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1 **Evidence for centennial-scale Mid-Holocene episodes of hypolimnetic anoxia in a high-**  
2 **altitude lake system from Central Tian Shan (Kyrgyzstan)**

3  
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19

20 **Keywords**

21 Hyperspectral imaging, Paleolimnology, Paleoclimate, Lake stratification, Hypolimnetic  
22 anoxia, Chemocline, Central Asia, Lake Son Kol, Holocene

23

24

25

26 **Abstract**

27 Few sedimentary archives of lake meromixis are available in palaeolimnological records,  
28 because long-term observations are limited in time and indisputable sediment proxies of  
29 hypolimnetic anoxia are still scarce. Here we use visible and near infrared (VNIR), and short-  
30 wave infra-red (SWIR) hyperspectral imaging combined with geochemical analyses to  
31 reconstruct lake stratification history, redox status and mixing conditions in the water column  
32 of Lake Son Kol (Kyrgyzstan) in Central Tian Shan during the last 8500 years. In particular,  
33 the detection of Bacteriopheophytin *a* (Bphe *a*), a pigment produced by anoxygenic  
34 phototrophic bacteria at the chemocline of meromictic lakes, emphasizes episodes of multi-  
35 decadal to centennial hypolimnetic anoxia in Lake Son Kol. Phases of hypolimnetic anoxia  
36 are inferred from the deposition of dark organic sediments within a stratified lake system,  
37 which occurred during periods of increased snowmelt (and solid winter precipitation) and  
38 warmer spring/summer temperatures that promoted floods and the **export** of terrestrial  
39 material from the catchment. Prolonged euxinic conditions in bottom waters (involving the  
40 development of a stable chemocline) are reported around 8500, 8400, 8200–7800, 7700–  
41 7500, 7300–7000, 6500–6100, 6000–5700 and 5500–5250 cal. yr BP. At ca. 5250 cal. yr BP,  
42 the chemocline abruptly vanished as Lake Son Kol tipped into a regime with predominantly  
43 cooler and well-mixed conditions (predominance of oxygenic phototrophs), coeval with  
44 higher lake levels. The disappearance of hypolimnetic anoxia in Lake Son Kol coincides with  
45 strengthened wind conditions that imply enhanced lake overturning and upward mixing of  
46 nutrients in the water column. This study reveals the strong potential of hyperspectral  
47 imaging, in combination with more classical palaeolimnological approaches, to reconstruct  
48 the lake trophic and mixing history and explore the controlling mechanisms at work on  
49 decadal to centennial timescales. Our results outline how abrupt ecosystem changes may  
50 occur even in the absence of anthropogenic climate change.

51

52 **1. Introduction**

53 Anthropogenic global warming is changing high-altitude ecosystems and environmental  
54 dynamics (Solomon et al., 2007; IPCC 2013). Regional model outputs for Central Asia  
55 predict that seasonal surface temperatures in the region will increase from 3 °C up to 11.4 °C  
56 on average, coeval with a decline in the amount of precipitation (Ozturk et al., 2012),  
57 although uncertainties in hydrological/rainfall projections are existing. High elevation regions  
58 of Central Asia (e.g., Tian Shan and Pamirs mountains, Tibetan Plateau) are particularly  
59 vulnerable to warmer climate conditions (e.g., Christensen et al., 2007). In high-altitude areas,  
60 a consequence of global warming may be shorter and wetter (and milder) winters along with  
61 longer and drier summers (Christensen et al., 2007; IPCC, 2013). Increases in solid winter  
62 precipitation, spring/summer temperatures, and (summer) rainstorms are likely to increase  
63 flooding of snowmelt-fed lake basins. Warmer temperatures are also expected to lead to  
64 eutrophication of lakes, depletion of oxygen in deep waters (Jenny et al., 2013), and longer  
65 and stronger temperature stratification in summer (e.g., Adrian et al., 1995; Livingstone,  
66 2003; Foley et al., 2012). Incomplete lake overturning may result in the development of a  
67 chemocline (Davison, 1993) with an anoxic hypolimnion underneath (Butz et al., 2016),  
68 which may last for prolonged periods (meromixis). Meromixis implies persistent lake  
69 stratification over multiple years caused by chemical differentiation of the bottom waters, (cf.  
70 seasonal thermal stratification which is common in deep temperate lakes and may occur  
71 above the chemocline in a meromictic lake). Investigations of long-term trends in oxygen  
72 depletion in lakes, and their relationship with climatic changes on timescales beyond the  
73 instrumental period, are still scarce in high Central Asia.

74 Located at the intercept between the influences of the Westerlies and the Siberian Anticyclone  
75 (Chen et al., 2008; 2010), the Tian Shan holds key information about past climate variability

76 and particularly temperature and hydrological changes in western Central Asia (e.g.,  
77 Rousseau et al., 2020). A number of valuable and high-quality lake sediment studies have  
78 examined Holocene climate development in the Tian Shan and Pamir mountains (e.g.,  
79 Aichner et al., 2015; Beer et al., 2007; Li et al., 2011; Mischke et al., 2010; Chen et al., 2016;  
80 Heinecke et al., 2016, 2017, 2018; Rousseau et al., 2020), including several at Lake Son Kol  
81 (or Son Kul) in Kyrgyzstan (Huang et al., 2014; Mathis et al., 2014; Pacton et al., 2015;  
82 Lauterbach et al., 2014; Schwarz et al., 2017, Wolff et al., 2017; Laug et al., 2020). The focus  
83 of these works has been on the Mid- to Late Holocene, with however limited  
84 palaeoenvironmental reconstructions between 8500 and 6000 cal. yr BP. Thus there are  
85 virtually no high-resolution palaeolimnological records of long-term lake mixing conditions  
86 from this region yet, which could provide information on the complex links between  
87 environmental, climatic and anthropogenic changes in high-altitude settings.

88 Here we combine hyperspectral imaging (HSI) and X-Ray fluorescence (XRF) analyses of a  
89 new core from Lake Son Kol (SK06) to obtain an unprecedented record of sedimentary  
90 pigment changes at ultra-high sampling resolution ( $\mu\text{m}$ -scale; sub-annual resolution).

91 Building upon the work of Huang et al. (2014), who inferred that most of the annual  
92 precipitation switched from the summer to the cold (winter) season between the Mid- and  
93 Late Holocene, we used sedimentary pigments and geochemical data to (i) decipher lake  
94 trophic status during the last 8500 years, (ii) investigate when anoxic conditions developed in  
95 the hypolimnion, and (iii) assess whether the occurrence of hypolimnetic anoxia can be linked  
96 to climatic changes in Central Tian Shan. We further evaluate whether phase boundaries  
97 identified in the multi-proxy dataset can be attributed to event-based phase transitions or more  
98 gradual shifts using multivariate statistical analyses. Comparisons with other  
99 palaeolimnological data from Lake Son Kol and regional records will be used to understand  
100 regional climate development in the Tian Shan, as well as implications for lake stratification

101 dynamics during the Holocene.

102

## 103 **2. Study site**

### 104 *2.1. Geological, environmental and lacustrine setting*

105 The geology, geomorphology, climatic and lacustrine characteristics of Lake Son Kol have  
106 been extensively studied (Mathis et al., 2014; Huang et al., 2014; Lauterbach et al., 2014).  
107 Son Kol (41°50'52"N, 75°07'45"E, 3016 m a.s.l.) is the second largest natural lake in  
108 Kyrgyzstan, within the central Tian Shan mountain ranges (Fig. 1). It fills the central  
109 depression of an intramontane basin located between two generally east-west stretching  
110 mountain ranges, which are the Son Kol Too and Baidula Ranges (3200–3500 m a.s.l.) in the  
111 north and the Boor-Albas Range (3800–4000 m a.s.l.) in the south (Shitnikov, 1980). These  
112 mountain ranges are composed of Carboniferous sedimentary rocks and Permian granitoids  
113 (east and south) and Cambro-Ordovician metamorphic granodioritic and granitic gneisses, as  
114 well as strongly folded sedimentary rocks in the north  
115 (<http://portal.onegeology.org/OnegeologyGlobal/>). The modern equilibrium line altitude  
116 (ELA) of glaciers at north-facing slopes in the Central Tian Shan is ~3850–4100 m a.s.l.,  
117 (Narama et al., 2007; Koppes et al., 2008), hence there are no glaciers in the lake watershed  
118 today and, based on ELA reconstructions, seem unlikely to have fed Lake Son Kol during the  
119 Holocene (Savoskul and Solomina, 1996). The widespread pasture-covered plains  
120 surrounding the lake are made up of Quaternary eroded material arranged in old Pleistocene  
121 moraines (Abramowski et al., 2006; Narama and Okuno, 2006) and inactive fan deltas. A  
122 large fluvial fan-delta nourished by cirque valleys of the Boor Albas range progrades from the  
123 southern shore into the lake. Along the southeastern and northeastern shore, large coastal  
124 sedimentary features such as sand spits and associated open lagoons have been developed by  
125 the prevailing westerly winds, indicating a stable lake level during at least the Late Holocene.

