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1 **Controlling weathering and erosion intensity on the southern slope of the**
2 **Central Himalaya by the Indian summer monsoon during the last glacial**

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Abstract

This paper reports the results of clay mineral analysis (the amount of clay fraction, clay mineral assemblages, illite crystallinity) of samples collected from a drilled core (Rabibhawan (RB) core) located in the west-central part of the Kathmandu Basin on the southern slope of the Central Himalaya. The amount of clay fraction in the core sediments between 12 m and 45 m depth (corresponding to ca. 17 ~ 76 ka), which belong to the Kalimati Formation, is variable and shows three clay-poor zones (19 ~ 31 ka, 44 ~ 51 ka, and 66 ~ 75 ka). The variations correspond with those of illite crystallinity index (Lanson index (LI) and modified Lanson index (MLI)) and kaolinite/illite ratio as well as the fossil pollen and diatom records reported by previous workers. These data reveal the following transformations occurring during the weathering process in this area:

micas (mainly muscovite) → illite

(→ illite-smectite mixed layer mineral (R=1)) → kaolinite.

The sedimentation rate (~ 50 cm/kyr) of clay-poor zones that correspond to dry climate intervals is only half that of clay-rich zones (~ 120 cm/kyr) that correspond to wet climate intervals, indicating weakened chemical weathering and erosion and low suspended discharge during dry climate intervals. The clay-poor zones commonly show unique laminite beds with very fine, authigenic calcite, which was probably precipitated under calm and high calcite concentration conditions caused by low precipitation and run-off.

The variations between dry and wet conditions in this area as deduced from clay minerals appear to follow the Indian Summer Monsoon Index (ISMI) (30°N – 30°S, 1 July) and northern hemisphere summer insolation (NHSI) signals (30°N) at 1 July, especially during the dry climate zones, whereas the wet maxima of the wet climate zones somewhat deviate from the strongest NHSI. On the other hand, the dry-wet

60 records lead markedly the SPECMAP stack (by about 5,000 years). These results
61 suggest that the Indian summer monsoon precipitation was strongly controlled by
62 the NHSI or summer insolation difference between the Himalayan-Tibetan Plateau
63 and the subtropical Indian Ocean, showing a major fluctuation on the 23,000 years
64 precessional cycle, and that it was not driven by changes in high-latitude ice
65 volume, although the records of clay mineral indices during the wet intervals leave
66 a question that other factors, in addition to insolation forcing, may play important
67 roles in weathering, erosion, and sedimentation processes.

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69 Keywords: Indian monsoon, last glacial, paleoclimate, weathering, clay minerals,
70 Nepal Himalaya.

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1. Introduction

The Indian monsoon system is one of the major weather systems on the Earth and affects most densely populated regions. Differential heating during summer results in a seasonal low pressure cell over the Indian continental landmass and a high pressure cell over the cooler Indian Ocean. As a consequence, warm humid southwest summer winds from the Indian Ocean flow onshore and contribute most to the rainfall (Colin et al., 1998; Kudrass et al., 2001; Rashid et al., 2007). Most of the monsoonal precipitation falls on the catchments of the Ganges-Brahmaputra-Meghna (GBM) river system, whose rivers drain most of the Himalayas and the northern Indian subcontinent (Kudrass et al., 2001; France-Lanord et al., 2003; Rashid et al., 2007). Water and suspended discharge of the river system, therefore, get concentrated during only five months (June to October) of the summer monsoon (Islam et al., 2002; Goodbred, 2003). Natural calamities, such as flooding or landslide, are also more frequent during this season (Rashid et al., 2007).

It is naturally expected that past modifications of the intensity of chemical and physical weathering and erosion of the Himalayan and Burman ranges and the GBM catchments are strongly related to past variations in the strength of the Indian summer monsoon (Colin et al. 1998). It is known that numerous paleoclimatic studies, based on several proxies such as % *Globigerina bulloides*, organic carbon content, lithogenic grain size, and pollen content, have permitted reconstruction of changes in the paleo-monsoon intensity (e.g., Anderson and Prell, 1993; Sirocko et al., 1993; Overpeck et al., 1996; Schulz et al., 1998; Ivanochko et al., 2005). These proxies, however, are generally of monsoon wind strength and monsoon wind-induced upwelling, rather than precipitation (Tiwari et al., 2006; Rashid et al., 2007; Shakun et al., 2007). Rashid et al. (2007) state that summer monsoonal precipitation on the Indian subcontinent is not linearly correlated to wind strength, because it depends on the moisture content of the incoming monsoon winds, which

101 is determined by sea surface temperature (SST) in the southern hemisphere and by
102 the convergence and rate of ascent of the air parcels after they cross the Indian coast.
103 On changes in the strength of the Indian summer monsoon precipitation or in the
104 intensity of weathering and erosion induced by precipitation, continental records
105 from the Himalayas and the north Indian subcontinent, where Indian summer
106 monsoon winds blow directly, are extremely rare (Sinha et al., 2005), while the
107 records from marine sediments are many (e.g., Bay of Bengal and Andaman Sea:
108 Colin et al., 1998, 1999; Rashid et al., 2007, Arabian Sea: Sirocko et al., 1991, 1993,
109 2000; Tiwari et al., 2006).

110 Precipitation plays a key role in the formation, weathering, erosion and transport
111 of clay minerals to the depositional basins (Singer, 1984; Chamley, 1989; Robert,
112 2004). Therefore, clay minerals can be useful indicators of paleoclimatic
113 conditions and have been used to estimate the intensity of precipitation or
114 continental wetness (Robert and Kennett, 1994; Diester-Haass et al., 1998; Robert,
115 2004). However, paleoclimatic interpretation of clay minerals or other
116 mineralogical indices such as grain size in sediments, especially of those
117 transported from large source areas such as the Himalayas and the northern Indian
118 subcontinent to the Indian Ocean, is anything but straightforward. This is because
119 weathering, erosion, transport and sedimentation processes are controlled by many
120 factors (e.g., mixing of exotic minerals, selective erosion and transport, mixing of
121 detrital and authigenic clay minerals, asynchronous weathering and
122 transport/deposition) (Singer, 1984; Chamley, 1989).

123 Moreover, Kübler Index (KI) (Kübler, 1964), which is conventional illite
124 crystallinity index defined by the full width at half maximum intensity (FWHM) of
125 10 Å illite X-ray diffraction (XRD) peak and has extensively been used to
126 reconstruct the paleoclimate (e.g., Chamley 1989; Fukuzawa et al. 1997; Lamy et al.
127 2000), is not always available for estimation of illite crystallinity (Srodon, 1979,
128 Srodon and Eberl, 1984; Lanson, 1997; Kuwahara et al., 2001). According to

129 Srodon (1979), the KI is significantly larger for finer fractions because the FWHM
130 of the illite 001 peak is mostly a function of the amount and composition of the
131 illite-smectite mixed layer (I-S) component of the sample and the I-S component
132 has a finer particle size than illite. To overcome this problem, Lanson (1997)
133 proposed a new illite crystallinity index (Lanson index (LI)), which accounts for the
134 relative proportion of illite crystallites with low coherent scattering domain size
135 (CSDS), by using decomposition procedure of X-ray diffraction (XRD) patterns.
136 The asymmetry of the complex 001 XRD peaks of illitic minerals near 10 Å is in
137 fact due to the presence of different mineral phases with different illite content and
138 different CSDS thickness (Lanson, 1997). Therefore, he decomposed the complex
139 peaks using three elementary peaks corresponding to three different phases with
140 different illite content and different CSDS thickness, that is, I-S, poorly crystallized
141 illite (PCI), and well-crystallized illite (WCI). The LI can be determined by the
142 characteristics of the three elementary peaks. The modified Lanson index (MLI),
143 which estimates illite crystallinity only from the difference between PCI and WCI,
144 is available for the estimation of variations in weathering and hydrolysis conditions
145 (Kuwahara et al., 2001).

