Erosional history of the eastern Tibetan Plateau since 190 kyr ago: clay mineralogical and geochemical investigations from the southwestern South China Sea

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Abstract

A high-resolution study of clay mineralogy and major element geochemistry has been conducted on high sedimentation rate cores (MD01-2393 and MD97-2150) collected off the Mekong River mouth in the southwestern South China Sea in order to reconstruct the erosional and weathering history of the Mekong Basin. The chronology is based upon planktonic foraminiferal oxygen isotope records combined with carbonate stratigraphy.

Clay minerals and major element results suggest that the Mekong River is the major sedimentary source over the past 190 kyr for both cores. Illite and chlorite were derived mainly from the eastern Tibetan Plateau. Kaolinite was derived mainly from active erosion of inherited clays from reworked sediments in the middle part of the Mekong Basin. Smectites originated mainly through chemical weathering of parent aluminosilicate and ferromagnesian silicate under warm and humid conditions in the middle to lower parts of the Mekong Basin.

Smectites/(illite + chlorite) and smectites/kaolinite ratios coupled to K₂O/SiO₂ and Al₂O₃/SiO₂ ratios allow us to reconstruct a history of chemical weathering versus physical erosion. Good correlations between those ratios suggest monsoon-controlled weathering and erosion over the eastern Tibetan Plateau and the Mekong Basin. Variations with time scales of 0.3–0.6 kyr for the last 70 kyr and 1–2 kyr for 70–190 kyr ago in the chemical weathering/erosional history are strongly related to the East Asian monsoon evolution, which is itself related to the summer solar radiation calculated for the Northern Hemisphere.

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1. Introduction

The Asian monsoon system is a major component of the southern and eastern Asian climate systems. It dominates seasonal patterns of winds, precipitation,
and runoff, and determines in part the character of land vegetation over southeast (SE) Asia (Webster, 1994; Ding et al., 1998; Wang et al., 1999). Winter monsoon is characterized by the development of high pressure over northern Asia, northeast winds across the South China Sea, and increased rainfall in the Austral-Asian equatorial zone. Conversely, summer monsoon circulation is characterized by the development of low pressure over central China, inducing southerly winds across all of southern Asia and high precipitation over southern and eastern Asia. On a regional scale, changes in the strength of summer monsoon rainfall represent an important factor driving weathering and erosion of the Himalayas and the Tibetan Plateau (Bouquillon et al., 1990; Derry and France-Lanord, 1996) as well as in East Asia (Clift et al., 2002). Over glacial–interglacial time scales, a strong relationship between weathering/erosion and climate has been established by previous studies (e.g., the erosional history of the Himalayan and Burman ranges, Colin et al., 1999, 2001; the Pliocene weathering of South China, Wehausen and Brumsack, 2002). Similar monsoon-controlled weathering and erosion, therefore, would be expected in the eastern Tibetan Plateau.

In this paper, we present a high-resolution study of clay mineralogy and major element geochemistry of two International Marine Past Global Change Study (IMAGES) cores, MD01-2393 and MD97-

Fig. 1. Topography and bathymetry of SE Asia and location of the studied cores. A dashed line shows the Mekong Basin. Arrows indicate modern near-surface Asian monsoon winds dominating during summer (solid) and winter (dashed), respectively.
2150, offshore from the Mekong River (Fig. 1). These two cores were retrieved in order to collect detrital materials supplied to the South China Sea from the Mekong River, which drains the eastern Tibetan Plateau and the Indochina Peninsula. Furthermore, the South China Sea is a region located between the “Western Pacific Warm Pool” and the Asian continent, recording short- and long-term paleoceanographic responses to both summer and winter East Asian monsoons (Wang et al., 1999). Therefore, sediments recovered off the Mekong River allow us to reconstruct both erosion and weathering of the eastern Tibetan Plateau and to establish the relationship between weathering and East Asian monsoon evolution since 190 kyr ago.

2. Geological setting of the Mekong Basin

The Mekong River, which is 4880 km in length (Gupta et al., 2002), drains an area of $790 \times 10^3 \text{ km}^2$ and ranks as the 10th largest river in the world in terms of sediment discharge with an average $160 \times 10^6 \text{ t/yr}$ (Fig. 1) (Milliman and Meade, 1983; Milliman and Syvitski, 1992).

