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Holocene biomass burning and global dynamics of the carbon cycle

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Abstract

Fire regimes have changed during the Holocene due to changes in climate, vegetation, and in human practices. Here, we hypothesise that changes in fire regime may have affected the global CO_2 concentration in the atmosphere through the Holocene. Our data are based on quantitative reconstructions of biomass burning deduced from stratified charcoal records from Europe, and South-, Central- and North America, and Oceania to test the fire-carbon release hypothesis. In Europe the significant increase of fire activity is dated ≈6000 cal. yr ago. In north-eastern North America burning activity was greatest before 7500 years ago, very low between 7500–3000 years, and has been increasing since 3000 years ago. In tropical America, the pattern is more complex and apparently latitudinally zonal. Maximum burning occurred in the southern Amazon basin and in Central America during the middle Holocene, and during the last 2000 years in the

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northern Amazon basin. In Oceania, biomass burning has decreased since a maximum 5000 years ago. Biomass burning has broadly increased in the Northern and Southern hemispheres throughout the second half of the Holocene associated with changes in climate and human practices. Global fire indices parallel the increase of atmospheric CO_2 concentration recorded in Antarctic ice cores. Future issues on carbon dynamics relatively to biomass burning are discussed to improve the quantitative reconstructions.

Keywords: Fire; Atmospheric carbon dynamics; Europe; North America; South America; Southeast Asia

1. Introduction

Carbon dioxide is one of the primary gases involved in the "greenhouse effect". It is thus essential to know the global dynamics of CO_2, both in terms of causes and consequences of variations through time. A high-resolution analysis of gas bubbles in Antarctic ice cores has demonstrated that global atmospheric CO_2 concentration has changed during the Holocene (Indermühle et al., 1999). One explanation of this CO_2 rise involves changes in the terrestrial tropical vegetation of Africa and Arabia, specifically the expansion of desert in response to drier and cooler conditions since 7000 years (Indermühle et al., 1999). However, at the same time in South America, tropical forests expanded at the expense of savannah (Desjardins et al., 1996; Salgado-Labouriau et al., 1997; de Freitas et al., 2001) in response to increasing moisture (Martin et al., 1997; Ledru et al., 1998). The carbon released from changes in African and Arabian vegetation were potentially compensated by the build-up of biomass in South America. Nevertheless, the estimate of carbon release by African and Arabian vegetation only represents 30 Gt, while the difference between 7000 and 1000 cal. yr BP represent 195 Gt, suggesting that changes in the vegetation of these regions are not enough to explain the atmospheric carbon rise (Indermühle et al., 1999). If the biospheric hypothesis of Indermühle et al. (1999) is correct, regions in addition to tropical areas of Africa and Arabia must have contributed to the rise in global atmospheric CO_2.

Among the processes of carbon fluxes, biomass burning is one of the most important (Seiler and Crutzen, 1980). Natural or human-induced fires affect most ecosystems from tropical forests to tundra (Bond and van Wilgen, 1996; Dwyer et al., 2000). Fire is thus a global process that depends on climate, volcanic activity, vegetation, and human practices. Paleoecological studies show that local fire regimes are not constant through time and vary at the decadal, centennial, and millennial time-scales (Clark, 1990; Swetnam, 1993; Long et al., 1998; Carcaillet and Richard, 2000). Although temporal variability has been documented, very little is known about the spatial variability and the carbon release to the atmosphere by burning of paleo-biomass, despite several regional syntheses (see Clark et al., 1997).

In this paper, we hypothesise that fire may play a role in the global carbon budget, particularly through the release of carbon gas to the atmosphere related to global fluctuations in biomass burning. To test this hypothesis, we show synthetic charcoal records from areas where several records are available, e.g. Europe, northeastern North America, tropical South and Central America, and Oceania.

2. Material and methods

2.1. Stratigraphic charcoal

2.1.1. Rationale and assumptions

The analysis of charcoal from lake and peat sediments is suitable for temporal reconstruction of biomass burning because of the stratigraphic nature of these deposits (Clark et al., 1997). Charcoal accumulation rates (CHAR) are a generally accepted proxy for the occurrence of fire (Clark et al., 1996b), although the relationship between products of biomass burning, such as charcoal, and the stratigraphic record is highly complex (Cofer et al., 1997). We assume that CHAR depicts the burning activity and is correlated with the carbon released by fires from biomass, regardless of the fire origins, natural or anthropogenic.

Several techniques are used to estimate charcoal (review in Patterson et al., 1987 and in Rhodes, 1998). They often involve tallying or measuring microscopic charcoal (<100 μm) in pollen-slides, and others methods estimate macroscopic charcoal (>200 μm) by sieving sediment. These two approaches do not provide the same burning signal. Micro-charcoal from pollen-slide is considered a proxy of both local (few hundred meters from the lake shore) and regional burning activity (Tinner et al., 1998; Carcaillet et al., 2001b), whereas sieved macro-charcoal represents stand- to local-scale fire history depending on sampling site area (Bradshaw and Hannon, 1992; Clark et al., 1998; Carcaillet et al., 2001b). This dichotomy is based on the fact that sieved charcoal has a low risk of being transported far in the air, while pollen-slide charcoal fragments are airborne transported (Wein et al., 1987; Clark et al., 1998; Ohlson and Tryterud, 2000). Water runoff can affect the charcoal record especially in large catchment areas covered by rivers and steep relief (Earle et al., 1996; Whitlock and Millspaugh, 1996; Laird and Campbell, 2000).

2.1.2. Source of data

The data used in this study are from several sources, including time series of pollen-slide and sieved charcoal from peat and lake sediments. Most of data are published (Table 1). Because of differences in charcoal quantification methods, some series display local and others show regional fire activity. The synthesis of all of the data per region provides a regional view of biomass burning. The selected charcoal series have more than four samples per millennium. All data are expressed in CHAR, e.g. $mm^2\,cm^{-2}\,yr^{-1}$, $mm^2\,g^{-1}\,cm^{-2}\,yr^{-1}$ or fragments $cm^{-2}\,yr^{-1}$ depending on the original data. Data cover at least 6000 years, and some cover more than the entire Holocene (11 000 cal. yr) including the Late-Glacial and the Full-Glacial. One data set comes from the cold temperate and boreal modern ecosystems of eastern Canada. A second set is from the Mediterranean to the boreal biomes in Europe, including sites from temperate plains and mountains. The third group of data are from studies in tropical areas of South and Central America. The last group is from Oceania, i.e. Indonesia and Papua New Guinea (Fig. 1; Table 1).

2.1.3. Calculation of the charcoal anomalies

Charcoal influx values vary greatly from site to site because of site size, physiography of catchment area, sedimentation rate, vegetation type and landscape structure (i.e. plant flammability and combustibility, fuel load and connectivity), and fire regime. Because intersite variability is important, we focused our analysis on long-term trends in CHAR through calculation of charcoal anomalies (Carcaillet and Richard, 2000). Here, the anomalies have been calculated first by the normalisation of charcoal series, i.e.:

$$N_{\mathrm{CHAR}} = \frac{\mathrm{CHAR}_i - \mu_s}{\sigma_s} \tag{1}$$

where $[\mathrm{CHAR}_i]$ corresponds to the CHAR at the ith spectra, and $[\mu_s]$ and $[\sigma_s]$ are the mean and the standard deviation of postglacial CHAR at the site, respectively. Normalisation allows the comparison of the magnitude of change in CHAR and is independent of site-specific taphonomical processes. Each charcoal series is displayed independently from the other with a comparable scale to allow an estimation of the real magnitude of changes (Fig. 2a and b).

