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

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## Abstract

33	This paper reports the results of clay mineral analysis (the amount of clay
34	fraction, clay mineral assemblages, illite crystallinity) of samples collected from a
35	drilled core (Rabibhawan (RB) core) located in the west-central part of the
36	Kathmandu Basin on the southern slope of the Central Himalaya. The amount of
37	clay fraction in the core sediments between 12 m and 45 m depth (corresponding to
38	ca. 17 ~ 76 ka), which belong to the Kalimati Formation, is variable and shows
39	three clay-poor zones (19 ~ 31 ka, 44 ~ 51 ka, and 66 ~ 75 ka). The variations
40	correspond with those of illite crystallinity index (Lanson index (LI) and modified
41	Lanson index (MLI)) and kaolinite/illite ratio as well as the fossil pollen and diatom
42	records reported by previous workers. These data reveal the following
43	transformations occurring during the weathering process in this area:
44	
45	micas (mainly muscovite) $\rightarrow$ illite
46	$(\rightarrow$ illite-smectite mixed layer mineral (R=1)) $\rightarrow$ kaolinite.
47	
48	The sedimentation rate (~ 50 cm/kyr) of clay-poor zones that correspond to dry
49	climate intervals is only half that of clay-rich zones (~ 120 cm/kyr) that correspond
50	to wet climate intervals, indicating weakened chemical weathering and erosion and
51	low suspended discharge during dry climate intervals. The clay-poor zones
52	commonly show unique laminite beds with very fine, authigenic calcite, which was
53	probably precipitated under calm and high calcite concentration conditions caused
54	by low precipitation and run-off.
55	The variations between dry and wet conditions in this area as deduced from clay
56	minerals appear to follow the Indian Summer Monsoon Index (ISMI) $(30^{\circ}N - 30^{\circ}S,$
57	1 July) and northern hemisphere summer insolation (NHSI) signals (30°N) at 1 July,
58	especially during the dry climate zones, whereas the wet maxima of the wet climate
59	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.

- 68
- 69 Keywords: Indian monsoon, last glacial, paleoclimate, weathering, clay minerals,
- 70 Nepal Himalaya.
- 71
- 72

#### **1. Introduction**

74 The Indian monsoon system is one of the major weather systems on the Earth 75 and affects most densely populated regions. Differential heating during summer 76 results in a seasonal low pressure cell over the Indian continental landmass and a 77 high pressure cell over the cooler Indian Ocean. As a consequence, warm humid 78 southwest summer winds from the Indian Ocean flow onshore and contribute most 79 to the rainfall (Colin et al., 1998; Kudrass et al., 2001; Rashid et al., 2007). Most of 80 the monsoonal precipitation falls on the catchments of the 81 Ganges-Brahmaputra-Meghna (GBM) river system, whose rivers drain most of the 82 Himalayas and the northern Indian subcontinent (Kudrass et al., 2001; 83 France-Lanord et al., 2003; Rashid et al., 2007). Water and suspended discharge of 84 the river system, therefore, get concentrated during only five months (June to 85 October) of the summer monsoon (Islam et al., 2002; Goodbred, 2003). Natural 86 calamities, such as flooding or landslide, are also more frequent during this season 87 (Rashid et al, 2007).

88 It is naturally expected that past modifications of the intensity of chemical and 89 physical weathering and erosion of the Himalayan and Burman ranges and the 90 GBM catchments are strongly related to past variations in the strength of the Indian 91 summer monsoon (Colin et al. 1998). It is known that numerous paleoclimatic 92 studies, based on several proxies such as % Globigerina bulloides, organic carbon 93 content, lithogenic grain size, and pollen content, have permitted reconstruction of 94 changes in the paleo-monsoon intensity (e.g., Anderson and Prell, 1993; Sirocko et 95 al., 1993; Overpeck et al., 1996; Schulz et al., 1998; Ivanochko et al., 2005). These 96 proxies, however, are generally of monsoon wind strength and monsoon 97 wind-induced upwelling, rather than precipitation (Tiwari et al., 2006; Rashid et al., 98 2007; Shakun et al., 2007). Rashid et al. (2007) state that summer monsoonal 99 precipitation on the Indian subcontinent is not linearly correlated to wind strength, 100 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 Therefore, clay minerals can be useful indicators of paleoclimatic 2004). 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 However, paleoclimatic interpretation of clay minerals or other 2004). 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).