The upper reach of the Mekong River reaches an elevation of $>2500 \text{ m}$. Its basement consists mainly of Mesozoic sedimentary rocks (meta-sandstone, shale, slate, and phyllite), with minor Precambrian metamorphic rocks and Mesozoic–Cenozoic extrusive igneous rocks (Figs. 2 and 3). The corresponding soils of the plateau consist of very primitive weathering complexes, labeled lithosols, whose clay fractions therefore consist principally of fragments of the minerals contained in the parent rocks (i.e., illite and chlorite in the case of the eastern Tibetan Plateau) (Liu et al., 2003a). Downstream from Tibet, the middle reach (channel length of about 2000 km) is mountainous with steep slopes and high relief between the valley bottoms and the ridge crests. This part of the basin is underlain mainly by Paleozoic–Mesozoic sedimentary rocks and Paleozoic–Cenozoic intrusive granitic rocks (Figs. 2 and 3). A few Precambrian rocks are also exposed in the western part of the basin. Two types of soils are developed in the middle reach: bissiallitic soil associated with Paleozoic sedimentary rocks and ferrallitic soil accompanying intrusive igneous rocks (Ségalen, 1995) (Fig. 3). Mineralogical components are 2:1 layer (the assemblage of two tetrahedral sheets with one octahedral sheet) clays (e.g., smectite) for the bissiallitic soil and 1:1 layer (the assemblage of one tetrahedral sheet with one octahedral sheet) clays (e.g., kaolinite) for the ferrallitic soil, respectively. The lower reach of the Mekong River starts around Vientiane in Laos (Fig. 2). Mesozoic sedimentary rocks (mainly sandstone and mudstone) prevail in most parts of the basin basement. Downstream from Vientiane, the Mekong enters a broad alluvial plain, in which several large Neogene basalt bodies outcrop (Figs. 2 and 3). In its lowest course, the river builds the largest delta in SE Asia in terms of drainage basin size and sediment discharge (Milliman and Syvitski, 1992). The surface soils in the lower reach are mainly bissiallitic (Ségalen, 1995).

3. Materials and methods

Cores MD01-2393 (10°30.15’N, 110°03.68’E, 1230 m water depth) and MD97-2150 (10°11.76’N, 119°31.51’E, 292 m water depth) were collected close to the Mekong River mouth during IMAGES cruises VII-WEPAMA in 2001 and III-IPHIS in 1997, respectively (Fig. 1).

The lithology of Core MD01-2393 is homogeneous, dominated by olive gray foraminifer-rich or diatom-bearing nannofossil ooze with terrigenous clay (Fig. 4). Although the core is located in the southern part of a modern upwelling zone (Wang et al., 1999), microscope observation shows that biogenic silicates (e.g., diatoms and radiolarians) constitute as little as 5–10% of the volume of particles $>63 \mu\text{m}$. A layer of grey volcanic ash is observed at 1946–1950 cm depth. The site of Core MD97-2150 was only 50 km away from the paleo-Mekong Estuary during the last glacial interval, when the sea level dropped 116 m (Hanebuth et al., 2000). The lithology consists of olive gray terrigenous clay and silt with foraminifer- or nannofossil-bearing ooze (Fig. 5). A level rich in quartz and feldspar sands is observed between 200 and 350 cm depth below seafloor.

Clay minerals were identified by X-ray diffraction (XRD) at the Laboratoire de Sédimentologie et Géodynamique, Université de Lille I on oriented mounts of noncalcareous clay-sized particles (Holtzapfel, 1985). Deflocculation was accomplished by succes-
Sieve washing with distilled water after decarbonation with 0.2 N HCl. Particles smaller than 2 μm were separated by sedimentation and centrifugation. XRD spectra were obtained using a Philips PW 1710 diffractometer with CuKα radiation and Ni filter, under a voltage of 40 kV and an intensity of 25 mA. Three XRD runs were performed following air drying, ethylene–glycol solvation for 12 h, and heating at 490 °C for 2 h. Identification of clay minerals was made mainly according to the position of the (001) series of basal reflections on the three XRD diagrams. Semi-quantitative estimates of peak areas of the basal reflections for the main clay mineral groups of smectites (smectite + mixed layers) (15–17 Å), mixed