Normalised charcoal series are averaged per millennium (Fig. 2c). Each millennium is centred on a *kilo annum*, e.g. 6 ka corresponds to time elapsed between 6500 and 5500 calibrated years before present (cal. yr BP). 0 ka represents the last 500 years. While Carcaillet and Richard (2000) excluded postglacial non-forest sequences for each site to reconstruct the variability of forest fire history in eastern Canada, the entire charcoal series since deglaciation is used here because our purpose is to reconstruct the biomass burning regardless of vegetation type. The anomalies are calculated for all millennium with a minimum of four charcoal data. The sequence of anomalies at a given site can thus be incomplete.

2.1.4. Carbon-release index

The carbon-release status corresponds to the mean of charcoal anomalies of all sites at a given millennium in a given area, such as eastern Canada or Europe. A negative index corresponds to a low carbon-release resulting from a low biomass burning activity, and vice versa for a positive index (Fig. 2d).

The variability in charcoal records varies greatly from series to series. Some sites contain 3–4 charcoal samples per 1000 years, while some others have up to 75 samples per 1000 years. For the calculation of carbon-release index the high-resolution series are given a greater weight than low-resolution series. Series with $\leqslant 5$ charcoal data-points per 1000 year have a weight of 0.5, a weight of 1 is assigned to series containing 6–10 data-points per 1000 years, 2 for series with 11–20 data-points per 1000 year, and 4 for series with >20 data-points per 1000 year.

2.2. Soil charcoal dating

2.2.1. Rationale and assumptions

Soil charcoal is found in all vegetation types except polar (e.g. Bryson et al., 1965; Jacquiot et al., 1973; Vernet et al., 1994). Soil charcoal has long been used to reconstruct fire history based on ^{14}C dating of fragments (Payette and Gagnon, 1985; Carcaillet, 1998). Generally, charcoal resulting from Holocene fires is buried in soils by bioturbation (Carcaillet, 2001a,b) or geomorphologic processes such as aeolian activity (Payette and Filion, 1993). Although physical and biotic processes weather the charcoal fragments, resulting a decreasing particle size with time (Carcaillet and Talon, 1996), dating of soil charcoal demonstrates that it is preserved in soil for tens of millennia (Servant et al., 1989; Hopkins et al., 1993).

If charcoal fragments are randomly sampled for dating in a soil profile (Carcaillet, 2001a), they provide a suitable proxy for reconstruction of fire history, and thus an indirect proxy of potential carbon released to the atmosphere by biomass burning.

2.2.2. Source of data

In the present synthesis, data have been compiled from many papers dealing with mountain and

Table 1
Characteristics of sites (location, methods, and references) for stratified charcoal series

Site	Country	Charcoal method	Site type	Latitude	Longitude	Elevation (m a.s.l.)	Size (ha)	Modern biome (F = forest)	Modern climate	First reference
Europe										
Raigejegge	Sweden	Sieving	Lake	66°06′ N	18°13′ E	485	<1	Evergreen needleleaf F	Boreal	Carcaillet et al., unpublished
Lattok	Sweden	Sieving	Lake	65°57′ N	18°21′ E	460	1	Evergreen needleleaf F	Boreal	Carcaillet et al., unpublished
Sävkrars mosse	Sweden	Sieving	Peat			65		Evergreen needleleaf F	Boreal	Von Stedingk, 1999
Lilla Gloppsjon	Sweden	Pollen-slide/ sieving	Lake	59°48′ N	14°37′ E	198	50	Evergreen needleleaf F	Boreal	Almquist-Jacob-son, 1994
Ljustjarnen	Sweden	Pollen-slide/ sieving	Lake	59°45′ N	14°29′ E	183	10	Evergreen needleleaf F	Boreal	Almquist-Jacob-son, 1994
Elferdalen	Norway	Sieving	Peat	59°39′ N	9°18′ E	360–400		Evergreen needleleaf F	Boreal	Tryterud, 2000
Bohult	Sweden	Sieving	Peat	57°14′ N	16°10′ E	?		Evergreen needle-leaf × deciduous broadleaf F	Cold temperate	Bradshaw et al., 1997
Solsø	Denmark	Pollen-slide	Lake	56°80′ N	8°37′ E	41		Deciduous broadleaf F	Cold temperate	Odgaard, 1994
Skånsø	Denmark	Pollen-slide	Lake	56°31′ N	8°50′ E	8	12	Deciduous broadleaf F	Cold temperate	Odgaard, 1994
Kragsø	Denmark	Pollen-slide	Lake	56°15′ N	9°06′ E	47		Deciduous broadleaf F	Cold temperate	Odgaard, 1994
Kis-Mohos Tó	Hungary	Pollen-slide	Peat	48°24′ N	20°24′ E	310		Deciduous broadleaf F	Temperate	Willis et al., 1997
Steeren-moos	Germany	Pollen-slide	Peat	47°49′ N	8°11′ E	1000	~12.5	Evergreen needle-leaf × deciduous broadleaf F	Mountain temperate	Rösch, 2000
Nussbau-mersee	Switzerland	Pollen-slide/ sieving	Lake	47°36′ N	8°49′ E			Deciduous broadleaf F	Temperate	Haas and Hadorn, 1998
Seedorf See	Switzerland	Pollen-slide	Lake	–	–	609	13	Evergreen need-leaf × deciduous broadleaf F	Temperate	Richoz et al., 1994
Lago di basso	Italy	Sieving	Lake	46°25′ N	9°17′ E	2250		Evergreen needleleaf F	Mountain temperate	Wick and Tinner, 1997
Origlio	Switzerland	Pollen-slide	Lake	46°02′ N	8°57′ E	416	8	Deciduous broadleaf F	Warm temperate	Tinner et al., 1999
Muzzano	Switzerland	Pollen-slide	Lake	46°00′ N	8°56′ E	337	22	Deciduous broadleaf F	Warm temperate	Tinner et al., 1999
Las Pardil-las	Spain	Pollen-slide/ sieving	Lake	42°20′ N	3°20′ W	1850	<1	Evergreen needleleaf F	Mountain Mediterranean	Sánchez Goñi and Hannon, 1999