126 Today, Lake Son Kol is located above the treeline, on a high plateau surrounded by mountain  
127 meadows. Because of pronounced altitudinal changes over short distances, the vegetation  
128 rapidly changes within a few kilometers with the development of a lush alpine meadow in the  
129 lake catchment at ~3000 m a.s.l., while conversely steppe to semi-desert environments prevail  
130 in open lowlands (below 2000 m a.s.l) with sagebrush steppe (*Artemisia*) and  
131 Chenopodiaceae (Mathis et al., 2014).

132 Lake Son Kol is ~29-km long and ~18-km wide, with a surface area of ~275 km<sup>2</sup> and a  
133 catchment of ~1130 km<sup>2</sup>. The lake water volume is ~2.82 km<sup>3</sup> and the lake is presently ~13-m  
134 deep (Lauterbach et al., 2014). According to the lake stratification classifications of Kalff  
135 (2002), the depth and climatic location of Lake Son Kol should make it susceptible to shifts in  
136 mixing regime. The lake has a single outlet at its southern end, which discharges into the  
137 Naryn River and then to the Syr Darya down to the Aral Sea basin. Apart from direct  
138 precipitation on the lake surface, most surface inflow is from several small mostly episodic  
139 rivers fed by rainfall, groundwater and snowmelt from neighbouring mountains (Huang et al.,  
140 2014). A few perennial rivers supply the lake from the west where they braid in a small  
141 alluvial plain at the shore. Today, however, most of the inflow is contributed by groundwater  
142 (Beloglasov and Smirnova, 1987). Average lake water temperature fluctuates between 16 °C  
143 (14.5 °C in summer 2007; Huang et al., 2014) and 0–2 °C in winter, with almost no stable  
144 thermal stratification throughout the year (Huang et al., 2014; Lauterbach et al., 2014). Earlier  
145 monitoring data reported average summer water temperatures of 12–13 °C (Beloglasov and  
146 Smirnova, 1987), with maxima of 20–23 °C during warmer years. Freezing starts in October  
147 and ice break-up may begin as early as March (up to late April) depending on short-term  
148 climatic variability (Huang et al., 2014).

149 The water column of Son Kol is well mixed and oxygenated to the lake bottom in the  
150 summer, though a small oxygen minimum has been noted at a water depth of 1.5 m revealing

151 respiration in the epilimnion (July 2007; Huang et al., 2014). Monitoring data from the 1960s  
152 and 1970s similarly reported a well-oxygenated water body throughout the whole year, but  
153 Beloglasova and Smirnova (1987) observed oxygen concentrations of 0.8–1.8 mg.l<sup>-1</sup> or less in  
154 the shallow water zone during winter. The strong fetch (maximum effective fetch of ~17 km;  
155 Håkanson et al., 1983) and the high wind frequency (wind speeds up to 20 km.h<sup>-1</sup> in summer  
156 and fall) impart efficient and complete mixing of the shallow water body. According to  
157 previous surveys the weather is calm only during 11% of the ice-free season (Beloglasov and  
158 Smirnova, 1987). In line with this, surface/intermediate flow currents reaching 2–8 cm.s<sup>-1</sup>  
159 across the water body (and up to 10 cm.s<sup>-1</sup> in the deepest part of the basin) reflect a highly  
160 dynamic lacustrine system (Beloglasov and Smirnova, 1987). Such conditions guarantee for a  
161 steady mixing of the whole water column and thus makes Son Kol mostly a polymictic lake  
162 nowadays (apart from under ice in winter). Son Kol is a moderate hard-water lake (4.56–4.96  
163 mg<sup>-1</sup>) with a mineralisation of 465 mg.l<sup>-1</sup> (summer) and 306 mg.l<sup>-1</sup> (winter) (Beloglasov and  
164 Smirnova, 1987). Accordingly, today it is a bicarbonate lake with weak magnesium  
165 dominance. Conductivity throughout the water column ranged between 515 and 530 mS.cm<sup>-1</sup>  
166 in the summer of 2012 (Schwarz et al., 2017).

167

## 168 2.2. Climatic setting

169 Local climatic conditions lead to short temperate summers and severe cold winters, with snow  
170 cover between November and April. Mean annual air temperature is -3.5 °C (mean of 11 °C  
171 in summer and -20 °C in winter), based on the nearest meteorological station (Dolon, 41.8° N,  
172 75.8° E, 3040 m a.s.l.) (Shnitnikov, 1980). Mean annual precipitation is 500–600 mm at Lake  
173 Son Kol, with a maximum in summer and a minimum in winter (Academy of Science of the  
174 **Kyrgyz** SSR, 1987). The current precipitation regime in the Tian Shan Mountains  
175 surrounding Son Kol is mainly regulated by the interaction between cyclonic activity driven



176 by the mid-troposphere Westerlies and the southwestern branch of the Siberian anticyclonic  
177 circulation (Aizen et al., 1997). Atmospheric dust concentrations (and loess) are maximal  
178 below 3000 m a.s.l in Central Asia, with thickest loess accumulations occurring between 1000  
179 and 1300 m a.s.l in the piedmont regions of the Tian Shan (Machalett et al., 2008). Different  
180 airflows govern convective surface winds in western Central Asia and thus, the regional  
181 distribution of remote dust particles over the year (Bugayev et al., 1957; Shettler et al., 2014).  
182 During the summer northwestern, western and northern air intrusions and northeast  
183 anticyclonic flows control up to 83% of the local wind directions. Dust monitoring near  
184 Bishkek reveals besides that wet deposition is a substantial process for aerosol deposition in  
185 mountain areas (Shettler et al., 2014). Nonetheless, backwards trajectories calculated from  
186 Son Kol document that remote air masses likely also come from southern (Afghanistan) and  
187 further western (Mediterranean) and northern (North Atlantic) sources (Shettler et al., 2014).

188

### 189 **3. Material and methods**

#### 190 *3.1. Lake coring and cross-core correlation*

191 Four sediment cores, including two sets of parallel cores, were extracted with a Uwitec piston  
192 corer in July 2007 from two different sites in the southeastern part of Lake Son Kol (Fig. 1).  
193 Core SK04 (133cm) was obtained at ca. 1.15 km from the southern shore (41°46'56.6"N,  
194 75°08'18.2"E, 10.5 m water depth) while a second set of parallel cores, SK06 and SK07 was  
195 retrieved at 1.6 km northeast of core SK04 (41°47'45.2"N, 75°08'22"E, 12.5m water depth)  
196 (Fig. 1). Although cores SK04 and SK07 have been extensively analyzed recently (Mathis et  
197 al., 2014; Huang et al., 2014; Pacton et al., 2015), no integrative study has been undertaken on  
198 core SK06 yet, although it represents the longest lacustrine archive ever retrieved in Lake Son  
199 Kol. In spite of the significant distance between the two coring sites, the cores exhibit very  
200 similar sedimentary successions (and lithological marker layers), thus allowing a robust core-

201 to-core correlation (Fig. 2a), as also demonstrated by the good match between magnetic  
202 susceptibility records for cores SK04, SK06 and SK07 (Fig. S1). Core SK06 was split in half  
203 lengthwise and photographed directly after core opening. A detailed macroscopic description  
204 established the different sedimentary facies (Units I-VI) as well as the stratigraphic  
205 correlations between cores SK04, SK06 and SK07 (Fig. 2a). No compaction and/or  
206 disturbance was observed in SK06.