Fig. 2. Schematic geological map of SE Asia (modified after the Commission for the Geological Map of the World, 1975).
layers (15 Å), illite (10 Å), and kaolinite/chlorite (7 Å) were carried out on the glycolated curve (Holtzapffel, 1985) using the MacDiff software (Petschick, 2000). Relative proportions of kaolinite and chlorite were determined based on the ratios of the 3.57/3.54 Å peak areas. Illite crystallinity was obtained from a half height width of the 10 Å peak. Lower values represent higher crystallinity, characteristic of weak hydrolysis in continental sources and dry and cold climate conditions (Chamley, 1989).

Major element contents were analyzed from samples taken from Core MD01-2393, using an electron microprobe Philips XL-30 at the Orsayterre Laboratory, University of Paris XI after fusion of the carbonate-free fraction following the procedure described by Colin et al. (1998). The carbonate fraction was removed by leaching 50 mg of sediment with 20% acetic acid, followed by rinsing several times and centrifuging to remove traces of carbonate solution. As the South China Sea sediments are characterized by a low and/or negligible weight percentage of biogenic silicates, opal fraction was not leached by conventional methods (sodium hydroxide or sodium bicarbonate) in order to preserve terrigenous material from dissolution of labile elements. The carbonate-free fraction was then carefully mixed by hand with 20% Li$_2$CO$_3$ (Merck) in an agate mortar. The mixture was fused in air on a platinum cell by radiofrequency induction heating. The cell was heated to 90 °C for 15 s to drive off H$_2$O and CO$_2$ from the sample, and the temperature was then increased sufficiently above the mixture liquidus (about 1350 °C) to ensure complete melting. After quenching at room temperature, glass samples were included in epoxy resin and polished for analysis.

For both cores, oxygen isotope compositions (δ$_{18}^O$) were measured on planktonic foraminifera Globigerinoides ruber (white) (250–315 μm) using an Optima VG mass spectrometer at the Laboratoire des Sciences du Climat et de l’Environnement, Gif-sur-Yvette. The measurements are reported versus PDB after calibration with NSB19. The mean external reproducibility is ± 0.07‰ for δ$_{18}^O$. Finally, carbonate contents were determined using the gasometric techniques of Jones and Kaiteris (1983) with a precision better than ± 2%.

4. Chronological framework

The chronology of Core MD01-2393 has been established using the high-resolution oxygen isotope record combined with carbonate stratigraphy and tephrrostratigraphy (Fig. 4). For the last 18 kyr, a graphic correlation was performed using the Analyses software (Paillard et al., 1996) between the oxygen isotopic record of planktonic foraminifera G. ruber (white) and those in nearby Core MD97-2151 (8°43.73’N, 109°52.17’E, 1550 m water depth) (Fig. 1). This method produces a detailed chronology con-
Fig. 4. Age model of Core MD01-2393, indicating lithological column and description, planktonic foraminifera *G. ruber* (white) $\delta^{18}O$ record, and carbonate contents versus depth. The age-based (kyr) *Globigerinoides ruber* (white) $\delta^{18}O$ record and linear sedimentation rate are also displayed. The YTT and the last occurrence (LO) of *G. ruber* (pink) are labeled. Marine isotope event (MIE) and MIS after Martinson et al. (1987). Planktonic foraminifera *Globigerinoides sacculifer* $\delta^{18}O$ record from Core MD97-2151 (Lee et al., 1999) and stacked SPECMAP $\delta^{18}O$ (Martinson et al., 1987) are included for their correlation. The data (fine) were smoothed by a five-point moving average (coarse) for the $\delta^{18}O$ curve of Core MD97-2151.
Fig. 5. Age model of Core MD97-2150, indicating lithological column and description, planktonic foraminifera *G. ruber* (white) δ¹⁸O record, magnetic susceptibility (sensor units) (Michel et al., 1997), and carbonate contents versus depth. The age-based (kyr) *G. ruber* (white) δ¹⁸O record and linear sedimentation rate are also displayed. MIE and MIS after Martinson et al. (1987). Planktonic foraminifera *G. sacculifer* δ¹⁸O record from Core MD97-2151 (Lee et al., 1999) is included for their correlation. The data (fine) were smoothed by a five-point moving average (coarse).
strained by 12 AMS $^{14}$C datings (Lee et al., 1999). Before 18 kyr, chronology was established by using the SPECMAP oxygen stratigraphy proposed by Martinson et al. (1987). According to our oxygen isotope stratigraphy, the volcanic ash layer observed at 1946–1950 cm interval depth corresponds to the Youngest Toba Tuff (YTT) (Fig. 4). This volcanic eruption has been already observed in the southern South China Sea (Bühring et al., 2000; Song et al., 2000) and was dated approximately at $74 \pm 2$ kyr (K/Ar), providing an additional time marker for our chronological framework (Ninkovich et al., 1978).