Ojos del Tremendal	Spain	Pollen-slide	Peat	40°32′ N	2°03′ W	1650	60	Deciduous broadleaf × evergreen needleleaf F	Mountain Mediterranean	Stevenson, 2000
Navarés	Spain	Pollen-slide/ sieving	Peat	39°06′ N	0°41′ W	225	?	Evergreen broadleaf F	Mediterranean	Carrı́on and Van Geel, 1999
North America										
LR1	Québec (Canada)	Pollen-slide	Lake	58°35′ N	75°15′ W	170	2	Tundra	Sub-polar	Gajewski et al., 1993
LB1	Québec (Canada)	Pollen-slide	Lake	57°55′ N	75°37′ W	200	3	Tundra × Evergreen needleleaf F	Boreal	Gajewski et al., 1993
EC1	Québec (Canada)	Pollen-slide	Lake	56°17′ N	75°06′ W	250	4	Tundra × Evergreen needleleaf F	Boreal	Gajewski et al., 1993
GB2	Québec (Canada)	Pollen-slide	Lake	55°06′ N	75°17′ W	300	5	Evergreen needleleaf F	Boreal	Gajewski et al., 1993
Pessière	Québec (Canada)	Sieving	Lake	49°30′ N	79°14′ W	280	4	Evergreen needleleaf F	Boreal	Carcaillet et al., 2001b
Triangle	Québec (Canada)	Pollen-slide	Lake	48°42′ N	65°24′ W	465	2	Evergreen needleleaf F	Boreal	Asnong, 2000
Petit Bouchard	Québec (Canada)	Pollen-slide	Lake	48°51′ N	64°35′ W	145	2	Evergreen needleleaf × deciduous broadleaf F	Boreal	Asnong, 2000
Pas-de-Fond	Québec (Canada)	Sieving	Lake	48°48′ N	78°50′ W	290	2	Evergreen needleleaf × deciduous broadleaf F	Boreal	Carcaillet et al., 2001b
Francis	Québec (Canada)	Pollen-slide/ sieving	Lake	48°31′ N	79°28′ W	305	<1	Evergreen needleleaf × deciduous broadleaf F	Boreal	Carcaillet et al., 2001b
Mirabel	Québec (Canada)	Pollen-slide	Peat	45°41′ N	74°03′ W	75	>10	Deciduous broadleaf F	Cold temperate	Müller, 2001
Synthesis: 30 sites	Québec/ Ontario (Canada)	Pollen-slide	Lake	45° to 55° N	64° to 80° W	17–965	<10	Deciduous broadleaf to Evergreen needleleaf F	Cold temperate to subarctic	Carcaillet and Richard, 2000
South America										
Synthesis: 5 sites	Equador Panama Brazil	Pollen-slide	Lake, peat, swamp	1° S to 9° N	47° to 81° W	35–3180	<100	Evergreen broadleaf F	Tropical	Haberle and Ledru, 2001
Oceanie										
Synthesis: 10 sites	Papua New-Guinea Indonesia	Pollen-slide	Lake, peat, swamp	1° S to 9° S	105° to 147° W	100–3630	<100	Evergreen broadleaf F	Tropical	Haberle and Ledru, 2001

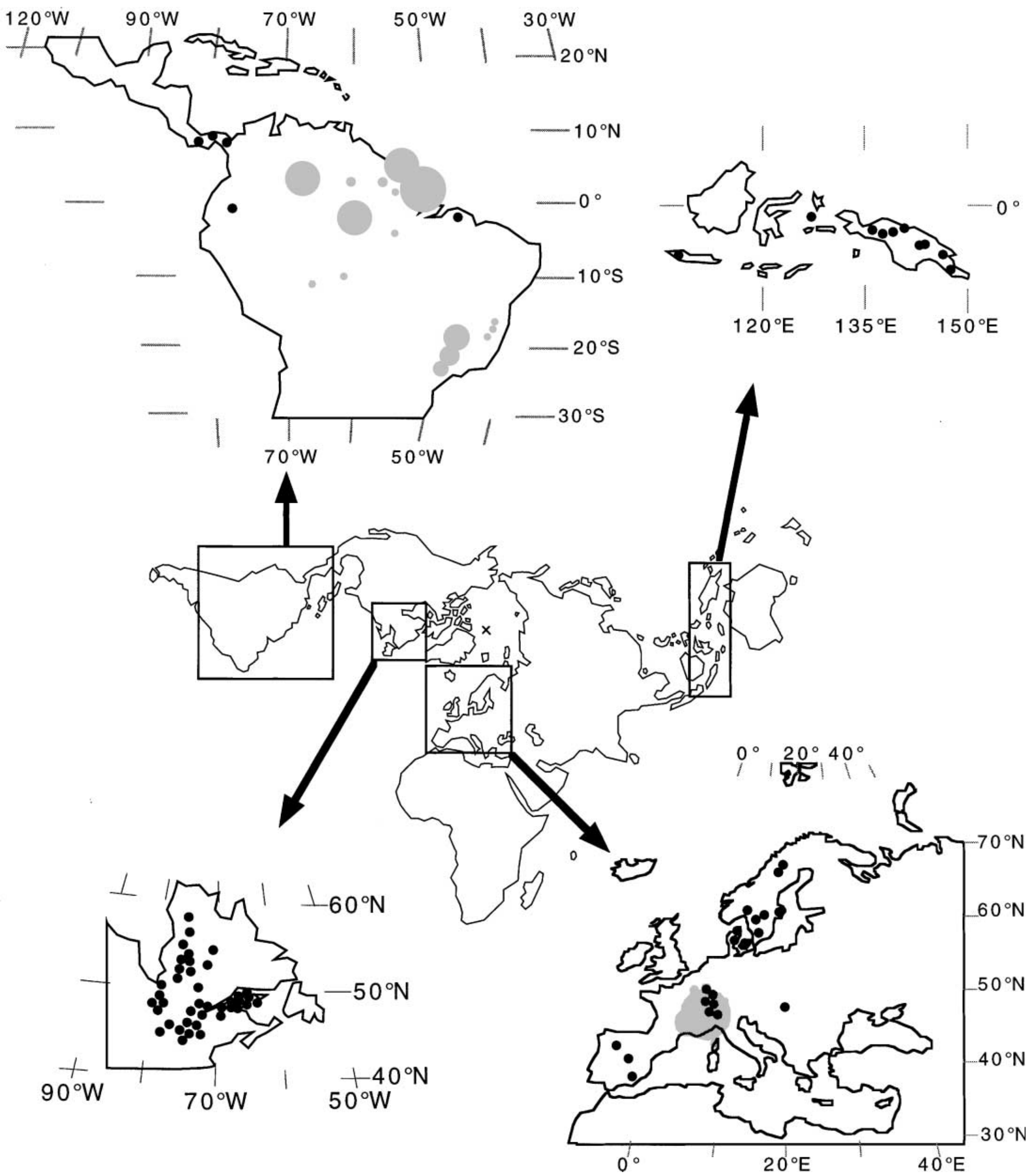

Fig. 1. Location of study sites. Stratigraphic charcoal series are indicated by black dots and, soil charcoal ^{14}C data set is by grey dots. Size of grey dots is proportional to the number of ^{14}C dates on soil charcoal.

Mediterranean soils in Europe, and tropical soils in South America (Fig. 1; Table 2). In Europe 95 ^{14}C dates were recorded (mostly in the Alps), and 225 in South America (mostly in the Brazil and French Guyana).

2.2.3. Fire history

No numerical treatment has been carried out on soil charcoal dating. The fire history is based on the distribution of calibrated ^{14}C dates on charcoal particles without taking into account the associated analytical error. Dates are distributed into age classes of 500 calibrated years.