207

### 208 *3.2. Age-depth model*

209 Owing to the robust core-to-core correlation between cores SK06, SK04 and SK07 (Figs. 2  
210 and S1) and SK06 and SONK\_11\_D1/2 from Lauterbach et al. (2014) (Fig. S2), the  
211 chronology of core SK06 (Fig. 2b) was established using available radiocarbon (AMS  $^{14}\text{C}$ )  
212 dating of both terrestrial plant macroremains and bulk sediment organic matter (OM),  
213 complemented by one short-lived radionuclide measurement ( $^{241}\text{Am}$ ). We selected and  
214 reappraised 8 bulk sediment samples and one terrestrial leaf from cores SK04 and SK07  
215 (Huang et al., 2014) along with one terrestrial plant remains from core SONK\_11\_D1/2  
216 (Lauterbach et al., 2014) (Table 1).  $^{14}\text{C}$  age anchor points were transferred to core SK06 using  
217 a detailed correlation of both lithological marker layers and magnetic susceptibility between  
218 core SK06 and SK04 and SK07 (Figs. 2a and S1) and between cores SK06 and  
219 SONK\_11\_D1/2 (Fig. S2). For the latter, the correlation at 33 cm in core SK06 of Poz-44073  
220 date (at 33 cm in SONK\_11\_D1/2; Lauterbach et al., 2014) is a robust anchor point for the  
221 core SK06 chronology, making the Poz-26363 radiocarbon age at 30 cm too old, and so this  
222 date was rejected from the age model. Note that not all the radiocarbon ages reported in  
223 Huang et al. (2014) were applied to core SK06 by cross-core correlation, since we only rely  
224 on age points that could be confidently transferred to core SK06. Poz-26363, Poz-26370 and  
225 Poz-26372 dates were rejected because they imply the reworking of older material in younger

226 sediments. Poz-37470 date (bulk sediment organic matter) was not used either because a  
227 terrestrial leaf dated in the same stratigraphic level was preferred (Poz-37469). Other  
228 discarded ages correspond to depths that could not been matched properly between cores  
229 SK04 and SK06 (Poz-26368).

230 A 150-yr reservoir age correction was further applied prior to calibration to each transferred  
231 conventional radiocarbon age by using the  $155 \pm 30$   $^{14}\text{C}$  yr BP reservoir age obtained on live  
232 deep-water algae by Lauterbach et al. (2014) (Table 1). Such a small reservoir age offset  
233 implies that the recent hardwater effect is small; this is further confirmed by nearly identical  
234 radiocarbon ages (within error bars) obtained on one bulk organic sample (not used here) and  
235 a terrestrial leaf from the same stratigraphical level in core SK04 (see Table 1 and Huang et  
236 al., 2014). More detail regarding reservoir effects in Lake Son Kol is provided in Lauterbach  
237 et al. (2014).

238 In order to constrain the age of the topmost sediments in core SK06, we used the close  
239 agreement of lithological markers in cores SK06 and SONK\_11D1/2 and assumed that the  
240 notable increase in  $^{241}\text{Am}$  activity at 5.5 cm in SONK\_11\_D1/2 occurs at 2 cm in core SK06  
241 (Fig. S2), and corresponded to the Northern Hemisphere nuclear weapons testing maximum  
242 in 1963 CE (Appleby et al., 1991). The topmost ca. 2 cm of SONK\_11\_D1/2 are missing in  
243 core SK06. Finally, the age-depth model for core SK06 was established using 10 corrected  
244 conventional radiocarbon ages (Table 1), as well as the  $^{241}\text{Am}$  anchor point at 2 cm sediment  
245 depth (most likely 1963 CE), as input parameters for a *P\_Sequence* deposition model  
246 implemented in Oxcal 4.4 (Ramsey, 2008) (Fig. 2b). The *k* parameter was set to 1 (as in  
247 Huang et al., 2014) and the radiocarbon ages were converted to calendar years by using the  
248 IntCal20 terrestrial calibration curve of Reimer et al. (2020). The high agreement index  $A_{\text{model}}$   
249 of 89.8% indicates that the chronology of core SK06 is reliable (Ramsey, 1995; 2001).

250

251 *3.3. Geochemistry*

252 Major elements in the sediments were measured by XRF at high-resolution (500- $\mu\text{m}$  step  
253 size) on the split-core surface with an Avaatech Core Scanner (EDYTEM Laboratory). Prior  
254 to measuring, the core surface was cleaned and foiled to prevent contamination and core  
255 desiccation. The X-ray beam was generated with a rhodium anode and a 125- $\mu\text{m}$  beryllium  
256 window, which allows a voltage range of 7–50 kV and a current range of 0–2 mA.  
257 Geochemical data were obtained with different settings according to the elements analyzed.  
258 They were adjusted at 10 kV and 1 mA for 10 s to detect Si, Ca, Al, Fe, Ti, K, Mn and S. For  
259 heavier elements (i.e., Sr, Rb, Zr, Br and Pb), measurements were performed at 30 kV and  
260 9.75 mA for 30 s. XRF data are displayed as element intensities in count per second (cps)  
261 (Fig. S3). Because no major element concentration was conducted on core SK06, we rely here  
262 on XRF core-scanner data available on core SK07 (unpublished data) in order to calibrate  
263 XRF measurements in core SK06. Major element concentrations on core SK07 were  
264 previously measured on 1.5-cm thick discrete samples using XRF at the University Lyon1.  
265 The sampling was conducted at a step of about 3–5 cm to allow for one calibration  
266 measurement every 3–5 core scanner measurement in core SK07.

267

268 *3.4. Hyperspectral imaging*

269 Hyperspectral imaging is a visible and near infra-red spectroscopic method measuring the  
270 reflectance of the sediment surface exposed to an incident light, and thus the sediment  
271 composition (Barranco et al., 1989; Balsam and Deaton, 1996; Balsam et al., 1997).  
272 Prior to analysis, the thin, superficial layer was scraped from the sediment to ensure a uniform  
273 surface. Reflectance spectroscopy was performed on a core logger bench equipped with two  
274 hyperspectral cameras (VNIR-PFD and SWIR, SPECIM®), which offers a short sampling  
275 time and allows for high spatial resolution (several dozens of  $\mu\text{m}$ ; Jacq et al., 2020). The

276 distance between the studied sample/core and the camera lens was 130 mm, with an  
277 acquisition angle of 90°. The surface of the sediment is indirectly illuminated by 18 halogen  
278 bulbs homogeneously distributed around the lens. The acquisition speed of the core logger  
279 was 0.5 mm.s<sup>-1</sup>. Pixels in the resulting Visible and Near Infrared (VNIR) image each cover an  
280 area of 47×47 μm<sup>2</sup> and contain raw spectral data from 400 to 1000 nm at 6.5 nm spectral  
281 resolution (given in digit number, DN), whereas the Short Wave-Infrared (SWIR) image  
282 pixels have a resolution of 150×150 μm<sup>2</sup> and cover a spectral range from 1000 nm to 2500 nm  
283 at 10 nm spectral resolution. Therefore, each pixel contains spectral information related to  
284 sediment composition. The depth of field of the imagers is ca. 2 mm. A normalization step is  
285 required to convert raw data (8 bits) to % reflectance. At each acquisition, a 100% reflective  
286 SPECTRALON standard was measured and used as a reference to express the data as %  
287 reflectance. Data processing was carried out with the ENVI (v. 5.3) and IDL 8.0 softwares.  
288 Pigment composition and concentrations were inferred from area indices (Fig. 3), which  
289 directly measure the surface area (here referred to as relative absorption band area or RABA;  
290 Trachsel et al., 2010) between the continuum and the reflectance spectrum at the absorption  
291 band targeted (for instance for the chlorophyll-a and its derivatives at 670 nm). The sum of  
292 the reflectance bands located throughout the absorption band is subtracted from the area under  
293 a line connecting the beginning and the end of the depression formed by the absorption band.  
294 In this study, area indices are followed by wavelengths indicating the width of the absorption  
295 band considered (e.g. RABA630-700). The main absorptions are determined by continuum  
296 removal, which also allows estimating the most interesting wavelength interval. In this study,  
297 we have defined four of them centered on 670 nm, 615 nm, 844 nm and 1670 nm (Fig. 3).