Carbonate contents vary between 5% and 22%, with higher values during interglacials (Fig. 4), in agreement with previous studies in the South China Sea (Wang et al., 1995; Huang et al., 1999), supporting our age model. This core provides a continuous sedimentary record extending down to marine isotope stage (MIS) 6 (about 190 kyr) (Fig. 4). The linear sedimentation rate varies over the range 10–41 cm/kyr, with higher values generally appearing during glacial stages 6, 4, 3, and 2, and the Holocene (Fig. 4).

The age model of Core MD97-2150 was established using the oxygen isotope stratigraphy in combination with variations in magnetic susceptibility and carbonate contents (Fig. 5). Several oxygen isotopic events were determined by correlating the planktonic foraminiferal G. ruber (white) $\delta^{18}$O with those of Core MD97-2151 (Fig. 5). MIS 2/1 and 3/2 transitions are determined using both planktonic foraminiferal $\delta^{18}$O records and magnetic susceptibility (Fig. 5). The linear sedimentation rate is much lower compared to the deep water Core MD01-2393 and shows less variation with an average of 17.9 cm/kyr (Fig. 5).

Both cores were sampled at 5–20 cm intervals in order to analyze clay mineralogy and major element content with average temporal resolutions of 0.3–0.6 kyr for the last 70 kyr and a lesser resolution of 1–2 kyr for the older part.

5. Results

5.1. Clay mineralogy

In Core MD01-2393, illite (21–40%) and smectites (22–58%) are the dominant clay minerals (Fig. 6). Clay minerals present in lesser abundance include chlorite (10–25%) and kaolinite (11–25%), accompanied by minor amounts of quartz and feldspar in the clay size fractions. The clay mineral distributions indicate strong glacial–interglacial cyclicity. In general, illite, chlorite, and kaolinite concentrations are similar and inversely correlated to the smectite concentration (Fig. 6). MIS 2, 3, 4, and 6 are characterized by higher contents of illite (35–40%) than interglacial MIS 1 and 5 (20–35%), whereas smectites show higher values during interglacials (30–40%) than during glacials (20–35%). On a shorter time scale, chlorite shows an opposite pattern compared to illite during the Holocene (last 10 kyr), and is characterized by an increasing chlorite content during the Holocene. Glacial periods are also characterized by a slight increase of the illite crystallinity (Fig. 6). Moreover, long-term changes in the quartz/feldspar ratio of clay fractions present a similar glacial–interglacial variation with relatively higher ratios during glacial periods (Fig. 6).

In Core MD97-2150, illite (23–45%) and kaolinite (12–29%) proportions are higher than those in Core MD01-2393 (Figs. 6 and 7). In contrast, smectites (11–40%) are less abundant than in Core MD01-2393, whereas chlorite presents a similar range between 11% and 25%. Illite crystallinity displays the same average value of about 0.23 $\Delta$2θ as seen in Core MD01-2393, but without any significant glacial–interglacial variations. During MIS 3 and 4, clay mineral distribution, illite crystallinity, as well as quartz/feldspar ratios do not exhibit any significant changes. Since the beginning of MIS 2, illite and chlorite concentrations have decreased, whereas kaolinite content has increased by about 10%. The quartz/feldspar ratio increased abruptly from an average of 0.3 to about 1.0 in the end of MIS 2 (Fig. 7). During MIS 1, the chlorite content increased after about 10 kyr ago, while illite decreased. A maximum in smectite content is observed at about 6 kyr.