3. Results and interpretation

3.1. Biomass burning in Europe

The mean pattern of charcoal anomalies shows very little burning during the early and middle Holocene

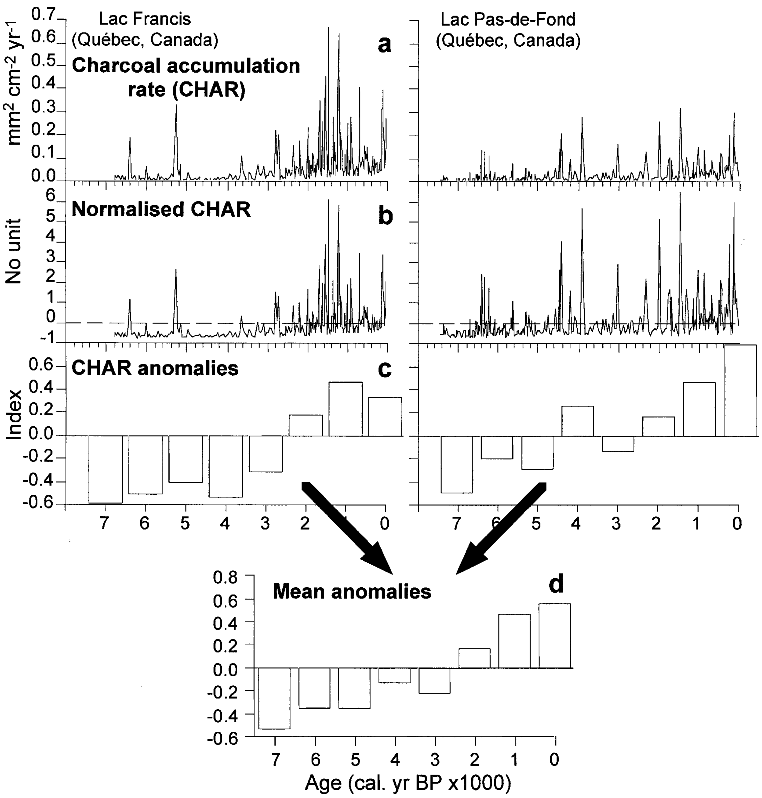

Fig. 2. Method to calculate the charcoal index based on charcoal time series. Here, the two charcoal series come from Abitibi (boreal Québec), south of James Bay.

(11 000 to ≈3000 cal. yr BP) and an increase in burning from the middle to late Holocene (Fig. 3a). The pattern of increasing burning is temporally gradual. The first rise occurs between 10 and 8 ka, and the second between 6 and 3 ka. Two maxima are observed at 3 and at 0 ka. In the past 3500 years, the period between 1500 and 500 cal. yr BP appears, on average, as a period of low biomass burning in Europe. The distribution of charcoal dates in Alpine soils shows a progressive increase since 7000 years ago, but fires are recorded since the beginning of the Holocene (Fig. 3b). The two main periods with abundant soil charcoal occur between 5500 and 3000 cal. yr BP, and since 2000 years (Fig. 3b).

The timing differences in charcoal maxima in soil and lakes and wetland records may be explained by their geographic location. Indeed, most soil charcoal studies have been carried out in areas where no lacustrine charcoal series exist, i.e. French Alps, Provence and Corsica (Table 1). On the other hands, most lake records are located in the northern (Scandinavia) and central (Germany, Switzerland) Europe. However, the two proxies of biomass burning, soil charcoal dates and stratigraphic charcoal, show a similar pattern of increasing burning activity since 8 to 7 ka.

In Europe, climatic changes of the last 11 000 years, affected, for instance, upper treeline (Kullman and Kjällgren, 2000), lake-level (Harrison et al., 1993), diatoms (Korhola et al., 2000), or molluscs assemblages (Rousseau et al., 1994). In addition, the development of the cultural landscape has been a major process during the Holocene (Berglund, 1991). Neolithic populations expanded between 7000 and 6000 cal. yr BP followed by Bronze- and Iron-Age civilisations. Neolithic and metal age populations deforested most of the continent, probably using the slash-and-burn cultivation system (e.g. Clark et al., 1989; Tinner et al., 1999; Pitkänen, 2000). This increasing human influence on European ecosystems drove expansion of evergreen sclerophyllous trees in the Mediterranean basin (Vernet, 1997; Pons and Quézel, 1998; Willcox, 1999) and conifers in cold temperate forests (Björkman and Bradshaw, 1996) broadly after 6 ka, but with some times more than 3000 years of temporal differences between sites. It favoured expansion of grasslands at high elevations (Tinner et al., 1996; Carcaillet et al., 1998; Talon et al., 1998), and heathlands in Atlantic regions (Odgaard, 1992). Consequently, increasing biomass burning since 6000 years ago is best explained by cultural development.

Table 2
Characteristics of sites for soil charcoal analysis

Site	Country	Number of ^{14}C dates	Latitude	Longitude	Elevation (m a.s.l.)	Modern biome (F = forest; G = grassland)	Modern climate	First reference
Europe								
Grand Ventron	France	2	?	?	1000	Evergreen needleaf × deciduous broadleaf F	Temperate	Carcaillet, unpublished
Alpe d'Esserte	Switzerland	3	46°09′ N	7°22′ E	1780–2380	Evergreen needleleaf F	Mountain temperate	Tinner et al., 1996
Intragna	Switzerland	4	46° N	9° E	530	Needleaf × deciduous broadleaf F	Warm temperate	Berli et al., 1994
Maurienne	France	44	45°15′ N	6°30′ E	1700–2700	Evergreen needleleaf F to alpine G	Mountain temperate	Carcaillet, 1998, 2001b
Southern Alps	France	23	44° to 45° N	6°30′ to 7°30′ E	1900–3000	Evergreen × deciduous needleleaf F to alpine G	Mountain temperate	Talon, 1997
Provence	France	12			0–850	Evergreen broadleaf F	Mediterranean	Thinon, 1992
Fango	Corsica France	7	42°20′ N	8°49′ E	600–900	Evergreen broadleaf F	Mediterranean	Carcaillet et al., 1997
South America								
S. Carlos de Rio Negro	Venezuela Columbia	40	1°56′ N	67°03′ W		Evergreen broadleaf F	Wet tropical	Sandford et al., 1985; Saldarriaga and West, 1986
Nouragues	French Guyana	37	4°05′ N	52°40′ W		Evergreen broadleaf F	Wet tropical	Tardy, 1998
Sinnamary	French Guyana	31	5°00′ N	53°00′ W		Evergreen broadleaf F	Wet tropical	Tardy, 1998
Marouini	French Guyana	1	2°20′ N	54°20′ W		Evergreen broadleaf F	Wet tropical	Tardy, 1998
Zone cotière	French Guyana	8	5°30′ N	53°00′ W		Evergreen broadleaf F	Wet tropical	Tardy, 1998
Boa Vista	Brazil	7	2°52′ N	60°32′ W		Semi-deciduous broadleaf F × G	Seasonal tropical	Desjardins et al., 1996
Km 41 Reserve	Brazil	31	2°30′ S	60°00′ W		Evergreen broadleaf F	Wet tropical	Santos et al., 2000
Transamazonian	Brazil	5	5°00′ S	56°10′ W		Evergreen broadleaf F	Wet tropical	Soubiès, 1980
Pimenta Bueno	Brazil	2	11°49′ S	61°10′ W		Semi-deciduous broadleaf F × G	Seasonal tropical	Pessenda et al., 1998
Vilhena	Brazil	2	12°42′ S	66°07′ W		Semi-deciduous broadleaf F × G	Seasonal tropical	Pessenda et al., 1998
Malachacheta	Brazil	1	17°40′ S	42°07′ W		Evergreen broadleaf F	Wet tropical	Servant et al., 1989
Santa Maria	Brazil	1	18°10′ S	42°18′ W		Evergreen broadleaf F	Wet tropical	Servant et al., 1989
Contagem	Brazil	4	18°25′ S	43°40′ W		Evergreen broadleaf F	Wet tropical	Servant et al., 1989
Salitre de Minas	Brazil	23	19°00′ S	46°46′ W		Semi-deciduous broadleaf F × G	Seasonal tropical	Boulet et al., 1995; Vernet et al., 1994
Jaguariúna	Brazil	17	22°40′ S	47°10′ W		Semi-deciduous broadleaf F	Seasonal tropical	Gouveia and Pessenda, 2000; Gouveia et al., 1999
Botucatu	Brazil	13	23°00′ S	48°00′ W		Semi-deciduous broadleaf F	Seasonal tropical	Gouveia et al., 1999