146 The Kathmandu Basin is one of the ideal targets for studying the variations in
147 the Indian monsoon climate and their bearing on the uplifting of the
148 Himalayan-Tibetan orogen, because the basin is located on the southern slope of the
149 central Himalaya and filled with a thick pile of Late Pliocene to Quaternary
150 sediments (Sakai et al., 2001a, b; Fujii and Sakai, 2001). The Kathmandu Basin is
151 also ideal for interpretation of paleoclimates from clay minerals in sediments,
152 because the basin has a diameter of only about 30 km and the river's catchment area
153 is confined to the inside slope of the basin, implying that the basin-fill sediments are
154 supplied only from the mountains surrounding the basin (Sakai, 2001; Kuwahara et
155 al., 2001). Yet, previous studies could not completely decipher the paleoclimatic
156 changes in the Kathmandu Basin, because of discontinuities in the surface

157 exposures sampled (Yoshida and Igarashi, 1984; Igarashi et al., 1988; Nakagawa et
158 al., 1996; Goddu et al., 2007). The scientific group of this study conceived the
159 “Paleo-Kathmandu Lake (PKL) project”, under which they carried out academic
160 drilling in the Kathmandu Basin, Nepal Himalaya, and investigated the cores and
161 surface exposures from various viewpoints and by different methods (Sakai, 2001).
162 Several earlier workers reported on the results of fossil pollen, fossil diatom and
163 organic geochemical analyses and sediment characteristics from surface geological
164 surveys of the Kathmandu basin and studies of the drill cores obtained from the
165 basin (Sakai et al., 2001a, b, Fujii and Sakai, 2001, 2002, Maki et al., 2002, Fujii et
166 al., 2004, Hayashi et al., 2007a, b, Mampuku et al., 2008 ; Hayashi et al., 2009). In
167 this paper, it is attempted to reconstruct the variations in the intensity of weathering
168 and erosion conditions, as recorded in the clay minerals of the sediments from the
169 Kathmandu Basin. The variations were probably controlled by Indian summer
170 monsoon precipitation during the past 76,000 years.

171

172

173 **2. Materials and methods**

174 **2.1. Sample preparation and XRD measurements**

175 The materials used were a 218 m long core (RB core), which was obtained from
176 drilling at Rabibhawan in the west-central part of the Kathmandu Basin under the
177 PKL Project in 2000 (Sakai et al., 2001b) (Fig. 1). For clay mineral analysis, core
178 sediment samples, collected at 10 cm interval between 7 m and 45 m depth, were
179 used. The topmost part of the sampled core, from 7 m to 11 m depth, is composed
180 of medium-to very coarse-grained micaceous granitic sand beds of the Patan
181 Formation, which corresponds to the sediments of the Bagmati river (Sakai et al.,
182 2001b) (Fig. 2). The sediments immediately below this zone belong to the Kalimati
183 Formation, and those between 12 m and 45 m depth are of organic black or dark
184 gray mud, known as “Kalimati Clay”. The top 1 m part of the Kalimati Formation,

185 which probably corresponds to the period covering the draining out of lake water, is
186 characterized by thin interbeds of silt and sand (for further details of the RB core,
187 see Sakai et al., 2001b). The chronology of the RB core has been constructed by
188 Hayashi et al. (2009), Mampuku et al. (2008) and Hayashi (2007) using ^{14}C
189 accelerator mass spectroscopy (AMS) dating and fine tuning of a pollen wet and
190 dry index record to the SPECMAP $\delta^{18}\text{O}$ stack record (Imbrie et al., 1984). Their
191 age-depth model of the RB core gives 15 ka at 11 m depth, which marks the
192 boundary between the Patan and the Kalimati Formations, and 76 ka at 45 m depth
193 (Fig. 3).

194 Each sample was first dried in an air-bath at 60°C for one day and then weighed.
195 The clay fraction under 2 μm was separated from each sample by gravity
196 sedimentation. Then, about 200 mg of this fraction was collected by the Millipore[®]
197 filter transfer method using the Gelman[®] GA-9, 0.45 μm pore, 47 mm diameter
198 Metrical[®] filter to provide optimal orientation (Moore and Reynolds, 1989). The
199 thickness of the clay cake formed on the filter was over 15 mg/cm^2 , which was
200 adequate for XRD quantitative analysis (Moore and Reynolds, 1989). The clay
201 cake was then transferred onto a glass slide. For each sample, both air-dried (AD)
202 and ethylene glycol solvated (EG) preparations were made. The EG preparation
203 was carried out to expose the sample to the vapor of the reagent in desiccator for
204 over 8 hrs at 60°C. On the other hand, the non-clay fraction of over 2 μm size was
205 dried and weighed to estimate the amount of the clay fraction.

206 All the XRD data were collected on a Rigaku X-ray Diffractometer RINT 2100V,
207 using $\text{CuK}\alpha$ radiation monochromatized by a curved graphite crystal in a step of
208 0.02° with a step-counting time of 4 seconds.

209
210

211 **2.2. XRD Decomposition and clay mineral analysis**

212 The decomposition (profile fitting) procedure of Lanson (1997) was followed to

213 obtain peak position, FWHM, and intensity (peak area) for each elementary peak,
 214 which were used for determination of the percentages of clay minerals and illite
 215 crystallinity. The XRD raw data were converted into ASCII format, transferred to
 216 an Apple Power Macintosh computer, and treated with a scientific graphical
 217 analysis program XRD MacDiff (Petschick, 2000). Basically, the treatment of a
 218 raw file begins with preliminary smoothing to decrease the effect of statistical
 219 counting errors. Then, a background was subtracted to eliminate most of its
 220 contribution to the peaks. Finally, the elementary peak fitting was done. All
 221 decompositions were performed with symmetrical elementary peaks with Gaussian
 222 shape. Fig. 4 shows a result of decomposition with five elementary peaks which
 223 correspond to smectite, chlorite, I-S (R=1), PCI and WCI, for the XRD pattern of
 224 AD sample. To check the reproducibility and detect the errors in the procedure, the
 225 decomposition was repeated four times for an XRD pattern of each sample and was
 226 also performed for four XRD patterns collected for each given sample. The errors
 227 on the measurement of FWHM and peak area were < 1% and <3%, respectively.

228 The percentage of each clay mineral in the core sediment samples was
 229 determined by the Mineral Intensity Factor (MIF) method (Moore and Reynolds,
 230 1989):

231

$$232 \quad CM_i (\%) = 100 \times (I_i / MIF_i) / \sum (I_i / MIF_i) \quad (i = 1, 2, \dots, n) \quad (1)$$

233

234 where CM_i is the percentage of clay mineral i and I_i is an integrated peak intensity
 235 for clay mineral i . The quantity MIF_i is the calibration constant for the diffraction
 236 peak used for clay mineral i that allows for quantitative estimation of its proportion
 237 in a mixture with clay mineral i' , and can be written as:

238

$$239 \quad MIF_i = MRI_i / MRI_{i'} \quad (2)$$

240

where MRI_i , called Mineral Reference Intensity, is the theoretical integrated intensity of clay mineral i under specified instrumental operating conditions. MRI was calculated by the computer program NEWMOD[®] (Reynolds and Reynolds, 1996), following the procedure suggested by Moore and Reynolds (1989). Note that the procedure forces the analysis to total 100%.

Illite crystallinity was estimated using the LI (Lanson, 1997) and MLI (Kuwahara et al., 2001). The LI is expressed thus:

$$LI = 0.1 / [PCI \text{ peak relative intensity} \times PCI \text{ peak FWHM} \times (PCI \text{ peak position} - WCI \text{ peak position})] \quad (3)$$

where

$$PCI \text{ peak relative intensity} = PCI \text{ intensity} / (PCI \text{ intensity} + WCI \text{ intensity} + I-S \text{ intensity}). \quad (4)$$

The MLI is defined thus:

$$MLI = PCI \text{ peak relative intensity} \times PCI \text{ peak FWHM} \times (PCI \text{ peak position} - WCI \text{ peak position}) \quad (5)$$

where

$$PCI \text{ peak relative intensity} = PCI \text{ intensity} / (PCI \text{ intensity} + WCI \text{ intensity}). \quad (6)$$

It is to be noted that the higher LI and the lower MLI, indicate higher illite crystallinity.

3. Results

The amount of the clay fraction in the core sediments of the Kalimati Formation between 12 m and 45 m depth varies between 2 wt% and 34 wt%, with an average

of 14 wt%. In the topmost part between 7 m and 12 m depth, which is composed of the sandy beds of the Patan Formation and the topmost part of the Kalimati Formation, the amount of clay fraction is much less (1 ~ 8 wt%, average 5 wt%) (Fig.5a). The Kalimati Formation between 12 m and 45 m depth consists of three clay-poor zones (17.8 ~ 22.0 m, 30.8 ~ 33.2 m and 41.1 ~ 44.8 m in depth) in which the amount of clay minerals is almost less than the average, except in some thin clay-enriched parts. The poorest part of the clay fraction in the clay-poor zones is at 19.6 ~ 22.0 m depth. In the other zones, the clay fraction varies around the average (14 wt%) at relatively short intervals (0.4 ~ 1 m), with some clay-rich peaks (> 20 wt%).