Moreover, Kübler Index (KI) (Kübler, 1964), which is conventional illite crystallinity index defined by the full width at half maximum intensity (FWHM) of 10 Å illite X-ray diffraction (XRD) peak and has extensively been used to reconstruct the paleoclimate (*e.g.*, Chamley 1989; Fukuzawa et al. 1997; Lamy et al. 2000), is not always available for estimation of illite crystallinity (Srodon, 1979, 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, 152because 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.
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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	<b>DVI</b> Derived in 2000 ( $0$ -let et al. 2001b) (Et al.) Esta elementaria esta la sector esta esta esta esta esta esta esta esta

PKL Project in 2000 (Sakai et al., 2001b) (Fig. 1). For clay mineral analysis, core 177178 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 Hayashi et al. (2009), Mampuku et al. (2008) and Hayashi (2007) using <sup>14</sup>C 188 189 accelerator mass spectroscopy (AMS) dating and fine tuning of a pollen wet and dry index record to the SPECMAP  $\delta^{18}$ O stack record (Imbrie et al., 1984). Their 190 191 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 2µm was separated from each sample by gravity 196 sedimentation. Then, about 200 mg of this fraction was collected by the Millipore<sup>®</sup> filter transfer method using the Gelman<sup>®</sup> GA-9, 0.45 µm pore, 47 mm diameter 197 Metricel<sup>©</sup> filter to provide optimal orientation (Moore and Reynolds, 1989). The 198 thickness of the clay cake formed on the filter was over  $15 \text{ mg/cm}^2$ , which was 199 200 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 2µm 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, using CuKa radiation monochromatized by a curved graphite crystal in a step of 207

- $208 \quad 0.02^{\circ}$  with a step-counting time of 4 seconds.
- 209
- 210

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

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

$$CM_i$$
 (%) = 100 × (I<sub>i</sub> / MIF<sub>i</sub>) /  $\sum$  (I<sub>i</sub> / MIF<sub>i</sub>) (i = 1, 2,...n) (1)

233

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

238

239  $MIF_i = MRI_i / MRI_i$ (2)

241	where $MRI_i$ , called Mineral Reference Intensity, is the theoretical integrated
242	intensity of clay mineral $i$ under specified instrumental operating conditions. MRI
243	was calculated by the computer program $\operatorname{NEWMOD}^{\mathbb{O}}$ (Reynolds and Reynolds,
244	1996), following the procedure suggested by Moore and Reynolds (1989). Note
245	that the procedure forces the analysis to total 100%.
246	Illite crystallinity was estimated using the LI (Lanson, 1997) and MLI
247	(Kuwahara et al., 2001). The LI is expressed thus:
248	
249	$LI = 0.1/[PCI peak relative intensity \times PCI peak FWHM$
250	$\times$ (PCI peak position – WCI peak position)] (3)
251	where
252	PCI peak relative intensity = PCI intensity /
253	(PCI intensity + WCI intensity + I-S intensity). $(4)$
254	
255	The MLI is defined thus:
256	
257	MLI = PCI peak relative intensity × PCI peak FWHM
258	$\times$ (PCI peak position – WCI peak position)] (5)
259	where
260	PCI peak relative intensity = PCI intensity / (PCI intensity + WCI intensity). (6)
261	
262	It is to be noted that the higher LI and the lower MLI, indicate higher illite
263	crystallinity.
264	
265	
266	3. Results
267	The amount of the clay fraction in the core sediments of the Kalimati Formation
268	between 12 m and 45 m depth varies between 2 wt% and 34 wt%, with an average

269 of 14 wt%. In the topmost part between 7 m and 12 m depth, which is composed of 270 the sandy beds of the Patan Formation and the topmost part of the Kalimati Formation, the amount of clay fraction is much less  $(1 \sim 8 \text{ wt\%}, \text{ average 5 wt\%})$ 271 272 (Fig.5a). The Kalimati Formation between 12 m and 45 m depth consists of three 273 clay-poor zones (17.8 ~ 22.0 m, 30.8 ~ 33.2 m and 41.1 ~ 44.8 m in depth) in which 274 the amount of clay minerals is almost less than the average, except in some thin 275 clay-enriched parts. The poorest part of the clay fraction in the clay-poor zones is at 276  $19.6 \sim 22.0$  m depth. In the other zones, the clay fraction varies around the average 277 (14 wt%) at relatively short intervals (0.4 ~ 1 m), with some clay-rich peaks (> 20 278 wt%).