298

299 *3.5. Coprophilous fungal spore analysis*

300 Catchment erosion processes in high-altitude landscapes may be re-shaped by livestock  
301 grazing related to pastoral activity (Giguet-Covex et al., 2011). The variability of grazing  
302 pressure and associated soil erosion in a catchment area can be reconstructed from the  
303 sedimentary abundance of coprophilous fungal ascospores in lake sediments, as a proxy for  
304 the density of wild and domestic herbivores (Davies and Schaefer, 2006; Gauthier et al.,  
305 2010; Etienne et al., 2013). Son Kol catchment is still an area of transhumance today;  
306 therefore, domestic herbivores are assumed to be the primary source of coprophilous fungal  
307 spores in lake sediments, and wild herbivores are regarded to be a trivial source. To evaluate  
308 the impact of grazing activity on soil erosion in the catchment, abundances of coprophilous  
309 fungi ascospores were measured in 37 samples of core SK06 (for the last 6000 years). The  
310 resolution is approximately 100 years from 6000 to 5000 cal. yr BP, and about 250 years from  
311 5000 to 1500 cal. yr BP. Samples were chemically prepared according to the procedure of  
312 Faegri and Iversen (1989). *Lycopodium clavatum* tablets (averaging 20,848 spores per tablet)  
313 were added to each sample (Stockmarr, 1971) to express the results as influx or accumulation  
314 rates (number.cm<sup>-2</sup>.yr<sup>-1</sup>). Coprophilous ascospores were identified and counted using a  
315 photonic microscope following pictures and descriptions in van Geel and Andersen (1988),  
316 van Geel et al. (2003), Van Geel and Aptroot, 2006, Feeser and O'Connell (2009), Cugny et  
317 al. (2010) and Gelorini et al. (2012). Total coprophilous fungi *sensu stricto* include *Delitschia*  
318 (UG-1066), *Neurospora* sp. (HdV-1), *Podospora*-type (HdV-368), *Sordaria*-type (HdV-55A),  
319 *Sporormiella* (HdV-113), and *Trichodelitschia* (HdV-546).

320

### 321 3.6. Data analysis

322 Linear dimension reduction was performed by applying a correlation matrix-based principal  
323 component analysis (PCA) using the PAST v. 4.02 software (Hammer et al., 2001). The PCA  
324 was supplied with XRF (major elements) and hyperspectral data from core SK06 (Fig. 4).

325 Due to very high levels of non-normality and heteroscedasticity of the original descriptors,  
326 data were first log-transformed and then separately centered on a per-sample basis in order to  
327 remove overall-count effects and to focus on relative variations among samples. Also using  
328 the PAST software, time series corresponding to the resulting PC1 and PC2 were further  
329 investigated using Gallagher et al.'s (2011) Bayesian "change-point modeling" algorithm for  
330 identifying abrupt regime shifts in time series, often referred as tipping points in geosystems  
331 (Lenton et al., 2008; Lenton, 2011; Scheffer et al., 2012). Number of MCMC iterations  
332 required by Gallagher et al.'s method was set to 1,000,000 with a burn-in phase of 20,000  
333 iterations; up to 300 changepoints were allowed a priori in order to generate fully-  
334 unconstrained changepoint models, resulting in an average number of changepoints of 74 and  
335 61 for the PC1 and PC2 time series, respectively.

336

## 337 **4. Results and interpretation**

### 338 *4.1. Sedimentology*

339 Sediments mostly consist of alternating yellowish and light-brownish calcareous muds in the  
340 upper 1.22 m of core SK06, whereas dark organic-rich muds alternating with yellow-brown  
341 calcareous muds and white carbonate layers predominate between 1.22 and 1.91 m (Fig. 2).  
342 The transition between the basal organic-rich sediments and the upper 1.22 m alternating  
343 calcareous muds is characterized by homogeneous greyish calcareous clays between 0.94 and  
344 1.22 m. Apart from this transition, the sediments are finely laminated in core SK06, and  
345 especially in dark organic sediments. XRF core scanning data allows deciphering the distinct  
346 geochemical composition of the three main lithologies (Fig. S3). Dark organic sediments are  
347 typified by their high abundances in Br and S (Figs. S3 and 5), whereas the intervening  
348 carbonate layers exhibit an enrichment in Ca and Sr (Fig. S3). In contrast, the transition  
349 interval (greyish homogenous clays) is best described by high abundances of Ti, Rb, Zr, Fe

350 and Si, emphasizing their high mineral content. The upper alternations between yellowish and  
351 light-brownish sediments are characterized by a mixed geochemical composition and exhibit  
352 fluctuating values of Ti, Rb and Zr, but also highly variable abundances of Ca and Sr (Fig.  
353 S3).

354

#### 355 *4.2. VNIR-SWIR proxy data*

356 Chlorophyll *a* (Chl *a*) and its derivative products (Rein and Sirocko, 2002; Das et al., 2005;  
357 Wolfe et al., 2006; Michelutti et al., 2010) was measured as the area index between 630 and  
358 700 nm (RABA630–700) normalized by the absorption maximum at 670 nm ( $R_{670}$ ) (Fig. 3) as  
359 defined by Wolfe et al. (2006). Although Chl *b* may contribute in the absorption range  
360 between 630 and 700 nm (i.e., at 652 nm, Bidigare et al., 1990; Yacobi et al., 2015), it is not  
361 observed in our spectral data (Fig. 3). We therefore assume that the measured absorption peak  
362 of particles at 670 nm almost exclusively represents the optical activity of Chl *a*, apart from  
363 the slight interference by the accessory chlorophylls Chl *b* and Chl *c*.

364 Phycocyanin is a water-soluble accessory pigment to chlorophyll exhibiting a notable  
365 absorbance trough around 610–620 nm for C-phycocyanin (Glazer et al., 1973; Yacobi et al.,  
366 2015). For this study, we propose to screen the phycocyanin content by measuring the area  
367 index between 600 and 630 nm (RABA600–630) normalized by the absorption peak at 615  
368 nm ( $R_{615}$ ) as described by Yacobi et al. (2015) (Fig. 3). It must be recalled, however, that the  
369 absorption of Chl *a* and accessory chlorophylls Chl *b* and Chl *c* partly overlap the range of  
370 optical activity of phycocyanins, although weakly (Yacobi et al., 2015). However, the  
371 RABA630-700 and RABA600-630 curves exhibit different patterns (Fig. 5), implying that a  
372 genuine signal does exist apart from the Chl *a* plus derivatives. This also suggests that  
373 phycocyanin predominates in some intervals where Chl *a* plus derivatives do not (Fig. 5), thus  
374 allowing their respective trends to be distinguished and analyzed separately in core SK06.



375 Phycocyanins are common in cyanobacteria (Williams et al., 1980), and **their** concentration in  
376 shallow lacustrine systems is used to estimate the presence and relative abundance of  
377 cyanobacteria (Horváth et al., 2013). Phycocyanin has been reported in lacustrine sediments  
378 of Lake Qinghai, Central China (Ji et al., 2009) and used to indicate cyanobacterial blooms in  
379 the Baltic Sea (Woźniak et al., 2016). Chl *a* plus derivatives and phycocyanin contents are  
380 thus interpreted here as a proxy for oxygenic phototrophs, which usually indicate aerobic  
381 conditions. Total Bacteriopheophytin *a* (Bphe *a*) is a stable degradation product of  
382 Bacteriochlorophyll *a* or Bchl *a* with a characteristic spectral absorbance in the infrared range  
383 around 750–770 nm in ethanol extracts (Coolen and Overmann, 1998; Rogozin et al., 2012)  
384 and 800–900 nm in sediments (Scheer, 2006; Ji et al., 2009; Blankenship, 2013). Bphe *a* was  
385 determined from the area index between 750 and 900 nm (RABA750–900) normalized by the  
386 absorption maximum at 844 nm ( $R_{844}$ ) as defined by Butz et al. (2015) (Fig. 3). Because no  
387 other known bacterial pigment absorbs at this wavelength, the relative abundance of Bchl *a*  
388 (and its derivative by-products Bphe *a*) can be evaluated by measuring the Bphe *a* absorption  
389 trough between 750 and 900 nm normalized by the absorption peak at  $R_{844}$ . Although  
390 Bacteriopheophytin *b* (Bphe *b*) may also contribute at 820 nm (Oren, 2011), no absorption  
391 peak is observed at that wavelength in our data (Fig. 3). We therefore assume that the  
392 measured absorption peak of particles at 844 nm represents exclusively the optical activity of  
393 Bphe *a*. Bphe *a* is produced by planktonic anoxygenic phototrophic bacteria (APBs) such as  
394 purple sulphur bacteria (Chromatiales, Rhodospirillales; Schmidt et al., 2002; Rogozin et al.,  
395 2012) which commonly live in the chemocline of stratified meromictic lakes, where reduced  
396 sulphur species are **very abundant** and oxygen levels are low, if sufficient light is available for  
397 photosynthesis (Butz et al., 2016). Hence Bphe *a* is a diagnostic pigment for photosynthesis  
398 in anaerobic conditions and the presence of that pigment in the sediment may help infer long-  
399 term lake stratification, and the presence of a chemocline (Fig. 5). Aromatic organic matter

400 content (ARO) has a characteristic absorption near 1670 nm (Wheeler, 1959) and it seems  
401 that no other compound absorbs at this wavelength, so we assume that an index in this  
402 spectral range can infer this property (Fig. 5). Nevertheless, we are aware that SWIR  
403 measurements may be influenced by sediment moistness and grain size, for which no known  
404 correction can be applied yet. For this study we propose the area index between 1660 and  
405 1690 nm (RABA1660–1690), normalized by the absorption maximum at 1670 nm ( $R_{1670}$ ).  
406 ARO values are interpreted as a proxy for tracking the inputs of terrestrial organic matter  
407 from the shore (e.g., soils) and related catchment, with high (low) ARO corresponding to  
408 increased (decreased) inputs of detrital organic matter.