The clay minerals in the core sediments include illite, kaolinite, chlorite, I-S (R=1), and smectite (Figs. 5c and 6). Among these, illite is the most dominant one (50 ~ 80% in the clay fraction, average 61%), followed by kaolinite (7 ~ 30% in the clay fraction, average 19%). The morphology and crystal structure (polytype) of illite in the basin sediments clearly suggest that the illite is detrital (Kuwahara, 2006). The curve depicting the variations in the percentage of illite is a mirror image of the corresponding curve for kaolinite. Chlorite (3 ~ 9% in the clay fraction), I-S (R=1) (4 ~ 12% in the clay fraction), and smectite (traces to 8% in the clay fraction) are of lesser importance in the clay fraction. In addition, the amounts of the three clay minerals “in the sediments” are extremely low; it is particularly so of smectite whose amount does not reach even 1 wt% (Fig. 5d). Besides the clay minerals, the sediments are composed of detrital and precipitated minerals (quartz, feldspars, micas, calcite) (Paudel et al., 2004), amorphous silica (diatom shell) (Hayashi, 2007; Hayashi et al., 2009), and organic materials (Mampuku et al., 2008).

The illite crystallinity indices (LI and MLI) also appear to vary in accordance with the variations in the amount of the clay fraction or of illite and kaolinite (Fig. 5b). The MLI in the Kalimati Formation varies between 0.1 and 0.27, with an

average of 0.17. In the high illite crystallinity zones (17.8 ~ 21.7 m, 30.6 ~ 33.0 m and 39.8 ~ 44.5 m depth), the MLI is almost less than the average, except in certain zones where some peaks can be seen denoting slightly high MLI. These three high illite crystallinity zones overlap the three clay-poor zones mentioned above. The MLI in the other zones appears to fluctuate around the average at relatively short intervals (0.4 ~ 1 m), with some high peaks (> 0.2) that indicate low illite crystallinity.

4. Discussion

4.1. Weathering and erosion processes in the Kathmandu Basin

The detrital minerals in the Kathmandu Basin sediments are mostly micas (mainly muscovite), feldspars and quartz for which the source rocks could have been the gneisses and granites of the Shivapuri injection complex and weakly metamorphosed rocks of the Phulchauki Group (Sakai, 2001). Besides these, no other source rock, such as hydrothermal ore body, which could have contributed clay minerals, is known from or near the basin. The illitic minerals – the most dominant clay minerals – in the basin sediments, therefore, could have been formed by the exfoliation of micas during weathering, and were eroded and transported from the surrounding mountains by rainfall and run-off. In the Kalimati Formation, the percentage of illite in the clay fraction decreases with increase in the total amount of the clay fraction (Figs. 5a and c, Fig. 7(e)). In addition, illite crystallinity becomes low when the percentage of illite in the clay fraction decreases (Fig. 7(d); note that the higher MLI indicate lower illite crystallinity). Hence, while the amount of clay minerals fed to the Kathmandu Basin, increased with intensification of chemical weathering or hydrolysis, the amount and crystallinity of illite, derived from parent micas, are expected to have been reduced.

Kaolinite, the other dominant clay mineral in the basin sediments, and I-S ($R=1$)

325 have clear negative correlations with illite (Figs. 7(a) and (b)). That is, the amount
326 of kaolinite, as also of I-S ($R=1$), increases with decrease in the amount of illite,
327 while the amount of clay minerals increases in and around the Kathmandu Basin
328 (Figs. 5 and 7). Kaolinite is typical of warm and humid areas with good drainage
329 conditions (Robert, 2004). Precipitation plays a key role in mineral deposition by
330 exposing fresh rock and mineral surfaces to chemical and physical weathering and
331 transporting the eroded minerals to the depositional basins. Steep continental relief
332 reinforces the role of precipitation and run-off in chemical weathering and erosion
333 (Chamley, 1989; Robert, 2004). Therefore, warm and wet conditions and steep
334 relief in and around the Kathmandu Basin could have contributed to the formation
335 of kaolinite.

336 Smectite in the clay fraction has no correlation with illite (Fig. 7(c)). In addition,
337 the amount of smectite in the Kalimati Formation does not anywhere reach even 1
338 wt% (Fig. 5(d)). Smectite, therefore, can not be considered the main clay mineral
339 or the main secondary clay mineral to have been derived from alteration of illite in
340 the Kathmandu Basin, although it is also indicative of warm and intense chemical
341 weathering. This does not, however, contradict that smectite occurs in areas of low,
342 rather than steep, relief characterized by alternating episodes of precipitation and
343 aridity (Chamley, 1989, Robert, 2004).

344 Based on these facts, the following transformations are inferred to have taken
345 place in this area during the weathering process:

346

347 micas (mainly muscovite) \rightarrow illite

348 (\rightarrow illite-smectite mixed layer mineral ($R=1$)) \rightarrow kaolinite.

349

350 Also, during this process, the feldspars must have altered mainly to kaolinite
351 (Chamley, 1989). With intensification of chemical weathering and consequent
352 erosion in and around the Kathmandu Basin, hydrolysis and leaching of parent

minerals were activated, followed by degradation of illite – derived from alteration of micas – (lowering of crystallinity and transformation to I-S), and finally the formation of kaolinite via illitic minerals (illite and I-S). With the waning of chemical weathering and erosion, the formation of clay minerals and the transformation of parent minerals to kaolinite would slow down, and consequently the clay mineral content in the sediments would decrease, resulting in a low kaolinite to illite ratio.

Also, the sedimentation rates differ between intensified and weakened chemical weathering and erosion conditions. Based on the age-depth model of the RB core (Hayashi et al., 2009; Mampuku et al., 2008; Hayashi, 2007), the average sedimentation rate in the clay-poor zone, between 18 m and 22 m depth (ca. 19 ~ 31 ka), is estimated to be about 50 cm/kyr, and that in the clay-rich parts (13 ~ 18 m (17 ~ 19 ka) and 22 ~ 28 m (31 ~ 44 ka)), below and above the clay-poor zone, are about 100 cm/kyr, twice that in the clay-poor zone (Fig. 8). Unfortunately, the details of the sedimentation rates between 28 m and 45 m depth (44 ~ 76ka) are unclear because the age between them is based only on one single datum (tie-point between 47.5 m depth and MIS 5.1 (81 ka) (Hayashi et al., 2009)). However, it is certain that the sedimentation rate in the clay-rich zone was faster than that in the clay-poor zone. Supposing that the sedimentation rate in the clay-rich zone below 28 m depth was twice as much as that in the clay-poor zone, in the same way as above 28 m depth, the former is estimated to be about 60 cm/kyr and the latter is about 30 cm/kyr (Fig. 8).

In the clay-poor zones corresponding to weakened chemical weathering and erosion conditions (or dry climate), unique laminite beds with alternating very thin white calcite-rich and black carbonaceous clayey layers (20 ~ 30 pairs/cm), which are not contradictory to the low sedimentation rates in the clay-poor zones, are recognized (Sakai, 2001; Paudel et al., 2004; Kuwahara, 2006). Paudel et al. (2004) and Kuwahara (2006) have reported that the calcite particles in the laminite beds

are very fine ($\sim 50 \mu\text{m}$), euhedral, and authigenic (precipitated in lake water). The calcite could have been formed under conditions tranquil enough to facilitate formation of laminite beds, when calcite concentration in lake water was high because of low precipitation and run-off and consequent shrinkage of the Paleo-Kathmandu Lake during dry climates. The fossil diatom study on the same core yielded similar evidence of falling lake-level during the dry climate intervals (Hayashi, 2007; Hayashi et al., 2009). Also, calcite formation was reported by the mineralogical study of the JW-3 core, which was drilled near the RB core site (Fujii et al., 2001). Such authigenic calcite in the Kathmandu Basin sediments, therefore, serves not only as an important indicator of dry climate but also as a key mineral in correlation of (core) sediments.