5.2. Major elements

In Core MD01-2393, carbonate-free sediments consist mainly of SiO$_2$, Al$_2$O$_3$, Fe$_2$O$_3$, K$_2$O, and MgO, with low concentrations of Na$_2$O, CaO, TiO$_2$, and MnO. In Fig. 8, we have selected six important elements that can be subdivided into three groups.
Fig. 6. Clay mineral proportions (%) on the <2-μm size fraction versus age (kyr) for Core MD01-2393. Illite crystallinity, quartz/feldspar (in clay fraction) ratio, and isotopic stages are also displayed. The illite crystallinity data (fine) were smoothed by a three-point moving average (coarse).
Fig. 7. Clay mineral proportions (%) on the <2-μm size fraction versus age (kyr) for Core MD97-2150. Illite crystallinity, quartz/feldspar (in clay fraction) ratio, and isotopic stages are also displayed. The illite crystallinity data (fine) were smoothed by a three-point moving average (coarse).
Fig. 8. Major element contents of Core MD01-2393.
Long-term changes of K$_2$O, CaO, and, to a lesser extent, Na$_2$O are observed and exhibit similar variations. MIS 6 and 1, as well as the interval 25–45 kyr, are characterized by a significant decrease in concentrations of these elements, particularly noticeable at 45 kyr in K$_2$O and CaO variations.

In general, long-term trends in SiO$_2$ concentrations are inversely correlated to those of K$_2$O, CaO, and, to a lesser extent, Al$_2$O$_3$ variations. Long-term variations in Al$_2$O$_3$ concentrations are also usually inversely correlated to those of K$_2$O and CaO fluctuations. For instance, at around 45 kyr, Al$_2$O$_3$ increased sharply, whereas K$_2$O and CaO concentrations decreased significantly. In addition, TiO$_2$ concentration generally displays an inverse correlation to Al$_2$O$_3$, but does not follow the climatic cycle pattern throughout the investigated interval (Fig. 8), in contrast to other datasets in the South China Sea for older time periods (Wehausen et al., 2003).

6. Discussion

The paleoclimatic interpretation of clay mineral and detrital geochemical records requires knowledge of the potential source areas as well as the mode and strength of the transport processes involved (Gingele et al., 1998; Wehausen and Brumsack, 2002). Glacial stages are characterized by a significant increase of illite, chlorite, and kaolinite proportions, which are relatively stable in the shallow Core MD97-2150, but exhibit large variations in the deep water Core MD01-2393 (Figs. 6 and 7). In contrast, interglacial periods are characterized by an increase of smectite contents in both cores.

Variations in clay mineral composition of the surface and Quaternary South China Sea sediments do not match variations in the lithological or diagenetic states. The primary control on such variations is provenance (Chen, 1978; Boulay et al., 2003; Tamburini et al., 2003; Liu et al., 2003b). Therefore, variations of both cores may result from: (i) variations in source materials induced by fluvial and/or Eolian contributions; (ii) change in the balance between chemical and physical erosion on land; and (iii) global sea level changes. We now examine each of these potential causes for variations in clay at both cores.

6.1. Sediment sources

Illite and chlorite are considered primary minerals, which reflect the decrease of hydrolytic processes in continental weathering and an increase of direct rock erosion under cold and arid climatic conditions. In the case of the Mekong Basin, illite and chlorite could be derived from physical erosion of metamorphic and granitic parent rocks. In addition, illite could also be formed by the weathering of nonlayer silicate, such as feldspar from granites under moderate hydrolysis conditions, and by the degradation of micas. Such formations are mainly located in the highland part of the Mekong Basin, the eastern Tibetan Plateau, where lithosols are dominant (Liu et al., 2003a; Figs. 2 and 3). Variations in illite and chlorite contents as well as in illite crystallinity in both cores reflect increasing physical erosion during glacial stages 6, 4, and 2, and decreasing erosion during interglacial stages 5 and 1 in the eastern Tibetan Plateau.