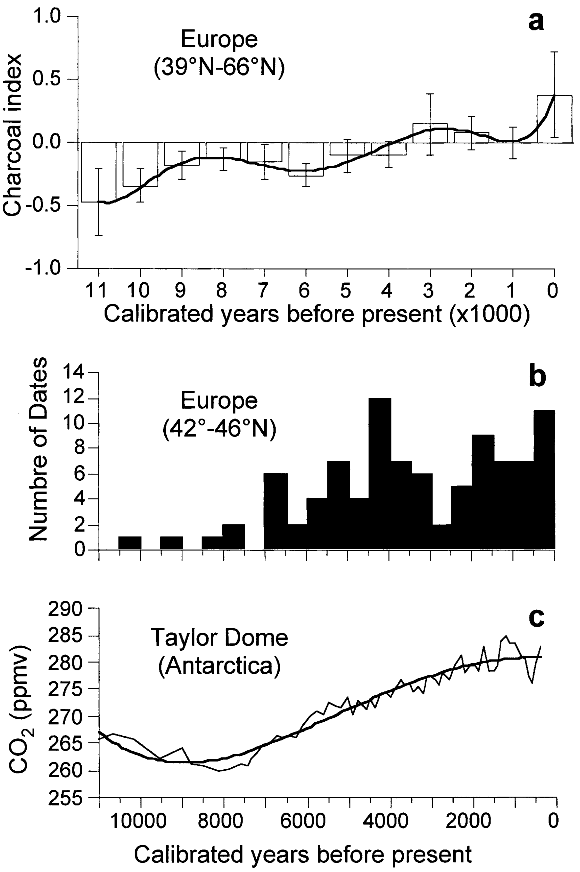

Fig. 3. Biomass burning reconstruction in Europe from the Mediterranean to upper Boreal ecosystems. (a) Charcoal index deduced from stratigraphic time series; error bars correspond to the standard error. (b) Distribution of ^{14}C dates in soil charcoal. (c) CO_2 concentration curve from Indermühle et al. (1999).

The rise of biomass burning between 10 and 8 ka, however, is more difficult to explain. Mesolithic populations may have used fire as a tool to create areas for game (Wick, 1994; Haas, 1996; Tipping, 1996). However, natural influences cannot be ruled out as pine forests were dominant in central and northern Europe, and significant climate warming occurred at the beginning of the Holocene. Regardless of whether fires were caused by Mesolithic peoples or by climate and vegetation conditions, biomass burning was low during the early Holocene, releasing very little carbon to the atmosphere.

If human influence on biomass burning is certain since $\approx$6000 years ago, it remains equivocal between 9000 and 8000 years ago, during early Mesolithic times. Climate effects on biomass burning in Europe are more difficult to determine because of the agricultural development since the Neolithic and the metal ages. Drier climate in Mediterranean areas since 6000 years ago (Harrison et al., 1993; Cheddadi et al., 1997) and positive feedback of highly flammable Mediterranean trees that expanded at the same time (Jalut et al., 2000) may have contributed to an increase in the risk of fire associated with human development.

3.2. Biomass burning in northeastern North America

Four main fire periods are shown by the anomalies of stratigraphic charcoal (Fig. 4a). The first period is at 11 ka (11 000 ± 500 cal. yr BP) and corresponds with low biomass burning index (Fig. 4a). The transition is abrupt with the second characterised by high burning index and occurs around 10.5 ka. The second period lasted until 8 ka. The third period between 7 and 3 ka corresponds with a low burning index. Since 3 ka the burning-index shows a progressive increase with a maximum during the last 500 years (0 ka).

The lowest burning activity is between 7 and 4 ka, and an increase occurred since 3 ka. Radiocarbon dates on charcoal from soil and paleo-sols in Quebec (not shown) indicate a slow and irregular rise in burning activity since 6500 cal. yr BP, but the maximum during the past 2000 years (Payette and Gagnon, 1985; Filion et al., 1991; Payette and Morneau, 1993; Bussières et al., 1996; Lavoie and Payette, 1996). In this area, woodland became established around 7000 cal. yr after the retreat of the Laurentide ice-sheet and the proglacial lakes,

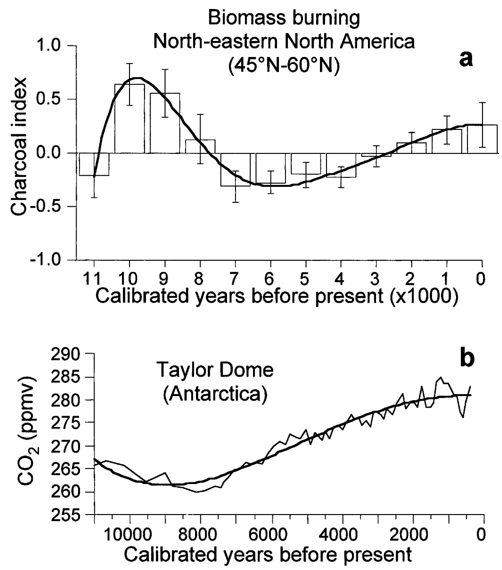

Fig. 4. Biomass burning reconstruction in north-eastern North America for temperate and boreal ecosystems. (a) Charcoal index deduced from stratigraphic time series; error bars correspond to the standard error. (b) Distribution of ^{14}C dates in soil charcoal. (c) CO_2 concentration curve from Indermühle et al. (1999).

explaining why no fires are recorded before that time. However, soil charcoal and stratified charcoal series indicate the same pattern of increasing biomass burning activity and associated carbon release during the late Holocene (Fig. 4).

The fire history in northeastern North America is associated with both climate and vegetation change (Clark et al., 1996a; Carcaillet and Richard, 2000). Although the presence of human is evidenced as early as 11 000 cal yr BP in northeastern North-America (Boisvert, 1999), the lack of agriculture among native peoples during most of the Holocene (Coté, 1993; Chapdelaine, 1996) meant that human activities were probably not a significant source of fire ignition. However, the southern edge of the area covered by fire data in the Saint-Laurent river valley was occupied by hunter-cultivators (Iroquoian civilisation) for several hundreds years before European settlement (McAndrews and Boyko-Diakonow, 1989; Cossette, 1996). Consequently, native American impacts on biomass burning cannot be totally ruled out for the past 2000 years in southern Québec and Ontario (Clark and Royall, 1995), although it was probably not of the same order of magnitude as that in Europe.

The first significant period in northeastern North America at 11 ka (Fig. 4a) corresponds to the tundra period with no-fire at the southern edge of the Laurentide ice sheet. The second and most significant period of biomass burning between 10 and 8 ka is associated with tree colonisation and forest expansion in the southern part of northeastern North America (Richard, 1994). The Laurentide ice-sheet completed its collapse around 7 ka (King, 1985; Lauriol and Gray, 1987) driving changes in atmospheric circulation, from an anti-cyclonic dominated system toward a more zonal climate controlled by westerlies and oceanic and air mass circulation (Sawada et al., 1999). Furthermore, conifers, which are more fire sensitive than broad-leaved species, dominated the vegetation during most of the early Holocene (Richard, 1994). A synchronic change in fire frequency at $\approx$7.5 ka has been recorded in sites with different coniferous vegetation composition 1000 km apart, south of the James Bay (Carcaillet et al., 2001a) and in eastern Quebec (Asnong, 2000). These observations suggest that fire is controlled more by sub-continental climate than by vegetation composition or structure at a large scale.