4.2. Variations in dry-wet conditions in the Kathmandu Basin and monsoonal response to insolation forcing

From the results of clay mineral analysis of the Kathmandu Basin sediments, three main dry climate intervals (clay-poor, high illite crystallinity, low kaolinite/illite (K/I) ratio zones) and four wet climate intervals (clay-rich, low illite crystallinity, high K/I ratio zones) were recognized between 17 and 76 ka. The three dry climate intervals are estimated to be 19 ~ 31 ka, 44 ~ 51 ka, and 66 ~ 75 ka (Fig. 8). The records prior to 17 ka (in the topmost part between 7 m and 12 m depth, which is composed of the sandy beds of the Patan Formation and the topmost part of the Kalimati Formation) are not suitable for the reconstruction of paleoclimate. The variation record of dry-wet climate in and around the Kathmandu Basin depicted by the clay mineral proxies (e.g., the K/I ratio) is very similar to that revealed by the pollen analysis of the same core (Fujii et al., 2004) (Fig. 8).

In the variation record of dry-wet climate in this area, one can observe a strong

409 long-term variation in the 23,000 years precessional cycle of solar radiation, as dry
410 maxima are centered around 25, 47 and 70 ka, corresponding to the northern
411 hemisphere summer insolation (NHSI) signal (Fig. 8). Similar results have been
412 obtained by the $\delta^{18}\text{O}$ record of the planktonic foraminifera *Globigerinoides ruber*
413 (Schultz et al, 1998) and carbonate content (Leuschner and Sirocko, 2003) in the
414 sediment cores of the Arabian Sea. The maximum signal at 47 ka is not seen in the
415 SPECMAP stack or GISP2 $\delta^{18}\text{O}$ records (Blunier and Brook, 2001). These signals
416 around Marine Isotope Stage (MIS) 3 may be of regional significance to Indian
417 monsoonal variability (Schulz et al., 1998). The dry maximum signal around 25 ka
418 is consistent with the coldest part, which corresponds to the last glacial maximum
419 (LGM), as revealed by the pollen analysis of the same core (Fujii et al., 2004).

420 Leuschner and Sirocko (2003) constructed an Indian Summer Monsoon Index
421 (ISMI) that is defined as the insolation difference between 30°N and 30°S on 1
422 August, based on the fact that the modern Indian summer monsoon is mainly driven
423 by low pressure over the Himalayan-Tibetan Plateau and high pressure over the
424 southern subtropical Indian Ocean. The ISMI or NHSI signal (21 June, perihelion)
425 showing the 23,000 years precessional tempo, leads the global ice volume record as
426 indicated by the SPECMAP stack by several thousand years (Ruddiman, 2001,
427 Leuschner and Sirocko, 2003, Wang et al., 2005). Further, Clemens and Prell
428 (2003) show that Arabian Sea summer monsoon stack and factor lag behind the
429 NHSI signal (21 June, perihelion) by about 8,000 years and behind the ice volume
430 record by about 3,000 years, at the precession band (23 kyr). Similar results on the
431 long lag of the monsoon record were also obtained from the windblown lake
432 diatoms in the sediment cores from the tropical Atlantic Ocean, as a proxy of the
433 North African monsoon (lagging behind the NHSI (21 June, perihelion) by 5,000 ~
434 6,000 years) (Pokras and Mix, 1987). Ruddiman (1997, 2001), however, suggests
435 that the net lag of the North African monsoon signal behind the NHSI (21 June,
436 perihelion) is probably only 1,000 ~ 2,000 years (not 5,000 ~ 6,000 years) because

437 of the delayed diatom deposition in the Atlantic Ocean.

438 The variation records of dry-wet condition in this area, deduced from clay
439 minerals as well as fossil pollen proxies, appear to follow the summer insolation
440 with no long lag, especially from the coincidence of the dry climate zones and the
441 low ISMI and NHSI intervals (Fig. 8) and the results of the cross-correlation
442 analysis of the NHSI with the record of the K/I ratio (Fig. 9). The ISMI and NHSI
443 in Fig. 8 were recalculated using the 1 July summer insolation signal, because
444 nowadays the summer monsoon precipitation in this area reaches its peak in July
445 (Meteorological Forecasting Division, Government of Nepal, 2006). We also show
446 in Fig. 8 variation curves of the K/I ratio and pollen dry index depicted based on the
447 age model reconstructed using the sedimentation rates below 28 m depth of the RB
448 core mentioned above and a tie point between 47.5 m depth and 82.5 ka
449 corresponding to one of the maxima of the NHSI at 1 July (instead of 81 ka (MIS
450 5.1)), as well as those depicted using the age-depth model of Hayashi et al. (2009),
451 Mampuku et al. (2008) and Hayashi (2007). Using the summer insolation signal at
452 21 June, perihelion, the dry-wet record in this area appears to lag slightly behind the
453 NHSI (by ~ 1,000 years) (Fig 9(b)).

454 However, the centers (or wet maxima) of the wet climate zones depicted by the
455 K/I ratio do not always coincide with the NHSI maxima. The wet interval between
456 32 and 44 ka leads the NHSI while the wet interval between 51 and 66 ka appears to
457 lag slightly behind the NHSI (Fig. 8). These results likely show that the K/I record,
458 especially during the wet intervals, is not as simple as a direct response to insolation
459 forcing. Other factors, in addition to insolation forcing, may play important roles in
460 weathering, erosion, and sedimentation processes or may complicate paleoclimatic
461 interpretation of clay mineral (e.g., lake level change and lake water flow that affect
462 the distribution of particle size of minerals or dispersion of clay minerals).

463 On the other hand, the dry-wet record leads markedly the SPECMAP stack (the
464 ice volume record) and $\delta^{18}\text{O}$ record of the planktonic foraminifer *G. ruber* in

465 sediment core of the Arabian Sea (Schultz et al, 1998) (by about 5,000 years) (Figs.
466 8 and 9(b)). Wang et al. (2005) suggest that if changes in monsoon strength take
467 place before changes in ice volume, then monsoon variance is definitely not driven
468 by changes in high-latitude ice volume. The results of the present study reveal that
469 the Indian summer monsoon precipitation was strongly controlled by the northern
470 hemisphere summer insolation or summer insolation difference between the
471 Himalayan-Tibetan Plateau and subtropical Indian Ocean.

472 The question that arises here is “Why the long lag of monsoon behind the
473 summer insolation does not show up in the clay mineral proxies of this study”? One
474 possibility is that the samples are not from deep-sea sediments, but from
475 intermontane basin sediments on the southern slope of the central Himalaya. For
476 instance, the $\delta^{18}\text{O}$ of planktonic foraminifera in the sediment cores from the Indian
477 Ocean, which was interpreted as a monsoon proxy, was certainly affected by
478 changes in global ice volume as well as the local temperature of the Ocean water
479 (Ruddiman, 2001). Sediment grain size or planktonic foraminifera shell flux in the
480 deep-sea sediments was probably influenced by oceanic conditions (e.g., surface
481 and deep-sea currents, sea-level, water temperature and chemistry) (Singer, 1984;
482 Chamley, 1989; Tiwari et al., 2006). Such problems or intervention is not inherent
483 in the monsoon proxies (clay minerals, fossil pollen and diatom, etc.) of the
484 Kathmandu Basin sediments. In addition, the detritals would have transported and
485 deposited into the Paleo-Kathmandu Lake “in an instant” as compared with those of
486 the deep-sea sediments, because the basin had a diameter of only about 30 km and
487 the catchment area of the river is confined to the inside slope of the basin (Sakai,
488 2001; Kuwahara et al., 2001). The paleo-lake sediments of the Kathmandu Basin
489 must allow one to obtain direct and valuable information on the Indian monsoon
490 variability.

491
492

5. Conclusions

The clay mineral study of the Paleo-Kathmandu Lake sediments reveals how the weathering processes operated in this area, how the variations in the intensity of weathering and erosion were controlled by the Indian summer monsoon precipitation, and how the Indian summer monsoon responded to the summer insolation. During the wet climate intervals, intense chemical weathering and erosion promoted the formation of clay minerals, lowered illite crystallinity, and transformed illite to kaolinite. On the other hand, during the dry climate intervals, the weakening of chemical weathering and erosion processes retarded the formation of clay minerals and the transformation of illite to kaolinite. The sedimentation rates during the wet climate intervals were roughly twice as much as those during the dry climate intervals. The variation records of dry-wet condition in this area probably follow the summer insolation with no long lag, especially during the dry climate zones, whereas the wet maxima of the wet climate zones somewhat deviate from the strongest insolation forcing. In contrast, the wet-dry records were far ahead of the SPECMAP stack (the ice volume record), indicating that the Indian summer monsoon precipitation was not driven by changes in ice volume but by the northern hemisphere summer insolation or summer insolation difference between the Himalayan-Tibetan Plateau and the subtropical Indian Ocean. It is stressed here that the comparative studies between continental and marine records will have to be pursued further for a more comprehensive understanding of the Indian summer monsoon.