Kaolinite is readily found in soils of intertropical land masses characterized by a warm, humid climate, and therefore displays a strong climatic dependence controlled by the intensity of continental hydrolysis (Chamley, 1989). Kaolinite is common on steep slopes within the drainage basin where there are good drainage conditions. In the Mekong Basin, ferrallitic soils located in a lower part of the middle reach are major potential sources of kaolinite to our two cores (Figs. 2 and 3). However, variations in kaolinite contents in both cores indicate a pattern similar to those of illite and chlorite during periods of more glacial conditions (Figs. 6 and 7), suggesting that most of the kaolinite could also be derived from active erosion of inherited clays. In addition, the exposed continental shelf of the southwest South China Sea during glacial sea level low-stand could also provide kaolinite to the deep sea by reworking. Therefore, the kaolinite contents of both cores could reflect the physical erosion in the Mekong Basin, instead of reflecting contemporary climates (Thiry, 2000). The general consistency between higher kaolinite contents and higher linear sedimentation rates during glacial periods and the Holocene in Core MD01-2393 supports this hypothesis (Figs. 4e and 6). Kaolinite could also be derived from erosion of soils in the Indonesian islands, where ferrallitic soils are common (Ségalen, 1995) and where kaolinite...
contents on surrounding sea surface sediments are high (Chen, 1978; Gingele et al., 2001). Enhanced summer monsoon (southwesterly) currents should transport both more smectite and kaolinite from the Indonesian islands to our investigated area. But at both cores, kaolinite contents are lower in sediments deposited during interglacials. This inconsistency suggests that the kaolinite contribution from the southern volcanic islands could be neglected.

Smectites are secondary minerals, which are derived from chemical weathering of parent aluminosilicate and ferromagnesian silicate under warm and humid conditions. Smectite is formed in confined environments, by recombination of released cations. In practice, smectite is not formed in the same part of the river basin as kaolinite. In the Mekong Basin, widely distributed bissiallitic soils in the upper part of middle reach and the lower reach are potential sources of smectites (Fig. 3). Annual rainfall in those parts of the Mekong Basin is heavy (2000–4000 mm) and nearly 85–90% of the annual rainfall arrives in the summer monsoon between May and October (Gupta et al., 2002). Such climatic conditions allow significant chemical weathering. The primary phyllosilicates (micas and chlorites) are first exfoliated by hydrolysis, giving way to “open” illite and chlorite. Successive steps develop if hydrolysis goes on or becomes more active. A typical evolution consists of successive formation of irregular vermiculitic mixed layers, clayey vermiculite, smectite mixed layers, and degraded smectite. Increased smectite concentrations during interglacials in both cores (Figs. 6 and 7) are consistent with enhanced summer monsoon rainfall, which is predominant in interglacials as observed at various sites in the South China Sea (Wang et al., 1999; Jian et al., 2001).

Although Thiry (2000) argued that sequential changes in sedimentary clay mineral assemblages with periods of less than 1 million years cannot be caused by climatic changes acting on soil mineralogy, recent investigations on chemical weathering of South Indian watersheds that contain similar precipitation of ~ 3000 mm and temperature of ~ 27 °C within the lower part of the Mekong Basin reveal that weathering rates can reach as high as ~ 150 kg/ha/year (Oliva et al., 2003). Given an average soil density of <2.78 g/cm³, the weathering rate is >5.4 cm/year. This value allows recordings of short-term changes in the paleoclimate of glacial–interglacial stages. Similar studies of the clay mineralogical implication for the last climatic cycle were successfully conducted both in the northern and southern South China Sea (Tamburini et al., 2003) and in the Bay of Bengal and the Andaman Sea (Colin et al., 1999).