279 The clay minerals in the core sediments include illite, kaolinite, chlorite, I-S 280 (R=1), and smectite (Figs. 5c and 6). Among these, illite is the most dominant one 281  $(50 \sim 80\%)$  in the clay fraction, average 61%), followed by kaolinite (7 ~ 30%) in the 282 clay fraction, average 19%). The morphology and crystal structure (polytype) of 283 illite in the basin sediments clearly suggest that the illite is detrital (Kuwahara, 284 2006). The curve depicting the variations in the percentage of illite is a mirror image of the corresponding curve for kaolinite. Chlorite  $(3 \sim 9\%)$  in the clay 285 286 fraction), I-S (R=1) (4 ~ 12% in the clay fraction), and smectite (traces to 8% in the 287 clay fraction) are of lesser importance in the clay fraction. In addition, the amounts 288 of the three clay minerals "in the sediments" are extremely low; it is particularly so 289 of smectite whose amount does not reach even 1 wt% (Fig. 5d). Besides the clay 290 minerals, the sediments are composed of detrital and precipitated minerals (quartz, 291 feldspars, micas, calcite) (Paudel et al., 2004), amorphous silica (diatom shell) 292 (Hayashi, 2007; Hayashi et al., 2009), and organic materials (Mampuku et al., 293 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.

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#### 4. Discussion

#### **4.1. Weathering and erosion processes in the Kathmandu Basin**

308 The detrital minerals in the Kathmandu Basin sediments are mostly micas 309 (mainly muscovite), feldspars and quartz for which the source rocks could have 310 been the gneisses and granites of the Shivapuri injection complex and weakly 311 metamorphosed rocks of the Phulchauki Group (Sakai, 2001). Besides these, no 312 other source rock, such as hydrothermal ore body, which could have contributed 313 clay minerals, is known from or near the basin. The illitic minerals - the most 314 dominant clay minerals - in the basin sediments, therefore, could have been formed 315 by the exfoliation of micas during weathering, and were eroded and transported 316 from the surrounding mountains by rainfall and run-off. In the Kalimati Formation, 317 the percentage of illite in the clay fraction decreases with increase in the total 318 amount of the clay fraction (Figs. 5a and c, Fig. 7(e)). In addition, illite crystallinity 319 becomes low when the percentage of illite in the clay fraction decreases (Fig. 7(d); 320 note that the higher MLI indicate lower illite crystallinity). Hence, while the 321 amount of clay minerals fed to the Kathmandu Basin, increased with intensification 322 of chemical weathering or hydrolysis, the amount and crystallinity of illite, derived 323 from parent micas, are expected to have been reduced.

324 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).

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

346

347 micas (mainly muscovite)  $\rightarrow$  illite

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

349

Also, during this process, the feldspars must have altered mainly to kaolinite (Chamley, 1989). With intensification of chemical weathering and consequent 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.

360 Also, the sedimentation rates differ between intensified and weakened chemical 361 weathering and erosion conditions. Based on the age-depth model of the RB core 362 (Hayashi et al., 2009; Mampuku et al., 2008; Hayashi, 2007), the average 363 sedimentation rate in the clay-poor zone, between 18 m and 22 m depth (ca. 19 ~ 31 364 ka), is estimated to be about 50 cm/kyr, and that in the clay-rich parts  $(13 \sim 18 \text{ m})$  m (17) 365 ~ 19 ka) and 22 ~ 28 m (31 ~ 44 ka)), below and above the clay-poor zone, are about 366 100 cm/kyr, twice that in the clay-poor zone (Fig. 8). Unfortunately, the details of 367 the sedimentation rates between 28 m and 45 m depth (44  $\sim$  76ka) are unclear 368 because the age between them is based only on one single datum (tie-point between 369 47.5 m depth and MIS 5.1 (81 ka) (Hayashi et al., 2009)). However, it is certain that 370 the sedimentation rate in the clay-rich zone was faster than that in the clay-poor 371 zone. Supposing that the sedimentation rate in the clay-rich zone below 28 m depth 372 was twice as much as that in the clay-poor zone, in the same way as above 28 m 373 depth, the former is estimated to be about 60 cm/kyr and the latter is about 30 374 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 381 are very fine (~ 50  $\mu$ m), euhedral, and authigenic (precipitated in lake water). The 382 calcite could have been formed under conditions tranquil enough to facilitate 383 formation of laminite beds, when calcite concentration in lake water was high 384 because of low precipitation and run-off and consequent shrinkage of the 385 Paleo-Kathmandu Lake during dry climates. The fossil diatom study on the same 386 core yielded similar evidence of falling lake-level during the dry climate intervals 387 (Hayashi, 2007; Hayashi et al., 2009). Also, calcite formation was reported by the 388 mineralogical study of the JW-3 core, which was drilled near the RB core site (Fujii 389 et al., 2001). Such authigenic calcite in the Kathmandu Basin sediments, therefore, 390 serves not only as an important indicator of dry climate but also as a key mineral in 391 correlation of (core) sediments.