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525

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721

722

723 **Figure Captions**

724 Fig. 1. Outline geological map of the Kathmandu Basin showing the location of the
725 Rabibhawan (RB) core (modified from Sakai, 2001).

726

727 Fig. 2. A columnar section of the RB core from 5 m – 45 m depth (modified from
728 Sakai, 2001).

729

730 Fig. 3. An age-depth model of the RB core based on the AMS ^{14}C dating (above
731 30.1 m depth) and fine tuning of a pollen wet and dry index record to the
732 SPECMAP $\delta^{18}\text{O}$ stack record with the LR04 age model (Lisiecki and Raymo, 2005)
733 (below 30.1 m depth) (Hayashi et al., 2009; Mampuku et al., 2008; Hayashi, 2007).

734

735 Fig. 4. Decomposition with 5 elementary peaks of the XRD patterns obtained from
736 AD sample of clay minerals in the RB core. The dotted lines represent observed
737 profiles and solid lines calculated profiles. Gray lines represent the residuum.

738

739 Fig. 5. Variation records depicting (a) the amounts of clay fractions ($< 2\mu\text{m}$) in the
740 RB core sediments, (b) illite crystallinity indices (Li and MLI), (c) the percentage
741 of each clay mineral in the clay fraction, and (d) the amount of each clay mineral in
742 sediments. The clay-poor zones (see text) are shaded.

743

744 Fig. 6. Representative XRD patterns of AD and EG samples of clay minerals in the

745 RB core.

746

747 Fig. 7. Correlation plots between (a) illite and kaolinite, (b) illite and I-S ($R=1$), (c)
748 illite and smectite, (d) illite and MLI, (e) overall clay and illite, and (f) overall clay
749 and kaolinite. Open marks indicate data points in the clay-poor zones and solid
750 marks indicate those in the other. Symbol “ r ” in figure is a correlation coefficient,
751 and “ F ” is a result of F test ($F = (n-2) r^2 / (1-r^2)$, where “ n ” is the number of samples).
752 In this test, random sampling ($n = 90$ selected from total number, $N = 319$, between
753 12 m and 45 m depth) was performed to each correlation plot to improve the power
754 of F test and to avoid reducing of degrees of freedom in the records (Chelton, 1982).
755 There is a correlation between the two elements when the F value is larger than the
756 $F_{88}^1(0.05) = 3.95$.

757

758 Fig. 8. Comparison of the records of pollen dry index (Fujii, et al., 2004) and
759 kaolinite/illite ratio (this study) for the Paleo-Kathmandu Lake sediments with the
760 ISMI (30°N – 30°S, 1 July), NHSI (30°N, 1 July), SPECMAP $\delta^{18}\text{O}$ stack (Imbrie et
761 al., 1984), $\delta^{18}\text{O}$ of the planktonic foraminifera *G. ruber* in sediment cores 88/93KL
762 (Schulz et al., 1998) and Greenland GISP2 $\delta^{18}\text{O}$ ice record (Blunier and Brook,
763 2001). The records of pollen dry index and kaolinite/illite ratio are depicted based
764 on the age model reconstructed using the sedimentation rates below 28 m depth of
765 the RB core and a tie point between 47.5 m depth and 82.5 ka (see text) (solid lines),
766 as well as based on the age-depth model of Hayashi et al. (2009), Mampuku et al.
767 (2008) and Hayashi (2007) (dotted lines). Changes in sedimentary environment
768 and paleoclimate in and around the Kathmandu Basin are also shown. Climates
769 shown in parentheses in the paleoclimate box were indicated by Fujii et al. (2004).
770 The dry climate zones are lightly shaded. Paleoclimate in the topmost part that is
771 darkly shaded (8 ~ 17 ka) is uncertain (see text).

772

773 Fig. 9. (a) Cross-spectral analyses on the record of kaolinite/illite ratio with the
774 NHSI at 1 July and the SPECMAP $\delta^{18}\text{O}$ stack. (b) Cross-correlation of the NHSI at
775 1 July, NHSI at 21 June, and SPECMAP $\delta^{18}\text{O}$ stack with the record of
776 kaolinite/illite ratio. These analyses were done using the Analyseries software
777 (Paillard et al., 1996).
778