409

#### 410 *4.3. PCA analysis*

411 XRF and hyperspectral (VNIR, SWIR) data, with a temporal resolution ranging between 3  
412 yr.mm<sup>-1</sup> and 7–8 yr.mm<sup>-1</sup>, were jointly analyzed by PCA in order to characterize sediment  
413 end-members and detect abrupt shifts in the multivariate time series (Fig. 4a). The XRF core-  
414 scanner of Ca, Fe and Ti was previously calibrated using XRF conducted on discrete samples  
415 in core SK07 (28 samples), exhibiting an excellent match between the two methods (Fig. 4b).  
416 PC1, PC2 and PC3 together explain about half of the total variance (Figs. 4a and S4),  
417 allowing for the identification of different end-members in the lake system: (i) high Ti, Rb, Fe  
418 and RABA1660–1690 values for PC1 > 0 (which are related to organic and minerogenic  
419 inputs from the catchment), while Sr, Ca (which indicate endogenic carbonate precipitation)  
420 and to a lesser extent Zr (coarse detrital input) drive PC1 < 0, making PC1 a proxy for  
421 allochthonous vs autochthonous products in the lake system; (ii) high Mn, RABA600–630  
422 and RABA630–700 values for PC2 > 0 (which are related to oxic conditions in the water  
423 column) vs. high Br, S (and to a lesser extent Zr) and RABA750–900 values (which are  
424 linked to organic matter content and infer reducing and sulphidic conditions in bottom waters)

425 for PC2 < 0, making PC2 a proxy for oxygenated vs. euxinic bottom water conditions in Lake  
426 Son Kol. Based on this result, prominent environmental change occurred around 8400, 8200–  
427 7800, 7700–7500, 7300–7000, 6500–6100, 6000–5700 and 5500–5250 cal. yr BP (Fig. S5),  
428 when increased influx of detritus (PC1 > 0) is in close correspondence with the development  
429 of euxinic conditions in the hypolimnion (and thus of a chemocline; PC2 < 0). The Si content  
430 appears to slightly differ from other terrigenous (river-borne) elements (PC1 and also PC3  
431 [14.4%], whose palaeoenvironmental signature is more difficult to interpret and is not  
432 discussed here) (Fig. S4). Si-value is high in the silt fraction and is, in addition to its weak  
433 riverine contribution, mostly linked with aeolian inputs into the lake (Huang et al., 2014).  
434 Judging from the lack of any conspicuous link to the other elements (Fig. 4a) and given the  
435 biochemical reactivity of this element, Mn is used to decipher bottom-water redox processes,  
436 as already reported in other lacustrine settings (Wirth et al., 2013; Davies et al., 2015).

437

#### 438 *4.4. Time series of hyperspectral and geochemical data*

439 Between 8500 and 7000 cal. yr BP (191–164 cm), phycocyanin (from cyanobacteria) has the  
440 highest abundances downcore, probably related to the development of benthic cyanobacterial  
441 mats in a shallow water body (Fig. 5). Chl *a* plus derivatives and Bphe *a* also oscillate at very  
442 **high** frequency. Bphe *a* and ARO maxima occur repeatedly around 8500, 8400, 8200–7800,  
443 7700–7500 and 7300–7000 cal. yr BP, corresponding to dark organic-rich sediments along  
444 with high Br and S content. No match was found, however, between phycocyanin and Chl *a*.  
445 Subsequently, Chl *a* plus derivatives and phycocyanin exhibit a pronounced decline between  
446 7000 and 6500 cal. yr BP. Phycocyanin abundances drop from 7000 cal. yr BP onwards and  
447 never recover to former levels. Bphe *a* and ARO values also exhibit a drastic decrease in this  
448 time interval, as also do Br and S content.

449 From 6500–5250 cal. yr BP, Chl *a* plus derivatives values increase and reach their highest  
450 abundances downcore, with however considerable fluctuations. Bphe *a* content recovers to  
451 high levels and even dominates between 5750 and 5250 cal. yr BP, but dramatically decreases  
452 and readily disappears from 5250 cal. yr BP onwards (Fig. 5). Coevally, Br, S and ARO  
453 values are also high between 5750 and 5250 cal. yr BP (corresponding to the deposition of  
454 dark organic sediments), but rapidly drop after 5250 cal. yr BP. Chl *a* plus derivatives and  
455 phycocyanin exhibit a prominent decline in their abundances between 5000 and 3850 cal. yr  
456 BP. Judging from XRF core-scanner data, enhanced detrital inputs occurred in this interval  
457 (Fig. S3). Subsequently, Chl *a* plus derivatives and phycocyanin increase from 3850–3100  
458 cal. yr BP, whereas Bphe *a* still oscillates at very low levels. Higher Br, along with enhanced  
459 ARO, also parallels the occurrence of dark organic-rich sediments in this interval. Over the  
460 course of the last 3000 years, oxygenic phototrophs fluctuate at lower levels, whereas Bphe *a*,  
461 Br and ARO remain low (Fig. 5).

462

#### 463 *4.5. Abrupt regime shifts*

464 Abrupt regime shifts were identified based on slope variations of the average changepoint  
465 model calculated separately for PC1 and PC2. For each principal component, average  
466 changepoint model's slope (absolute value) is plotted against time, with 99% and 95%  
467 percentiles taken as cut-off values (Figs. 6 and 7). Time intervals with average changepoint  
468 model's slope values above these cut-offs indicate abrupt changes in the time series, and can  
469 therefore be regarded as tipping points. For PC1, tipping points are recognized as 10–20 yr  
470 intervals (beyond the 99% percentile value) at ca. 3670, 5050–5080 (highest slope value),  
471 5250, 5370, 5850, 6150, 6220, 6900, 7030, 7980, 8230, 8280 and 8340 cal. yr BP (Fig. 6).  
472 For PC2, tipping points are similarly identified as 10–20 yr intervals (beyond the 99%  
473 percentile value) around 5250 (highest slope value), 5950, 6040, 6230, 6280, 6380, 6670,

474 6700, 6900, 7060, 7750, 7800, 8320 and 8380 cal. yr BP (Fig. 7). Hence the time intervals  
475 around 5050–5250, 6220–6240 and 8200–8400 cal. yr BP are detected for both PC1 and PC2,  
476 and probably correspond to abrupt regime shifts in the lacustrine system. Interestingly, low  
477 PC2 values match with a strong peak of magnetic susceptibility (Figs. S1 and S5) around  
478 8200–8400 cal. yr BP, suggesting the occurrence of abrupt changes in the lake and  
479 surrounding catchment that are tied with the 8200 cal. yr BP cold event widely recognized in  
480 the Northern Hemisphere (Alley et al., 1997; Rohling and Pälike, 2005).

481

## 482 **5. Discussion**

483 Between ca. 8500 and 5250 cal. yr BP anaerobic conditions (and the penetration of light deep  
484 in the lake to support photosynthesis for APB) are inferred from high Bphe *a* and Br (PC2 <  
485 0), along with the occurrence of dark organic layers (Fig. 5). Coevally, high ARO and  
486 minerogenics content (PC1 > 0) suggest enhanced inputs of detrital organic matter and a link  
487 between terrestrial influx, sedimentary dynamics and the advent of hypolimnetic anoxia in the  
488 earlier part of the record. After 5250 cal. yr BP, the near absence of Bphe *a*, lower ARO  
489 values but persistently high abundances of oxygenic phototrophs suggest prolonged aerobic  
490 conditions within a well-mixed water body (Fig. 5).

491

### 492 *5.1 Episodes of increased snowmelt and indicators of lake-level changes*

493 Grain size and geochemical/isotopic (TC, TN, TOC, C/N,  $\delta^{13}\text{C}_{\text{org}}$  and  $\delta^{18}\text{O}$ ) data can be useful  
494 in supporting interpretations from hyperspectral and XRF data. Here we compare previously  
495 unpublished and available data from SK07 (Huang et al., 2014) to help interpret the 8500–  
496 5000 cal. yr BP time interval during which dark organic intervals are recurrently interbedded  
497 with yellow-brown calcareous muds (Fig. 7). Altogether, these parameters help to distinguish

498 changes in onshore and in-lake environmental conditions such as sheet-wash from the shore,  
499 productivity and nutrient availability, and the possible mechanisms at work.

