/interglacial carbon cycle change since the last glacial maximum. At present, it is difficult to estimate such a response because there is insufficient information about the hydroclimatic situation during LGM. It is especially not clear to what extent the amount of global runoff changed, which is, as our results imply, the most important controlling factor for present-day. Another open question concerns the rates of chemical weathering that may take place underneath and at the margins of ice sheets. Finally, more work has to be devoted to study the influence of climate on rock weathering. We present indications that, for a given drainage and a given lithology, HCO_3^- concentrations are greater on average in dry and cold climates than in wet and hot ones. If this is right, it may also have influenced weathering during LGM.

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