Figure 1

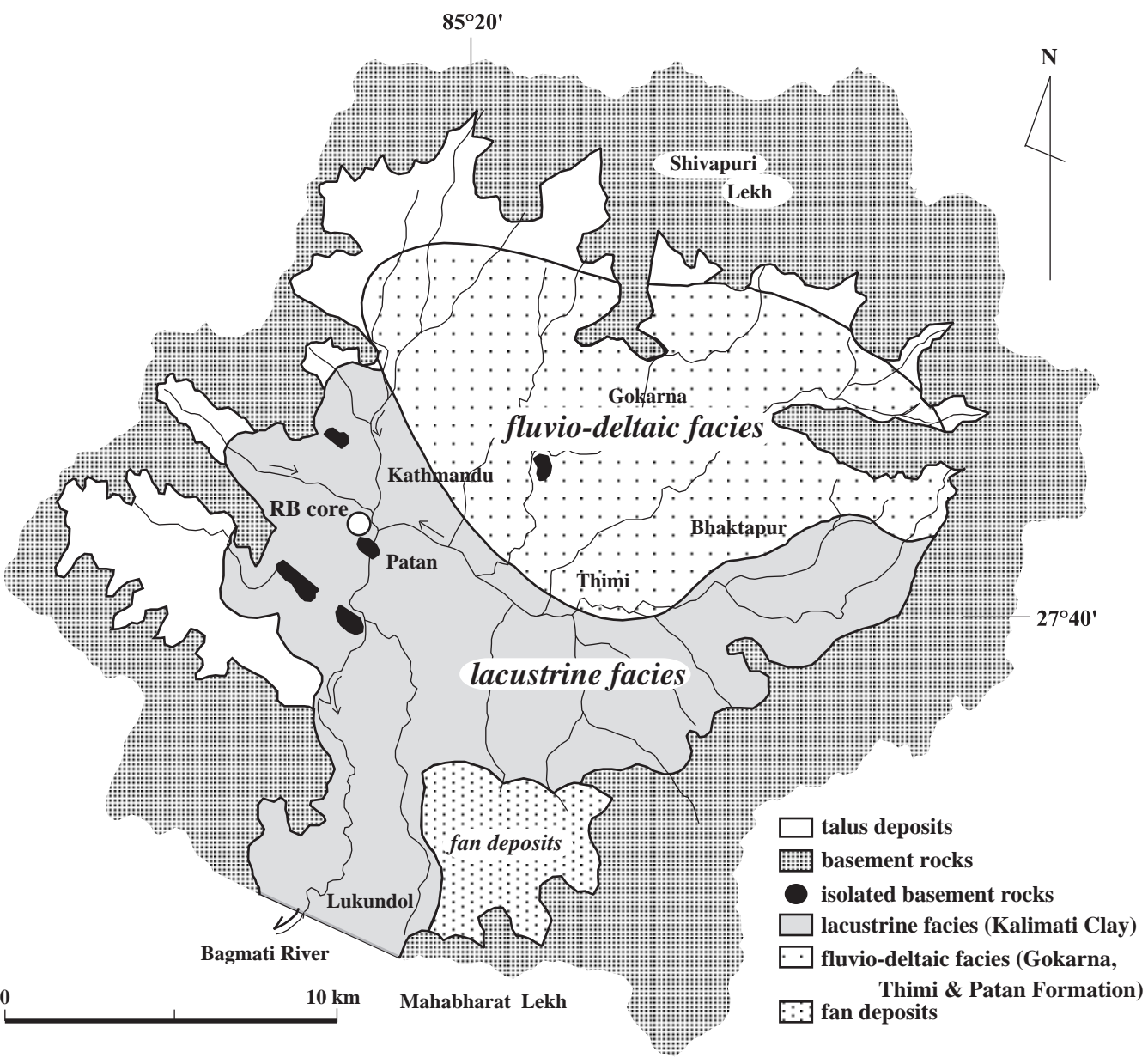


Figure 2

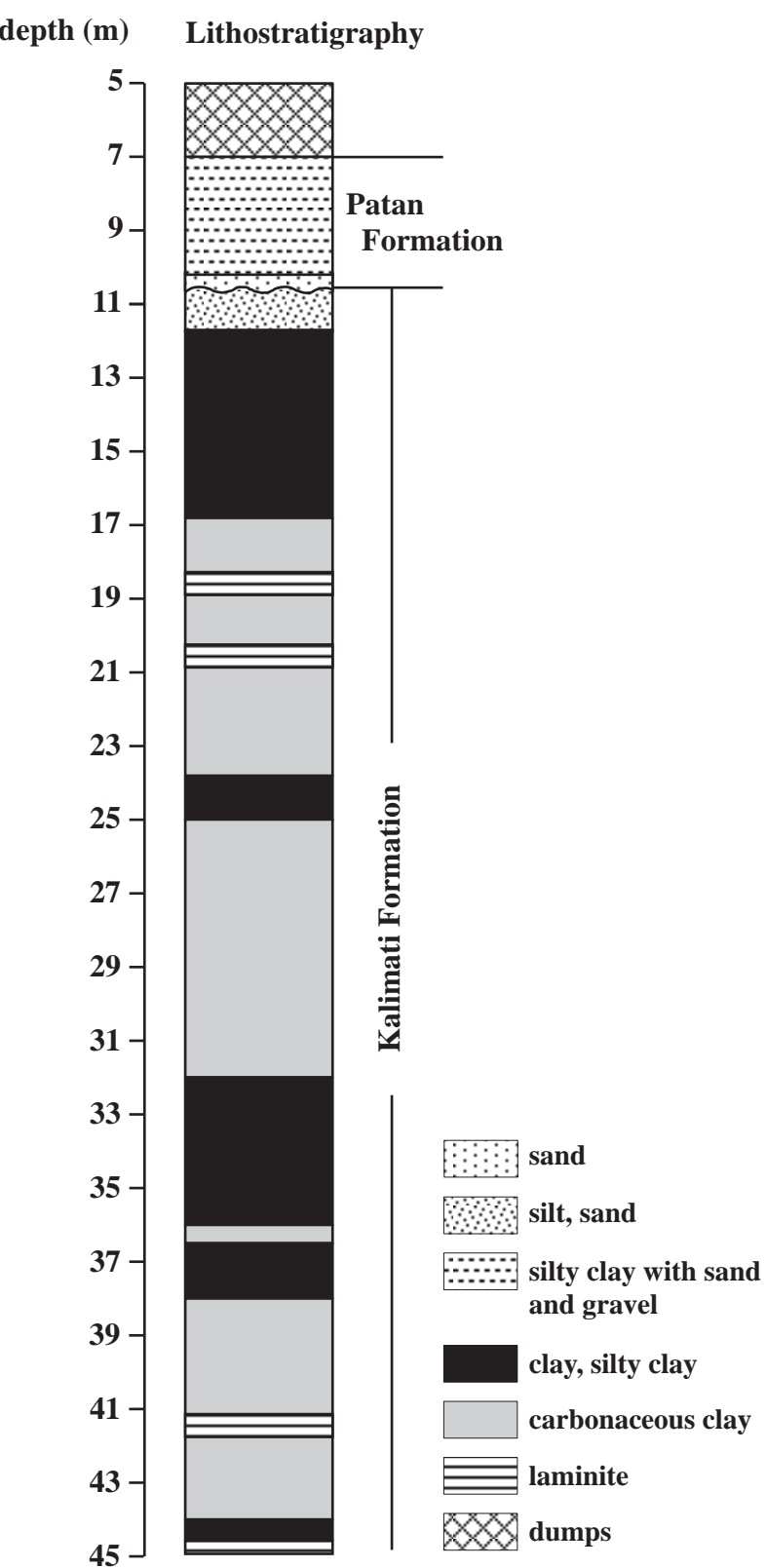


Figure 3

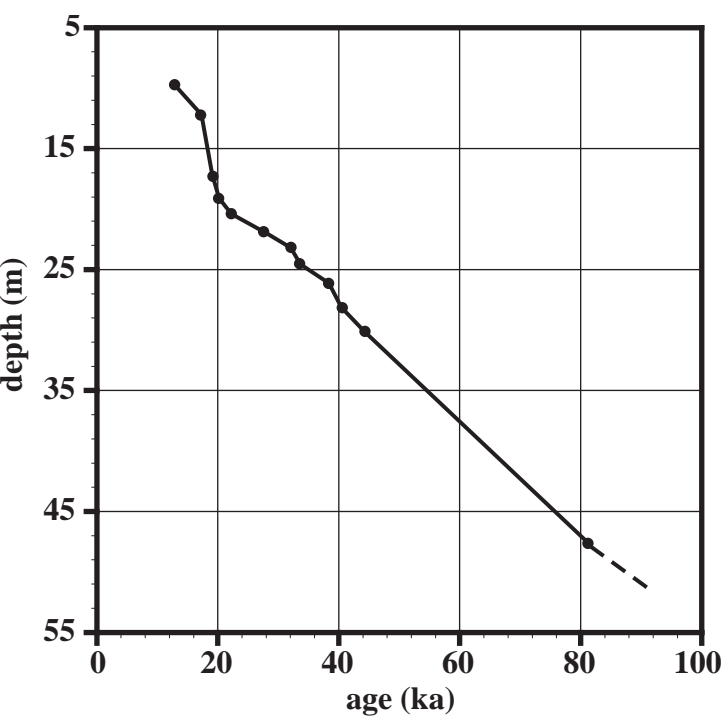