Smectite could also derive from the chemical weathering of basaltic rocks in the Mekong Basin plain and the volcanic activity in the Indonesian islands. However, the small volume of basalt in the plain of the Mekong Basin (Fig. 2) could produce by weathering only a negligible amount of smectite. The region of the Indonesian islands provides 30–40% smectites to the Java Sea (Gingele et al., 2001), in turn feeding the South China Sea through the Borneo Strait. Northeasterly sea surface currents, which are driven by the East Asian summer monsoon, could transport smectites from the south, particularly during interglacials (Liu et al., 2003b). However, the location of the two cores close to the Mekong delta excludes significant contribution of smectites from the Indonesian islands and from the northern Sunda Shelf for the following four reasons:

(i) The most important sediment source for the South China Sea is the Mekong River, which has an average suspended sediment discharge of $160 \times 10^6$ t/year (Milliman and Meade, 1983). Rivers in north and west Borneo mainly include the Baram and the Rajang in Malaysia and the Kapuas in Indonesia, with an estimated combined annual sediment discharge of $36 \times 10^6$ t/year on basis of the value of the Baram ($12 \times 10^6$ t/year; Hiscott, 2001). Moreover, clay compositions in rivers from Borneo are characteristic of high illite and low smectite values (Eisma et al., 1989). We suggest that the smectite contribution from Borneo is considered of minor importance given the small annual discharge and the clay compositions. Both our cores are close to the Mekong River mouth on the continental slope, ensuring that the Mekong sediment source is predominant.

(ii) Effect from global sea level changes. Clay mineral (smectites, illite, chlorite, and kaolinite) contents of both cores are quite similar, implying similar sediment sources. Core MD01-2393 is characterized by a slightly higher content of smectites. This could be due to an effect of differential settling of
clay minerals during marine sediment transportation. Pedogenetic smectite crystals are very small and can be transported more easily over a greater distance from the Mekong River mouth than the other clay species (Gibbs, 1977; Gingele et al., 1998). This could also explain the different patterns of smectites between the two cores during the Holocene (Figs. 6 and 7). In the interval ~ 7 to ~ 4 kyr, relatively higher smectite contents are encountered in the shallow water core, with lower contents in the deep water core, corresponding to the local highest sea level stand during the Holocene (Hanebuth et al., 2000; Ta et al., 2002), when the shallow core location was far from the Mekong River mouth.

(iii) The Indonesian islands produce high contents of smectites, but also high kaolinite contents (Gingele et al., 2001). In both cores, the increase of smectites is not associated with the increase of kaolinite, which means that smectite sources in the Indonesian islands can be excluded for both cores.

(iv) The $^{87}$Sr/$^{86}$Sr ratio of detrital fractions from Core MD01-2393 ranges between 0.7205 and 0.7235 (C. Colin, unpublished data). Such values are similar to those of sediments derived from the Mekong River (0.7230) and are much higher than volcanic rocks from Indonesia (0.7045–0.7095; Vroon et al., 1993), suggesting the Mekong River as the main sedimentary source over the past 190 kyr for the deep core.

6.2. Chemical weathering and East Asian monsoon evolution

Given that the Mekong River Basin represents the main source of siliclastic detritus for both cores, the clay minerals and major element compositions can be used to reconstruct the weathering/erosion history of the eastern Tibetan Plateau and the lowland of the Mekong Basin over the past 190 kyr. In general, multiple sources and transport processes, as well as the dilution of individual clay minerals and elements, make it difficult to assign changes in the downcore record of a single component to a change in a paleoclimatic factor (Gingele et al., 1998; Wehausen and Brumsack, 2002). However, the comparison of two components on the basis of their ratio offers the advantage of eliminating dilution effects by other components.

We adopt ratios of smectites/(illite + chlorite) and smectites/kaolinite as mineralogical indicators to reconstruct the history of chemical weathering versus the physical erosion in the eastern Tibetan Plateau. In both cores, variations in smectites/(illite + chlorite) and smectites/kaolinite ratio show similar changes during the last 70 kyr (Fig. 9). The relatively higher ratios developed during interglacials throughout the past 190 kyr suggest strengthened chemical weathering and weak physical erosion; by contrast, the lower ratios during glacial periods indicate strongly intensified physical erosion and weakened chemical weathering (Fig. 9). These results are also supported by higher glacial quartz/feldspar ratio in clay fractions (Figs. 6 and 7), indicating enhanced physical erosion during glacial periods.