500 Generally, the timing of changes in geochemical and isotopic proxies in core SK07 agree with  
501 those of XRF and hyperspectral data from core SK06 (Fig. 8). Dark organic intervals with Br  
502 and ARO in core SK06 have elevated TOC, TC, TN and C/N values but strongly depleted  
503  $\delta^{13}\text{C}_{\text{org}}$  and  $\delta^{18}\text{O}$  values in core SK07. Such prominent  $\delta^{13}\text{C}_{\text{org}}$  depletions might indicate the  
504 predominance of APB during phases of lake stratification because pronounced carbon isotope  
505 fractionation towards lighter  $\delta^{13}\text{C}$  has been associated with purple sulphur bacteria in  
506 meromictic lakes (Zyakun et al., 2009). Alternatively, short-lasting depletions in  $\delta^{18}\text{O}$  can  
507 indicate seasonal shifts in the hydrological regime; i.e., increased winter precipitation and/or  
508 increased meltwater discharges linked to erosive runoff during spring snow thaw, but also  
509 increased temperatures (Huang et al., 2014; Lauterbach et al., 2014) that would have fueled  
510 meltwater runoff. The dark organic-rich and  $\delta^{18}\text{O}$ -depleted intervals co-occur with coarser  
511 (e.g., silt- and sand-sized) minerogenic and organic (e.g. phytoclasts) inputs into the lake  
512 (Figs. 8 and S6), whereas the intervening yellow-brown calcareous muds are enriched in finer  
513 (clay-sized) detritus with low sand content (Fig. 8). Erosive runoff after thawing during spring  
514 (Huang et al., 2014), and/or higher winter snowfall (Lauterbach et al., 2014), can account for  
515 coarse detritus discharge into the lake. Interestingly, the PCA analysis reveals that higher  
516 detrital inputs ( $\text{PC1} > 0$ ) are not associated with elevated Zr content in core SK06 (Fig. 4);  
517 instead, dark organic layers exhibit low Zr content (Figs. 8 and S3). Zr is usually more  
518 concentrated in sand-size lacustrine sediments (e.g., Cuven et al., 2010; Davies et al., 2015)  
519 and so, this Zr depletion in a few coarse dark organic intervals is puzzling, since otherwise all  
520 detritics (Ti, Rb and Zr) exhibit similar trends in the intervening yellow-brown calcareous  
521 muds (Fig. S3). Therefore, enhanced levels of Ti and Rb along with lower Zr (thus  $\text{PC1} > 0$ )  
522 in the Br-enriched ( $\text{PC2} < 0$ ) dark organic layers either suggest the dilution of Zr input by Ti-

523 and Rb-laden terrestrial organic matter delivered during snow thaw (as inferred from high  
524 ARO values), and/or the dilution of Zr (associated with the coarse detrital fraction) during  
525 relatively high lake water level phases. Zr mainly originates from erosion processes in the  
526 upper catchment of Lake Son Kol because Zr is abundant in local Paleozoic intrusives  
527 (granites and granodiorites), but these tend to be covered by Quaternary sediments and soils  
528 (colluvial and glacifluvium deposits) on the shoreline belt and further up-valley. Although Zr  
529 is linked to Br and S on PC2, Zr is also significantly associated with elevated Sr and Ca (PC1  
530  $< 0$ ; Fig. 4), suggesting that Zr is enhanced during low lake water levels, which promotes  
531 endogenic carbonate precipitation (Huang et al., 2014). Hence higher Zr infers the increasing  
532 influence of riverine discharges at relatively low lake levels and the reduced distance between  
533 the coring site and proximal detrital sources. Dilution of Zr most probably occurs during  
534 short-lived increases in lake levels, which in turn enhance the formation of organic-rich  
535 deposits in Lake Son Kol when elevated terrestrial inputs from snowmelt strengthened the  
536 export of dissolved organic matter into the lake. The most probable sources of material during  
537 snowmelt discharges (e.g., floods) are the abundant fan delta/fluviial material at the southern  
538 edge of the lake, and/or superficial sediments on the slopes (e.g., colluvium) directly adjacent  
539 to Lake Son Kol. In contrast, higher Ti, Rb and Zr occur coevally during lower lake level  
540 phases, which correspond to the deposition of grey calcareous muds enriched in fine  
541 siliciclastics (Fig. S3). Hence higher Zr, Sr and Ca values indicate lower lake water levels and  
542 coincide with the deposition of yellow-brown calcareous muds. In contrast, dark organic  
543 deposits enriched in Br, S, C/N and TOC are associated with the injection of large volumes of  
544 sediment-laden sheet-wash from the shores (high ARO values) during intensified winter  
545 precipitation and/or stronger snowmelt (Huang et al., 2014; Lauterbach et al., 2014), which  
546 are likely to occur under warmer climate conditions.

547

548 *5.2. Redox conditions and oxygenation of the hypolimnion*

549 The geochemical behavior of Mn-enriched phases can be used to infer past redox conditions  
550 and thus help diagnose oxic and sulphidic redox regimes in the water column (Fig. 9), as  
551 reported from Lake Cadagno (Wirth et al., 2013). Based on elevated Mn content, oxic  
552 conditions ( $PC2 > 0$ ) prevailed around 8300–8200, 7400–7250, 5500–5400 and in particular  
553 between 5000 and 3800 cal. yr BP in Son Kol when Mn was flushed to the lake with detrital  
554 input, and probably occurred in the water column as dissolved unstable  $Mn^{2+}$  or as oxidized,  
555 particulate Mn-(oxyhydr)oxides. However, no Mn-(oxyhydr)oxides were found in SK06,  
556 possibly due to a reduction of  $Mn^{2+}$  (leading to a further recycling of Mn into the water  
557 column) if the sediment-interface and/or the sediments were reducing. Oxic water column  
558 conditions were accompanied by the precipitation of various carbonate minerals judging from  
559 the regular occurrences of carbonate layers (Pacton et al., 2015), as well as high XRF-Ca  
560 content in the yellow-brown calcareous muds separating dark organic-rich sediments, thus  
561 reflecting the high alkalinity of lake waters during those time intervals. The regular  
562 occurrence of sediment-dwelling ostracods and molluscs (which require well-oxygenated  
563 bottom waters) further supports that a well-mixed (polymictic) water body, with oxic bottom  
564 waters, has been established during the last 5000 years (Huang et al., 2014), in accordance  
565 with recent monitoring data (Lauterbach et al., 2014).

566 In contrast, there is compelling evidence that the hypolimnion recurrently turned to a  
567 sulphidic (euxinic) redox regime between 8500 and 5250 cal. yr BP, involving a stabilization  
568 of the water-column stratification and the full development of a chemocline. Low sedimentary  
569 Mn concentrations (Fig. 9), the absence of carbonate phases and the drastic increases in Br  
570 and S content ( $PC2 < 0$ ) characterize several phases of euxinic bottom waters between 8500  
571 and 5250 cal. yr BP. Coevally, high Bphe *a* values (inferring that purple sulphur bacteria  
572 thrived in the water column), and the rising TOC content in dark organic-rich sediments (Fig.



573 8), are good indicators of reducing bottom water conditions. Although such a hypothesis was  
574 ruled out by Lauterbach et al. (2014) based on the absence of varves and/or vivianite crystals  
575 (the latter indicating anoxic conditions in a non-sulphidic environment during sediment  
576 deposition, which is obviously not the case here) in dark organic-rich sediments, several lines  
577 of evidence advocate for the recurrence of hypolimnetic euxinia (and thus the development of  
578 a chemocline) in Lake Son Kol before 5250 cal. yr BP. Prolonged lake stratification and  
579 reducing oxygen conditions around 8500, 8400, 8200–7800, 7700–7500, 7300–7000, 6500–  
580 6100, 6000–5700 and 5500–5250 cal. yr BP (Fig. 9) are inferred from (1) high TOC contents  
581 (up to 16%) in dark organic-rich layers (Fig. 8) that imply both reducing bottom water  
582 conditions which prevents organic matter degradation and especially of photosynthetic  
583 pigments (Figs. 7 and 9); (2) the unusual preservation of numerous colonies of coccoid  
584 prokaryotes altogether with the preserved intracellular content of cyanobacteria that requires  
585 anoxic conditions (Pacton et al., 2015) and in particular, (3) the regular occurrence of Bphe *a*  
586 pigments that indicate photosynthesis in anaerobic conditions and thus hypolimnetic anoxia  
587 between 8500 and 5250 cal. yr BP.