Figure 4

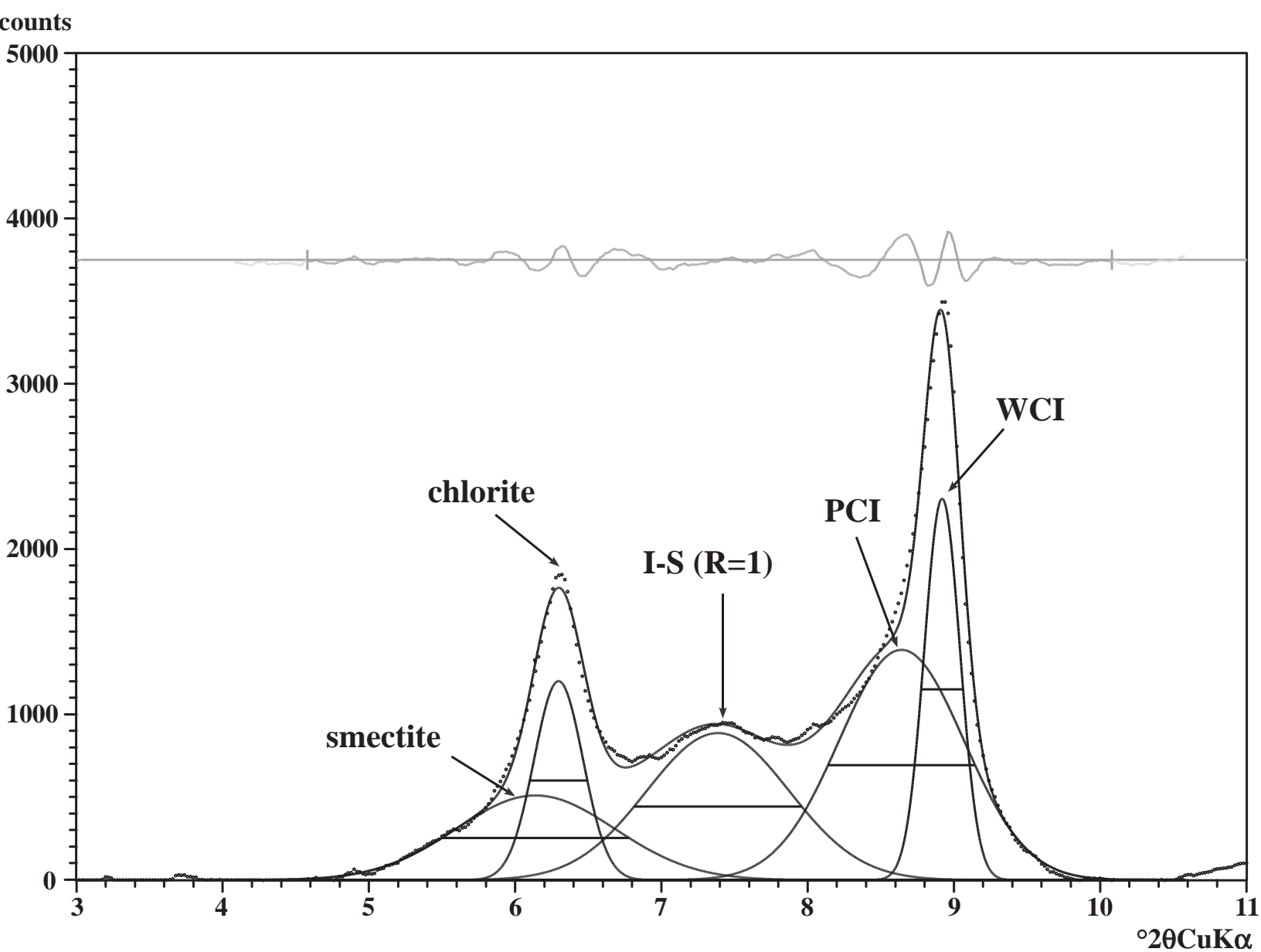


Figure 5

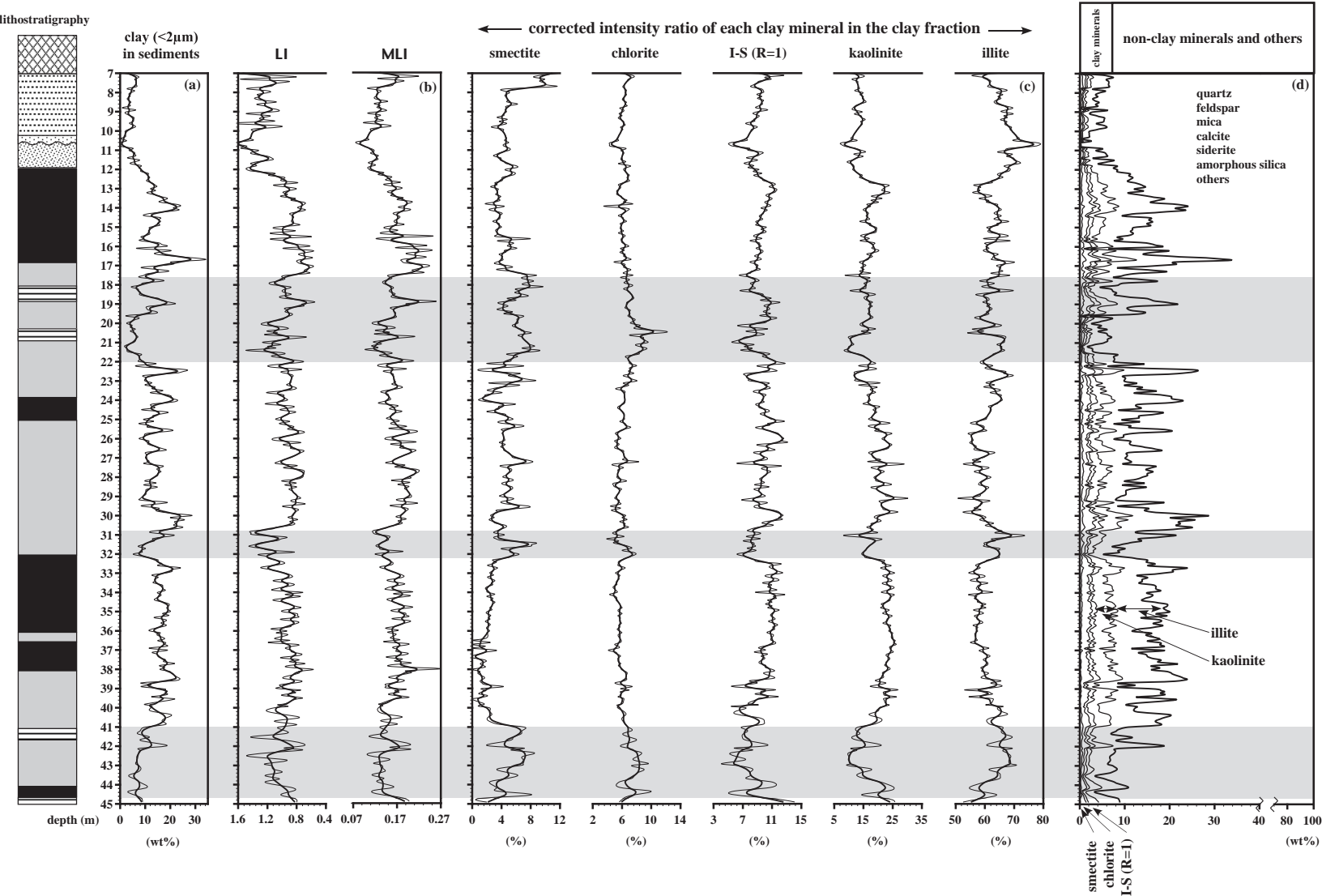


Figure 6

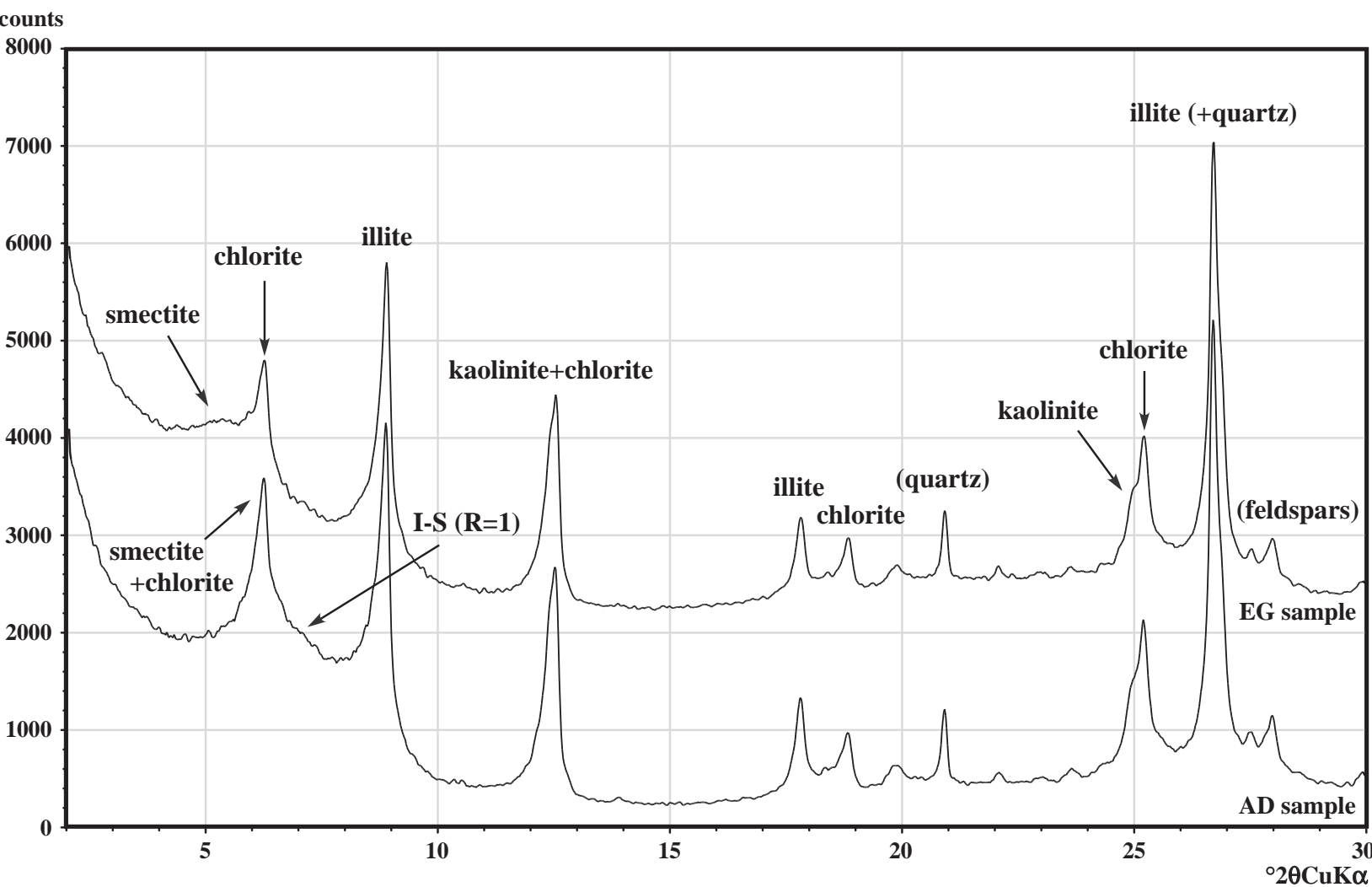


Figure 7