As the contribution of biogenic materials is negligible in Core MD01-2393, we suggest that minerals...
Fig. 9. Comparison of smectites/(illite + chlorite) and smectites/kaolinite ratios of cores MD01-2393 and MD97-2150 and K$_2$O/SiO$_2$ and Al$_2$O$_3$ ratios of Core MD01-2393 with the summer insolation of the Northern Hemisphere and Planktonic foraminifera G. ruber (white) $\delta^{18}$O record. The summer insolation refers to an average insolation during June and July at 65°N following the Laskar (1990) solution using Analyserie software (Paillard et al., 1996).
rich in SiO$_2$ could correspond mainly to those present in the silt fraction (e.g., quartz) (Boulay et al., 2003). In northern South China Sea sediments, SiO$_2$ is mainly related to Eolian dust and the variability of SiO$_2$ contents is mainly used to reconstruct the intensity of the Eolian contribution to marine sediments (Wehausen and Brumsack, 2002). Nevertheless, in the southern South China Sea, carbonate-free TiO$_2$/Al$_2$O$_3$ contents suggest that the Mekong River delivered most of the sediments by enhanced summer monsoon rainfall (Wehausen et al., 2003), attesting to a negligible Eolian contribution into the southern South China Sea. Hence, an increase of silt fraction (e.g., quartz) would characterize an increase in the physical erosion of granitic and metamorphic formations within the Mekong Basin highland, as also indicated by the quartz/feldspar ratio of clay fractions. Conversely, K$_2$O and Al$_2$O$_3$ are mainly present in clay minerals. While Al$_2$O$_3$ is the major component of all clay minerals, significant amounts of K$_2$O are preferably found in the clay mineral illite. Thus, K$_2$O/SiO$_2$ and Al$_2$O$_3$/SiO$_2$ ratios could be employed as a proxy to assess the relative proportion of clay and silt fraction in the sediment (Fig. 9), which is dependent on the strength of physical erosion and the effect of transport of detrital material in the Mekong Basin. In Core MD01-2393, K$_2$O/SiO$_2$ and Al$_2$O$_3$/SiO$_2$ ratio variations show a similar pattern and increase significantly during interglacial stage 5. The relatively high ratios during interglacial stage 5 suggest an increase of summer monsoon rainfall producing clays minerals in Mekong plain soils.

7. Conclusions

High-resolution clay mineralogy, major element geochemistry, and oxygen isotopic records combined with carbonate stratigraphy over the past 190 kyr from two piston cores MD01-2393 and MD97-2150 were used to reconstruct the erosional history of the eastern Tibetan Plateau. We conclude that:

1. The Mekong River is the major sedimentary source to the study area over the past 190 kyr for cores MD01-2393 and MD97-2150. Illite and chlorite were derived mainly from the eastern Tibetan Plateau, where physical erosion of metamorphic and granitic rocks is dominant. Kaolinite was derived mainly from active erosion of clays in reworked sediments in the middle part of the Mekong Basin. Smectites originated mainly through chemical weathering of parent aluminosilicates and ferromagnesian silicates under warm and humid conditions in the middle to lower reaches of the Mekong River.

2. Smectites/(illite + chlorite) and smectites/kaolinite ratios coupled to K$_2$O/SiO$_2$ and Al$_2$O$_3$/SiO$_2$ ratios permit the history of chemical weathering versus the physical erosion, which occurred in the eastern Tibetan Plateau and Mekong Basin to be reconstructed. An increase of smectites/(illite + chlorite) and smectites/kaolinite ratios suggests an intensification of the chemical weathering in the Mekong plain soils.

3. Variations in the chemical weathering/erosion history are mainly controlled by changes in the summer monsoon rainfall. Wet periods of summer monsoon reinforcement are characterized by an
increase of the chemical weathering of the Mekong plain soils relative to physical erosion of the highland Mekong Basin. The production of pedogenic clays in the plain (smectite) increases relative to detrital minerals (chlorite, illite, inherited kaolinite, and quartz).

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