3.3. Biomass burning in South and Central America

The pattern of biomass burning in Central and South America varies with location. The northern tropical area between 1° S and 10° N (Fig. 5a) shows little burning activity between 11 and 9 ka, followed by five millennia of greater activity (7–3 ka). The intensity of biomass burning in the last 1000 years appears low compared with that of the middle Holocene and seems more similar to the early Holocene (Fig. 5a). The greatest concentration of fire evidences based on ^{14}C dates from soil charcoal from areas between 5° N and 5° S in the northern Amazon Basin falls between 2000 years ago and the present (Fig. 5b). Between 5 and 25° S, in the southern Amazon Basin, ^{14}C dates cover most of the Holocene, with two maxima around 7500–6000 cal. yr BP and between 2000 and 500 cal. yr BP, and a minimum between 3000 and 2000 cal. yr BP (Fig. 5c).

The pattern of biomass burning in tropical South and Central America during the Holocene appears latitudi-

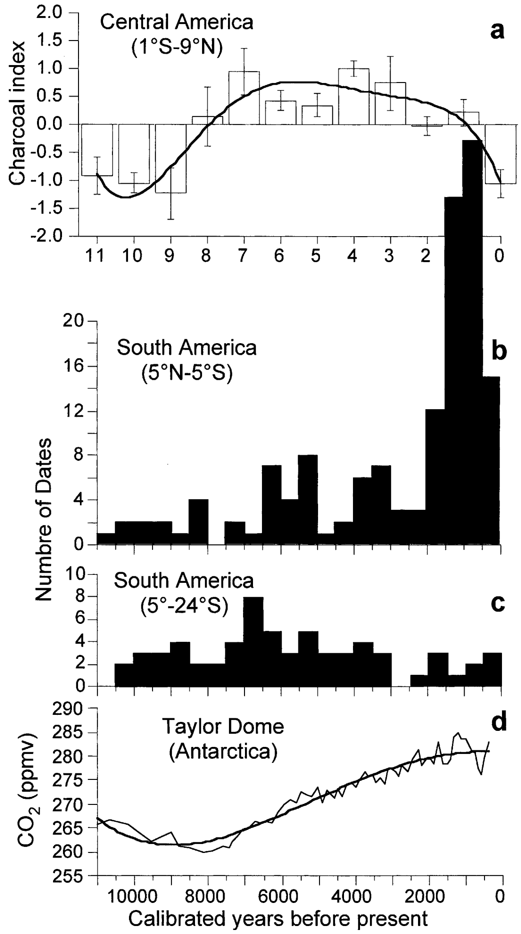

Fig. 5. Biomass burning reconstruction in Central and South America for tropical ecosystems. (a) Charcoal index deduced from stratigraphic time series in northern American tropical areas; error bars correspond to the standard error. The charcoal index is deduced from cumulative charcoal series corrected with respect to the number of sites, and averaged by millennia [method of Anderson and Smith (1997)]. (b) Distribution of ^{14}C dates from soil charcoal from northern Amazon Basin. (c) Distribution of ^{14}C dates from soil charcoal from southern Amazon Basin. (d) CO_2 concentration curve from Indermühle et al. (1999).

nally and thus may be connected with movements of the Inter-Tropical Convergence Zone (Martin et al., 1997). However, human populations probably arrived before 12 000 cal. yr BP, and certainly by the early Holocene (Roosevelt et al., 1991), and might have contributed to biomass burning. The role of climatic *versus* anthropic activities in prehistoric fire activity is debated in South America (Saldarriaga and West, 1986; Markgraf and Anderson, 1994; Hansen and Rodbell, 1995; Behling and da Costa, 2000). Whatever the cause of fire ignition, the pattern of biomass burning is poorly correlated with the curve of CO_2 from Antarctica during the late Holocene (Fig. 5). Taken together, reconstructions of tropical burning from South and Central America do not explain Holocene changes in atmospheric carbon. However, fires in areas between 5 and 25° S and 1° S and 9° N may have contributed to the early increase of the CO_2 concentration curve between 8000 and 3000 cal. yr. In Central America, Native American agricultural civilisations expanded during the middle Holocene using fire for deforestation (Horn, 1993; Goman and Byrne, 1998; League and Horn, 2000) and thus have potentially act on the carbon dynamics by biomass burning. A decrease in biomass burning in the last 500 years (Fig. 5a) is associated with the dramatic decline of native population including the Maya, Aztec, and Olmec. In the southern Amazon basin, savannah vegetation, which can be promoted by positive feedbacks to fire occurrence (Cochrane et al., 1999), expanded during middle Holocene (Desjardins et al., 1996) in association with biomass burning (Fig. 5c). A wetter climate appears to have developed during the late Holocene triggering forest expansion (Salgado-Labouriau et al., 1997; Ledru et al., 1998; de Freitas et al., 2001) and is associated with a decline in biomass burning (Fig. 5c). In the northern Amazon basin, increasing burning activity since 1500–2000 cal. yr BP (Fig. 5b) seems to be associated with human occupation or expansion (Saldarriaga and West, 1986; Behling and da Costa, 2000).

3.4. Biomass burning in southeast Asia

In Indonesia and Papua New-Guinea, stratigraphic charcoal from lakes and swamps clearly indicates two different fire regimes during the Holocene (Fig. 6). The first corresponds with the early Holocene between 10 and 6 ka when very little burning occurred, and the second occurred since 5 ka with more significant biomass burning, particularly between 4 and 3 ka.

The record of biomass burning between 8 and 3 ka in Indonesia and Papua New-Guinea parallels the global atmospheric CO_2 concentration from Antarctica (Fig. 6a versus b). The late Holocene follows CO_2 concentrations less closely, although fire activity remains much more important than during the first half of the Holocene. Biomass burning in these regions is interpreted as resulting from both agricultural expansion (Haberle, 1996, 1998) and intensification of El Niño Southern Oscillation effects by orbital forcing, such as severe drought and associated fires (Haberle and Ledru, 2001).

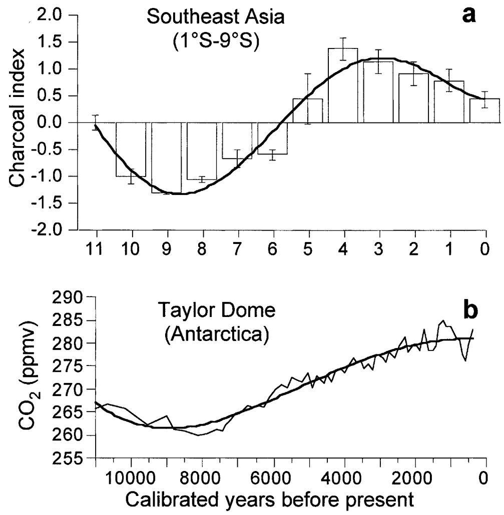

Fig. 6. Biomass burning reconstruction in southeast Asia (Indonesia and Papua New-Guinea) for tropical ecosystems. (a) Charcoal index deduced from stratigraphic time series; error bars correspond to the standard error. The Charcoal index is deduced from cumulative charcoal, series corrected in respect to the number of sites, and averaged by millennia [method of Anderson and Smith (1997)]. (b) CO_2 concentration curve from Indermühle et al. (1999).