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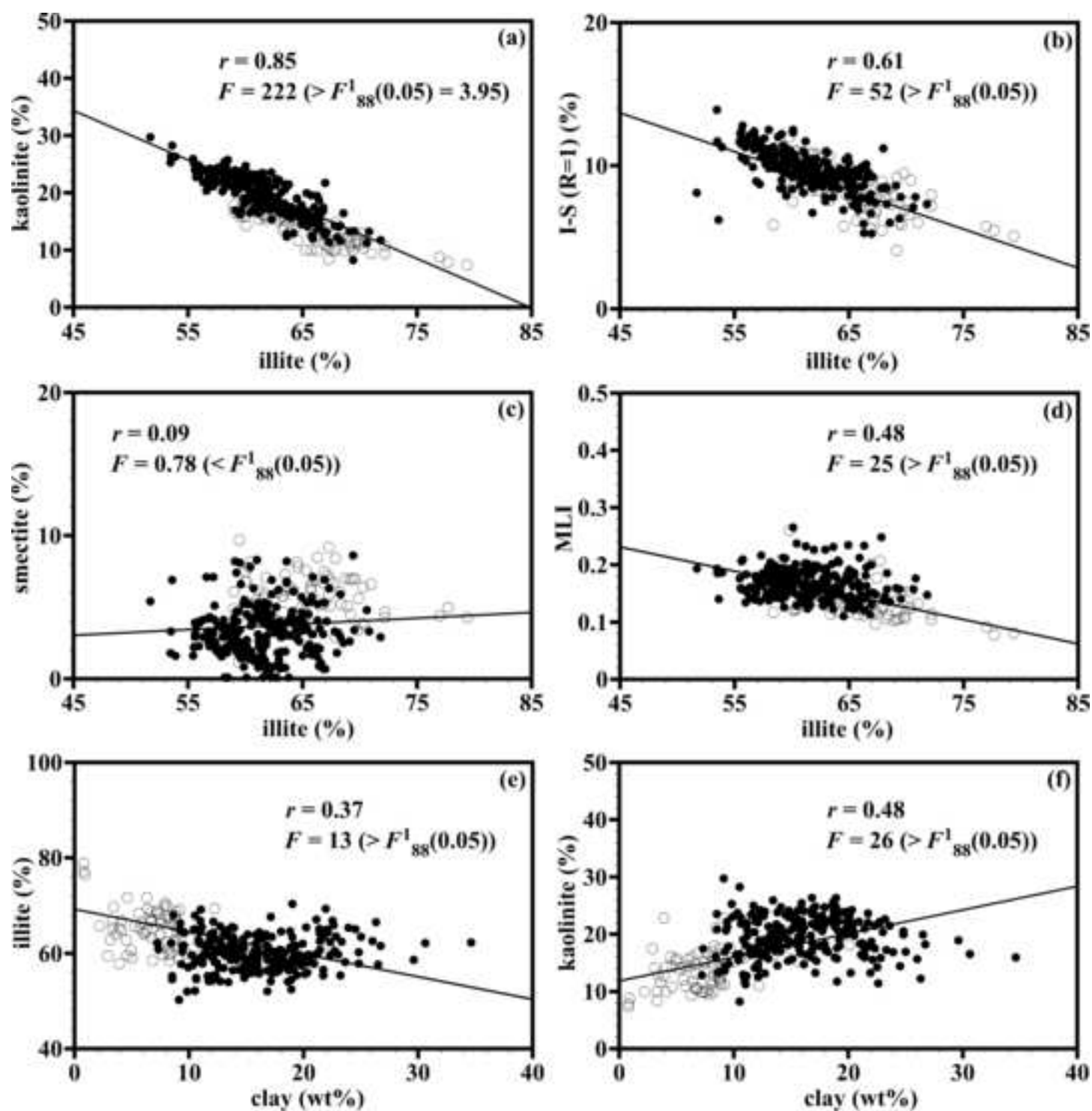


Figure 8

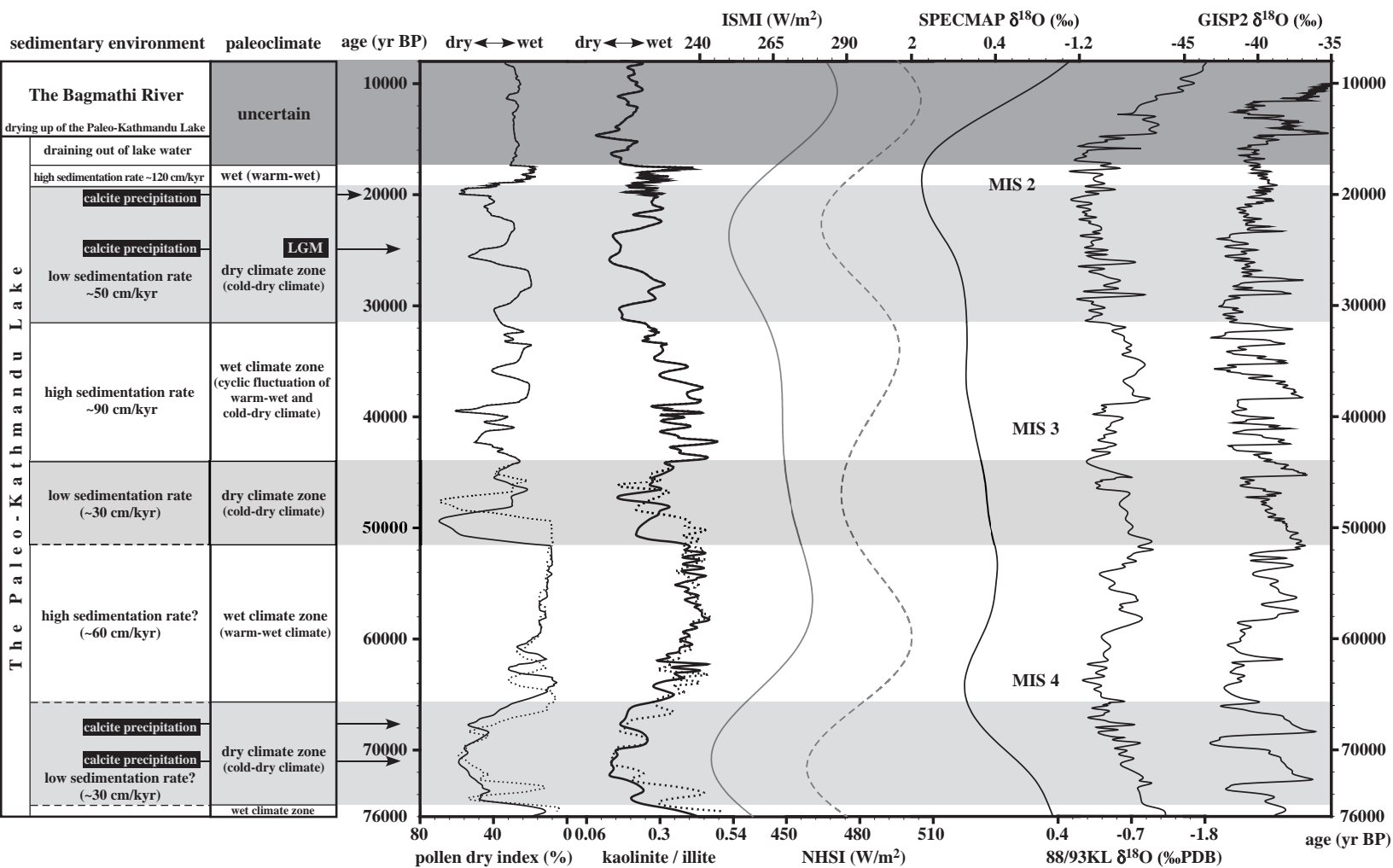


Figure 9