588  
589

### 590 *5.3. Origin and driving mechanisms of detrital-rich organic layers in Lake Son Kol*

591 As developed previously, the onset of euxinic bottom water conditions in Lake Son Kol  
592 between 8500 and 5250 cal. yr BP most likely occurred under warmer climate conditions that  
593 fostered meltwater runoff. Climate warming possibly fostered permafrost melting, which  
594 activated the subaquatic springs that in turn fed the lake with solute-rich waters. Such  
595 conditions are highly favorable for the sustainment of stable lake stratification and the  
596 development of a chemocline, as observed in other high-altitude lake settings (Wirth et al.,  
597 2013). However, this seems to be inconsistent with the regular occurrence of coarse material  
598 in dark organic-rich sediments, which implies the transport of organic and minerogenic

599 detritus over long distances in Lake Son Kol. In meromictic lakes, the deposition of coarse  
600 detritics in a coring site today located more than 3 km from the shore could be attributed to  
601 hyperpycnal or mesopycnal currents (Mulder and Chapron, 2011), depending on the density  
602 contrast between inflowing and lake waters. Hypopycnal flows are not considered here  
603 because hypopycnal flows in lakes generally supply small quantities of sediments and would  
604 require further reconcentration processes in the lake (over a 3-km distance in the case of Lake  
605 Son Kol). Aeolian deposition of coarse-grained detritics (including large phytoclasts, Fig. S6)  
606 is unlikely as we would expect much finer grain sizes in the case of loess deposits (i.e., silts)  
607 (Huang et al., 2014), and loess deposits are generally restricted to lower altitudes in the  
608 piedmont of the Tian Shan Mountains (Machalett et al., 2006; Lauterbach et al., 2014).  
609 Finally, increased coastal wave erosion driven by winds can be rejected as well because  
610 higher wave activity would also have broken down the chemocline. Moreover, hyperpycnal  
611 deposits are often typified by sharp-based lag rich in organic debris (Mulder and Chapron,  
612 2011), which is never encountered in core SK06 where coarse detritics are rather  
613 homogenously distributed in the organic-rich layers (apart from the occurrence of a few  
614 normal graded beds; Fig. S6). Hence, the most reasonable candidates for the deposition of  
615 coarse minerogenic and organic detritus at the coring site are homopycnal or mesopycnal  
616 flows. Homopycnal flows result from a small density contrast between the entering flow and  
617 the surrounding water (Bates, 1953) and can export suspended sediment load either in  
618 stratified or non-stratified lake basins (Mulder and Chapron, 2011). Alternatively, mesopycnal  
619 flows are partially buoyant mesopycnal flows (interflows) of turbid plumes forming along  
620 each pycnocline (that separates each water mass) (Van Rensbergen et al., 1998, 1999;  
621 Schröder et al., 1998; Chapron et al., 2004, 2005; Støren et al., 2010; Mulder and Chapron,  
622 2011).

623

624 Formation of mesopycnal currents is promoted when the inflowing waters enter more saline  
625 lake waters, as occurred between (at least) 6000 and 5000 cal. yr BP when higher  
626 conductivities (very low lake levels) were inferred in Lake Son Kol (Schwarz et al., 2017).  
627 Modern conductivity of Lake Son Kol (515–530 mS.cm<sup>-1</sup>) is about twice that of meromictic  
628 Lake Bourget in France (300–350 mS.cm<sup>-1</sup>). Therefore, mesopycnal (or homopycnal) flow  
629 deposits occurred during periods of strong snowmelt (and/or increased temperatures),  
630 indicated by the strong depletion of  $\delta^{18}\text{O}$  in dark organic layers (Fig. 8), thus implying the  
631 remobilization of sediments from the lake catchment whenever sufficient stream discharge  
632 occurred to form a turbid plume. Two main factors control floods triggered by snowmelt: the  
633 snowmelt potential and the actual melting. Snowmelt depends on the amount of seasonal solid  
634 winter precipitation (e.g., snow) in the catchment while the intensity of melting is mostly  
635 controlled by spring/summer temperature (Støren et al., 2010). During the warm Mid-  
636 Holocene in Central Asia (Chen et al., 2008), periods of higher solid winter precipitation  
637 coupled with relatively cool springs and warm summers may have favored increased  
638 availability of snow when early summer temperatures started to rise in the mountains,  
639 intensifying spring melt floods and the amount of terrestrial material washed into the lake.  
640 Convective summer rainstorms during warm summers could also lead to lower lake water  
641  $\delta^{18}\text{O}$  values during the Holocene thermal maximum. Intensive rainstorms in the summer  
642 would increase hillslope erosion and stream discharge at a time when the lake was probably  
643 more saline, stratified and shallow due to more intensive evaporation. Increased intensities of  
644 rainstorms and snowmelt are therefore both possible drivers of mesopycnal or homopycnal  
645 flows (and coarse sediment deposition) in Lake Son Kol. Alternatively, hypolimnetic anoxia  
646 could have developed seasonally, by occurring under ice during the winter when organic  
647 matter decomposes (if there is a limited snow cover that enables photosynthesis) and breaking  
648 down towards the end of summer. However, this hypothesis is unlikely as (1) very few graded

649 beds related to ice melt are observed (i.e., resulting from the settling of fine-grained aeolian  
650 particles deposited on ice to the lake bottom), (2) the regular occurrence of large phytoclasts  
651 in dark organic layers with coarse minerogenic detritus (Fig. S6) implies a soil erosion from  
652 the catchment, and thus is not compatible with aeolian depositional processes and, (3) Bphe *a*  
653 data exhibit multi-annual periods of stratification (i.e., meromixis), rather than a seasonal  
654 signal (Fig. 10).

655

#### 656 *5.4. Holocene chemocline development*

657 The prominent changes in oxygenation conditions in the lake water column led to distinctive  
658 changes in Lake Son Kol autotrophs (Figs. 9 and 10). Bphe *a* is produced by anoxygenic  
659 phototrophic bacteria (APBs) which commonly live in the chemocline (oxic/anoxic transition)  
660 of meromictic lakes (Overmann et al., 1991; Rogozin et al., 2012; Blankenship, 2013; Butz et  
661 al., 2016). APBs require nutrients, sulphides and light to flourish, as provided by stable  
662 chemocline (Parkin and Brock, 1980; Van Gemerden and Mas, 1995, Hodgson et al., 1998;  
663 Madigan, 2003), and thus APBs may form dense populations in meromictic lakes (Van  
664 Gemerden and Mas, 1995; McGowan et al, 2008). Such conditions usually develop when  
665 lakes permanently stratify, but oxygen drawdown **could** also develop under ice with  
666 respiration of organic matter, potentially leading to anoxic conditions during certain times of  
667 the year when light is available and APB could flourish. Since dark organic-rich sediments  
668 represent more than a single year in duration (Figs. 5 and 9), Bphe *a* suggest that the stability  
669 of stratification increased and that there may have been periods of prolonged hypolimnetic  
670 anoxia in Lake Kol during the Mid-Holocene. Recurring Bphe *a*, Br, S and ARO positive  
671 anomalies (i.e., PC2 < 0) at 8500, 8400, 8200–7800, 7700–7500, 7300–7000, 6500–6100,  
672 6000–5700 and 5500–5250 cal. yr BP thus imply periods characterized by the development of  
673 a chemocline, which are tied to increased terrestrial inputs into the lake (PC1 > 0) (Fig. 9). A