4. Discussion

4.1. Holocene fire history and carbon release

Most sub-continental reconstructions of biomass burning indicate low but increasing activity during the first half of the Holocene (Figs. 3, 5 and 6). Only northeastern North America displays comparatively high, although decreasing levels of burning activity between 9 and 7 ka (Fig. 4). This pattern most likely resulted from effects of the Laurentide ice sheet until 8–7 ka, which created a sub-continental anti-cyclonic dominated climate that promoted fire ignition and spread. Compared with the tropics of South America and southeast Asia and Europe, where burning activity was very low during the early Holocene, the area along the Laurentide ice sheet covered by the present data-set is insignificant. Consequently, the early Holocene is globally characterised by few fires and very low carbon released to the atmosphere.

The rise of atmospheric CO_2 concentration starting around 7000 cal. yr BP recorded in Antarctic ice-core gas bubbles (Indermühle et al., 1999) matches the increasing burning activity in Europe and the tropics starting between 8000 and 7000 years ago. Fire activity in the middle Holocene arrived at intermediate levels with high fire activity in Europe, Indonesia and Papua New-Guinea, and northeastern North America (Figs. 3, 4 and 6). Data from South and Central America argue against increasing biomass burning and related carbon release in the middle Holocene. Indeed, in South America, fire reconstructions between 5° N and 5° S only indicate an increase of biomass burning (Fig. 5b; Hansen and Rodbell, 1995; Behling and da Costa, 2000), while sites in Central America and the southern Amazon Basin indicate a decrease in fire occurrence (Fig. 5a,c). Other sites record more fire during the late Holocene (League and Horn, 2000) suggesting regional variability in Central America.

The late Holocene is a period of significant burning and consequent carbon release to the atmosphere in Europe, north-eastern North America, South America between 5° N and 5° S and southeast Asia (Figs. 3, 4, 5b and 6). The increase in burning partially resulted from the increase deforestation, charcoal production, and domestic fuel, but also from changes in atmospheric circulation. Little charcoal data are available from eastern Asia or India where human populations are dense and fires have certainly been set for deforestation and domestic purposes for millennia. New Zealand's fire history is a model illustrating the role of humans in biomass burning: almost no charcoal is recorded before 2000 yr ago, while most peat and lake sediment records indicate first occurrence or rise of charcoal since about 1000 years ago associated to the Polynesian arrival (e.g. Ogden et al., 1997; Elliot et al., 1998; Horrocks et al., 2000). Nevertheless, the European burning history (Fig. 3), and the Central American experience until the diachronic collapse of different Amerindian civilisations in the last millennium (Fig. 5a) show that agricultural activity is an important factor controlling biomass burning. Extrapolation to other main prehistoric population centres, e.g. eastern Asia, India, and Africa, suggests that human-induced fires might explain the increasing atmospheric CO_2 concentrations recorded in Antarctica since 7000 years.

We therefore suggest that the carbon released by fires to the atmosphere has intensified during the late Holocene, i.e. during periods of slash-and-burn deforestation, which characterised agricultural development since about 7000 to 6000 years ago in Europe. The role of fire in the global carbon cycle is more important than previously expected, not only in terms of gaseous carbon emissions, but also in terms of carbon sequestration in the form of soil charcoal (Carcaillet and Talon, 2001). The missing source of 165 Gt of carbon calculated by Indermühle and co-workers (1999) to explain the rise of CO_2 atmospheric uptake may be located in primeval forests from middle and high latitudes. The high and middle latitudes of the northern hemisphere corresponding to Mediterranean, temperate, and boreal forests are crucial sinks for carbon, which is stored primarily in peat and humus (Van Campo et al., 1993; Zoltai and Martikainen, 1996; Gajewski et al., 2001) and thus source of carbon if biomass burning increases in frequency or severity. Fires from these latitudes seems to be a cause of global atmospheric carbon flux (Figs. 3 and 4).

The historical importance of tropical fires needs to be better documented because some areas show high burning activity while others indicate low activity (Figs. 5 and 6). For example, the tropical zones in South America have recorded an expansion of forests resulting from wetter climate during the second half of the Holocene (Desjardins et al., 1996; Salgado-Labouriau et al., 1997; Behling, 1998; Ledru et al., 1998), while Africa experienced an expansion of desert and savannah at the expense of woodlands (Barakat, 1995; Jolly et al., 1998; Maley and Brenac, 1998). Consequently, the carbon release by tropical vegetation from Africa and Arabia (Indermühle et al., 1999) might have been compensated by an uptake in South American tropical forests. Consequently, effect of biomass burning on the carbon budget could be less important in low latitudes than in middle and high latitudes.

Fires release different gaseous and particulate products. Most biomass is converted to CO_2, while on an average less than 10% (Novakov et al., 1997) becomes recalcitrant carbon (charcoal, organic carbon particles) that is temporarily withdrawn from the carbon cycle and stored in soils and sediments. Reliable quantitative estimations of this sink are hardly possible, also considering that fire itself can partly remobilize recalcitrant carbon through combustion of charcoal-rich soil horizons and (peat) sediments. Nevertheless, based on estimations for different biomes (e.g. Tinner and Hu, 1999; Carcaillet and Talon, 2001), we assume that on millennial time scales the fraction of biomass and soil organic matter converted to CO_2 during fire events play a greater role than does the pool of recalcitrant carbon. Hence, fires are strong net CO_2 suppliers and in biogeochemical cycling this effect is not compensated by the cumulative withdrawal of black and organic carbon particles.

4.2. Future issues on biomass burning and carbon dynamics

4.2.1. Fire, vegetation, and carbon release

Quantification of carbon emissions resulting from paleo-fires cannot be easily accomplished from CHAR values. High CHAR is generally attributed to high fire occurrence. This assumption is roughly acceptable for analyses of low temporal resolution. If a fire is very severe, i.e. if the biomass is deeply burned, the biomass may be converted to ash and gases rather than charred particles. Frequent severe fires would thus release large quantities of carbon gases and very relatively few charcoal particles. The rate of carbonisation (ratio of charred to uncharred wood) depends on the moisture content of fuel (Trabaud, 1976) and is associated with severity of drought. If the fuel is very dry, mineralization (burning to ash) will be intense with lots of ash and gaseous emissions, and the production of charred particles will be proportionally low. Consequently, high mineralization rates would occur when vegetation is highly flammable and combustible, and when chronic droughts are severe and last several weeks to months (Stocks and Kauffman, 1997). These conditions of complete or almost complete combustion predominate in ecosystems dominated either by needle-leaved and schlerophyllous trees and shrubs with high concentrations of volatile compounds (e.g. Mediterranean), or by herbs (e.g. savannah and steppes). In boreal forests carbon release comes primarily from the ground layer (woody debris, humus), which is more susceptible to becoming completely consumed than is the above-ground biomass (Kasischke et al., 2000). Fire severity relative to vegetation structure and composition is thus a critical parameter in estimating carbon release based on stratigraphic charcoal analysis. A present-day characterisation of charcoal and carbon gaseous emissions with respect to vegetation- and fire-type is needed to transform stratigraphic charcoal data into estimates of carbon flux to the atmosphere. The apparent relationship between mean CHAR and pollen assemblages (Clark et al., 1996a) suggests that processes of particles record are vegetation dependent, and that carbon release could be quantifiable with respect to vegetation structure and composition. Thus, the type of vegetation burned must be determined in order to estimate the amount of gaseous carbon released.