392

393

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

396 From the results of clay mineral analysis of the Kathmandu Basin sediments, 397 three main dry climate intervals (clay-poor, high illite crystallinity, low 398 kaolinite/illite (K/I) ratio zones) and four wet climate intervals (clay-rich, low illite 399 crystallinity, high K/I ratio zones) were recognized between 17 and 76 ka. The 400 three dry climate intervals are estimated to be 19 ~ 31 ka, 44 ~ 51 ka, and 66 ~ 75 ka 401 (Fig. 8). The records prior to 17 ka (in the topmost part between 7 m and 12 m 402 depth, which is composed of the sandy beds of the Patan Formation and the topmost 403 part of the Kalimati Formation) are not suitable for the reconstruction of 404 paleoclimate. The variation record of dry-wet climate in and around the 405 Kathmandu Basin depicted by the clay mineral proxies (e.g., the K/I ratio) is very 406 similar to that revealed by the pollen analysis of the same core (Fujii et al., 2004) (Fig. 8). 407

408 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 obtained by the  $\delta^{18}$ O record of the planktonic foraminifera *Globigerinoides ruber* 412 413 (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 SPECMAP stack or GISP2  $\delta^{18}$ O records (Blunier and Brook, 2001). These signals 415 416 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 indicated by the SPECMAP stack by several thousand years (Ruddiman, 2001, 426 427 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, perihelion) is probably only 1,000 ~ 2,000 years (not 5,000 ~ 6,000 years) because 436

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 in Fig. 8 were recalculated using the 1 July summer insolation signal, because 443 444 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 K/I ratio do not always coincide with the NHSI maxima. The wet interval between 455 456 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  $\delta^{18}$ O record of the planktonic foraminifer *G* ruber in sediment core of the Arabian Sea (Schultz et al, 1998) (by about 5,000 years) (Figs. 8 and 9(b)). Wang et al. (2005) suggest that if changes in monsoon strength take place before changes in ice volume, then monsoon variance is definitely not driven by changes in high-latitude ice volume. The results of the present study reveal that the Indian summer monsoon precipitation was strongly controlled by the northern hemisphere summer insolation or summer insolation difference between the 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  $\delta^{18}$ O of planktonic foraminifera in the sediment cores from the Indian Ocean, which was interpreted as a monsoon proxy, was certainly affected by 477 478 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

#### **5.** Conclusions

494 The clay mineral study of the Paleo-Kathmandu Lake sediments reveals how the 495 weathering processes operated in this area, how the variations in the intensity of 496 weathering and erosion were controlled by the Indian summer monsoon 497 precipitation, and how the Indian summer monsoon responded to the summer 498 insolation. During the wet climate intervals, intense chemical weathering and 499 erosion promoted the formation of clay minerals, lowered illite crystallinity, and 500 transformed illite to kaolinite. On the other hand, during the dry climate intervals, 501 the weakening of chemical weathering and erosion processes retarded the 502 formation of clay minerals and the transformation of illite to kaolinite. The 503 sedimentation rates during the wet climate intervals were roughly twice as much as 504 those during the dry climate intervals. The variation records of dry-wet condition 505 in this area probably follow the summer insolation with no long lag, especially 506 during the dry climate zones, whereas the wet maxima of the wet climate zones 507 somewhat deviate from the strongest insolation forcing. In contrast, the wet-dry 508 records were far ahead of the SPECMAP stack (the ice volume record), indicating 509 that the Indian summer monsoon precipitation was not driven by changes in ice 510 volume but by the northern hemisphere summer insolation or summer insolation 511 difference between the Himalayan-Tibetan Plateau and the subtropical Indian 512 Ocean. It is stressed here that the comparative studies between continental and 513 marine records will have to be pursued further for a more comprehensive 514 understanding of the Indian summer monsoon.

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516

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- 526

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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 $5 \text{ m} - 45 \text{ m}$ 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 $\delta^{18}$ 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 (< $2\mu m$ ) 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, and "F" is a result of F test ( $F = (n-2) r^2/(1-r^2)$ , where "n" is the number of samples). 751 In this test, random sampling (n = 90 selected from total number, N = 319, between 752 753 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  $F^{1}_{88}(0.05) = 3.95.$ 756

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 ISMI (30°N – 30°S, 1 July), NHSI (30°N, 1 July), SPECMAP  $\delta^{18}$ O stack (Imbrie et 760 al., 1984),  $\delta^{18}$ O of the planktonic foraminifera *G. rubber* in sediment cores 88/93KL 761 (Schulz et al., 1998) and Greenland GISP2  $\delta^{18}$ O ice record (Blunier and Brook, 762 763 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).

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