The problem is much more complex for soil charcoal. Soil charcoal abundance depends on weathering processes that might contribute to fragmentation of charcoal particles with time, and thus decrease the sampling hazard of very old fragments large enough for dating. Furthermore, the abundance of ^{14}C dates does not mean that a large amount of carbon has been released because of the taphonomy of charcoal production with respect to fire severity. One way to infer carbon release from ^{14}C-dated charcoal records involves estimation of the charcoal content of the soil (Carcaillet and Talon, 2001), and the taxonomic identification of charcoal assemblages to deduce the nature of past burned vegetation (Carcaillet and Brun, 2000). As with peat and lake-sediment charcoal records, experimental burning is needed to monitor processes of soil charcoal accumulation and dynamics with respect to fire and vegetation type (Clark et al., 1998; Ohlson and Tryterud, 2000) and the dynamics of charcoal in soil with respect to stand and local physiography of sites (Carcaillet, 2001a,b).

4.2.2. High-resolution reconstruction

High-temporal resolution of charcoal analyses allows reconstruction of changes in fire frequency through time (Clark, 1990; Bergeron et al., 1998; Long et al., 1998; Hallett and Walker, 2000; League and Horn, 2000; Millspaugh et al., 2000), which is the best proxy of carbon release by biomass burning. However, most available data are from low-resolution and discontinuous records, which do not allow accurate determination of the timing of changes in fire regime (Fig. 7). Low-resolution analysis might lead to confusion between changes in fire regime and changes in background

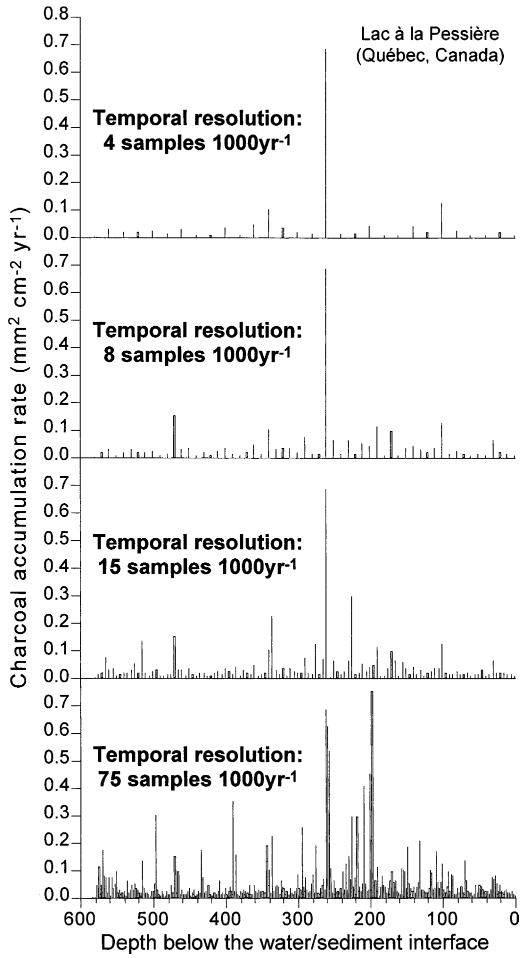

Fig. 7. Example of low and high-time resolution analysis of charcoal of the same core. The high-resolution analysis increases the possibility of detecting major periods of biomass burning and allows the assessement of the transition between periods with high precision.

charcoal. The background is the charcoal that accumulates almost continuously in the sediment and that does not necessarily represent local fire regime (Clark and Royall, 1996; Long et al., 1998). The background charcoal may be dependent on regional fire regime for small charcoal fragments, but also on lake-level fluctuation, bioturbation at the water–sediment interface, and erosion processes that create a constant disturbance of the sediment and thus a frequent release of charcoal particles to the water column. Consequently, to base the biomass burning reconstruction on low-resolution analyses carries the risk of confusing temporal changes in charcoal-collecting processes with charcoal sedimentation from fire events. Furthermore, high-temporal resolution could provide the possibility to study temporal variability in fire severity based on the magnitude of charcoal peaks (Carcaillet, unpublished).

4.2.3. Spatial variability of biomass burning

The pattern of biomass burning and carbon release to the atmosphere varies at different spatial scales. Fires and carbon release depends (i) on air mass circulation, orography (Flannigan et al., 2001), and vegetation type (French et al., 2000) on the continental scale, (ii) on climate and vegetation on a regional level (Clark et al., 1996b; Tryterud, 2000), and (iii) on topography and vegetation pattern on the landscape level, (Carcaillet and Talon, 2001). Human practices act on spatial variability by modifying the natural pattern of biomass burning by either suppressing fires as during the 20th century (e.g. Heyerdahl et al., 2001), or by increasing fire frequency and decreasing the size of fires as occurred between the 17th and the 19th centuries in Sweden (Niklasson and Granström, 2000). Furthermore, burning history of the Mediterranean and dry tropical areas is poorly documented. Now about 95% of present-day fire occurrences and burned area in Europe are concentrated in Mediterranean forests (Päätalo, 1998), and tropical fires represent 35% of global fires (Dwyer et al., 2000), suggesting that fires in Mediterranean-type and tropical ecosystems may have contributed for a long time to the global carbon dynamic by carbon release to the atmosphere. An estimate of the geographic variability of biomass burning is thus a prerequisite for a global reconstruction of carbon flux from vegetation to the atmosphere as a result of fire activity.

4.2.4. Controls of biomass burning

One major issue of studies on fire and carbon fluxes is the explanation of spatial and temporal variability of biomass burning. Although climate seems to be the best candidate to explain the variability of fire regime in northeastern North America, in other regions human activity is the best explanation for a sustained high level of biomass burning. However, if human practices have driven fire regime, climate may amplify the hazard of ignition and spread. For instance, the expansion of Mediterranean-type vegetation in the last 6000 years in the northwestern Mediterranean basin (Jalut et al., 2000) probably resulted from cultural development since the Neolithic (Chabal, 1997; Vernet, 1997; Pons and Quézel, 1998), but may also be connected to changes in climate that favoured fire; furthermore, the increasing biomass burning may be directly driven by the spread of schlerophyllous and needle-leafed species in northwestern Mediterranean basin and not by human practices since Neolithic. Human practices and climate may have influenced vegetation since 6000 years ago, thereby indirectly affecting the fire regime and associated carbon release. The problem appears simple, but its resolution is complex because each factor can be both a cause and consequence of biomass burning.

Because the current global warming and changing land-use throughout the world affects the fire regime and thus the global carbon emissions by biomass burning, it is crucial to analyse the past role of each component in determining fire regime. Efforts are required to produce high-resolution analyses of paleo-biomass burning to detail the timing of processes, and to reconstruct the fire frequency and severity. The climatic, vegetation/fuel or human causes of change in fire regime may be tested by coupling biomass burning reconstructions with out put of (1) climate models (e.g. GCM), (2) vegetation models (e.g. BIOME) and reconstruction (based on vegetation proxies), and (3) human population kinetics and distribution inferred from archeology (e.g. Bocquet-Appel and Demars, 2000).

5. Conclusion

Because charcoal is a proxy of carbon release by biomass burning, our reconstruction of charcoal accumulation in Europe, northeastern North America, Central and South America, and Oceania suggests that fire has been a process in the increase of atmospheric CO_2 since 7000 years. The biomass burning reconstructions are currently insufficient for providing quantitative estimates of carbon released by fires because the links between stratigraphic charcoal and gaseous carbon fluxes are difficult to assess. Future reconstructions of global and regional carbon release based on biomass burning require experiments to measure the relationship between CHAR, and vegetation and fire-typology. More high-resolution analyses of charcoal from small lakes and peat would improve our understanding of the temporal variability of carbon emissions and would help detail transition changes in frequency of biomass burning. More detailed spatial reconstructions of fire history are necessary to assess the spatial variability of biomass burning and carbon release associated with human activity, vegetation, and climate, the main causes of changes of fire regime.

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