674 very similar assemblage of spectral signatures (i.e., Bphe *a* and phycocyanin) was reported  
675 from Lake Qinghai (China), where higher Bphe *a* concentrations are tied to elevated  
676 freshwater inputs and increased warmth between ca. 8200 and 5500 cal. yr BP (Ji et al., 2009)  
677 (Fig. 10). This also coincides well with higher (alkenone-inferred) summer temperatures in  
678 Lake Balikun and Lake Qinghai at 8500–5000 cal. yr BP (Hou et al., 2016; Zhai et al., 2017).  
679 Warmer summers in Lake Jaczno (Poland) were also linked to the onset of lake meromixis  
680 (Butz et al., 2016). Besides, vegetation expansion in lake catchments during warm periods  
681 (and more specifically the expansion of **broad-leaved** trees, such as birch) is known to  
682 increase dissolved organic matter (DOM) fluxes to lakes and **decrease** UV light penetration  
683 (Sommaruga and Psenner, 1997; Schmidt et al., 2002). At Lake Son Kol, the presence of trees  
684 in the catchment and around the lake probably influenced lake trophic status and mixing  
685 conditions by (1) stimulating pedogenesis, which fueled oxygen consumption rates into the  
686 water body due to enhanced organic compounds exported to the lake and (2) sheltering the  
687 lake surface from wind-induced mixing (i.e., reduced wind exposure) thus favoring a  
688 prolonged time of stratification. Interestingly, the initiation of euxinic conditions in the  
689 hypolimnion (coeval to the deposition of dark organic-rich layers) followed the mass  
690 expansion of birch and arboreal **taxa** in the catchment between 8500 and 5000 cal. yr BP (Fig.  
691 10), which enhanced terrestrial inputs of DOM and thus organic matter decomposition in the  
692 lake. Therefore pollen, geochemical and sedimentological proxies all suggest that  
693 hypolimnetic euxinia developed during warmer summers, and probably milder  
694 winters/springs. As proposed by Mathis et al. (2014), centennial-scale episodes of weakened  
695 seasonality indeed occurred between 8500 and 5000 cal. yr BP, alongside increased winter  
696 temperatures and warm summers hence promoting a longer growing season for the trees. Our  
697 findings also agree with other speleothem records from Central Asia which indicate increased  
698 moisture transport by the Westerlies, and higher winter/spring precipitation in the early part of

699 the Holocene (Fig. 10; Chen et al., 2016; Cai et al., 2017; Liu et al., 2019). In contrast, the  
700 breakdown of lake stratification at 7500–7300 and 7000–6500 cal. yr BP probably coincides  
701 with cooler climate conditions (and longer and thicker ice cover) linked to the strengthening  
702 and southward position of the Siberian High in arid Central Asia. Increased influence of the  
703 Siberian High would weaken the mid-latitude Westerlies and/or push their tracks further to  
704 the south, resulting in comparably drier conditions in the Tian Shan (Aichner et al., 2015;  
705 Wolff et al., 2017).

706

707 Enhanced nutrient concentrations (linked with higher soil development DOM supply to the  
708 lake) are associated with a decrease in lake water transparency between 8500 and 7000 cal. yr  
709 BP, judging from unstable values of oxygen phototrophs (Fig. 9). At roughly the same time,  
710 phycocyanin abundances rapidly oscillate at high levels, suggesting that cyanobacteria  
711 predominated over Chl *a* and derivatives. A similar pattern was reported from shallow lakes  
712 in southern Sweden (Randsalu-Wendrup et al., 2014) and northern Germany (Dressler et al.,  
713 2007). Therefore, cyanobacteria, which have the ability to sustain relatively high growth rates  
714 at low light intensities (Mur et al., 1999), likely overtook oxygen phototrophs between 8500  
715 and 7000 cal. yr BP, concurrently with the occurrence of APB blooms at the chemocline (Fig.  
716 9). Before 5000 cal. yr BP, Lake Son Kol was a shallow and closed lacustrine system (Fig. 9;  
717 Schwarz et al., 2017) and it is likely that a substantial proportion of the oxygen phototrophs  
718 could derive from benthic algae (e.g., benthic diatoms and cyanobacteria). This also agrees  
719 well with the occurrence of biofilms and/or microbial mat structures between 8500 and 5000  
720 cal. yr BP (Pacton et al., 2015).

721

722 A major and abrupt change in lake regime is recorded around 5250 cal. yr BP in Lake Son  
723 Kol, with a drastic decline of Bphe *a* concentrations that occurred within a decade or so (Figs.

724 9, 10). The abrupt decline in APB pigments indicates a rapid collapse of stratified conditions  
725 and the breakdown of anoxic conditions in the hypolimnion after 5250 cal. yr BP, although  
726 high Br and S are still recorded until ca. 5050 cal. yr BP. Because this event tipped the system  
727 over a threshold (as observed in both PC1 and PC2 average model slopes; Figs. 6 and 7) into  
728 another stable state **from** which former trophic and mixing conditions never recovered, we  
729 consider this time interval as a tipping point in Lake Son Kol. A similar abrupt ecosystem  
730 threshold was reported at 4900 cal. yr BP based on chironomid remains, and attributed to the  
731 increased influence of the Westerlies causing enhanced precipitation and higher windiness at  
732 Lake Son Kol (Schwarz et al., 2017; Laug et al., 2020). Drastic environmental change in the  
733 lake may also be triggered by land-use processes in the catchment, which **are** mainly driven  
734 by changes in grazing pressure at high altitudes. Abundance of coprophilous fungal  
735 ascospores is assumed to reflect changes in grazing activity (Gauthier et al., 2010; Etienne et  
736 al., 2013). Influxes of coprophilous fungi *sensu stricto* fluctuated between 0 and 10  
737 spores.cm<sup>-2</sup>.yr<sup>-1</sup> between 6000 and 4000 cal. yr BP and during the last 2500 years (Fig. 10),  
738 suggesting the predominance of wild animals in the Lake Son Kol catchment. Nevertheless  
739 the presence of domestic herbivores cannot be excluded during 6000–4000 cal. yr BP, which  
740 would match the occurrence of human fecal biomarkers in Lake Chatyr Kol sediments around  
741 6000–5000 cal. yr BP (Schroeter et al., 2020). The imprint of domesticated livestock and  
742 human occupation was likely highest around 3700 and 2500 cal. yr BP, where influxes  
743 reached their maxima. Interestingly, influx values were generally stable during the period  
744 5500–5000 cal. yr BP, when an abrupt ecosystem shift occurred in Lake Son Kol (Figs. 9,  
745 10), thus implying that **pastoralism** apparently did not have a significant impact in the  
746 catchment and that climate is the main driver of the abrupt regime shift identified around  
747 5250 cal. yr BP.

748 After 5050 cal. yr BP, the lake became persistently well-mixed and the chemocline no longer  
749 existed in the lake water column, as indicated by pervasive PC2 > 0 values (Fig. 9) and  
750 elevated Mn content between 5000 and 3850 cal. yr BP that argue for oxic conditions in the  
751 lake. Surprisingly, the advent of well-mixed conditions is closely related to a strong decline of  
752 oxygen phototrophs (including cyanobacteria) between 5050 and 3850 cal. yr BP, with strong  
753 negative anomalies around 4400–3800 cal. yr BP, coeval to high Mn anomalies (Fig. 9).  
754 Because this phase corresponds to a noticeable decline in summer humidity and low lake  
755 levels (PC1 < 0; Fig. 9), we posit that cooler summer temperatures and longer lasting and  
756 thicker snow cover may have prevented the growth of oxygen phototrophs in the water  
757 column (as well as the establishment of a stable stratification in summer after ice cover).  
758 Alternatively, regular wind-driven lake mixing resulting from strengthened winds during the  
759 spring ice break-up season (inducing stronger vertical water circulation) may have caused low  
760 pigment deposition and preservation in the sediments (Cuddington and Leavitt, 1999). The  
761 disappearance of the chemocline after 5050 cal. yr BP coincides with a decrease in terrestrial  
762 DOM inputs into the lake, as catchment vegetation opened up with a decline of birch and  
763 other arboreal taxa and a rise in *Artemisia* (Fig. 10; Mathis et al., 2014), similarly as in Lake  
764 Qinghai (Ji et al., 2009). Tree removal around the lake thus probably led to higher wind  
765 exposure of the open-water area (due to increased wind fetch), inducing in turn a better  
766 mixing of the water column. Lake levels increased around 3700 cal. yr BP, corresponding to  
767 another main tipping point based on high PC1 values (Fig. 6), in line with increased  
768 planktonic diatom abundances and lower conductivities (Fig. 9; Schwarz et al., 2017). High  
769 PC2 (and low Bphe *a*) values from 3800 cal. yr BP onwards infer that repetitive mixing  
770 conditions prevented the establishment of a chemocline by oxygenating the hypolimnion  
771 (Butz et al., 2016), which hampers the onset of stable temperature stratification during  
772 summer. Such conditions promoted nutrient cycling in the photic zone to support oxygenic



773 phototrophs in the epilimnion, which diminished light penetration to the chemocline and led  
774 to the **destruction** of a suitable lake habitat for APB growth. Hence, all proxy data suggest that  
775 Lake Son Kol has had persistently high lake levels with well-mixed and oxygenated  
776 conditions for the last 5000 years, in accordance with its modern polymictic regime.

777

## 778 **6. Conclusions and perspectives**

779 Ultra-high temporal resolution geochemical and hyperspectral palaeolimnology data can yield  
780 a more complete understanding of changing ecological patterns and redox processes in  
781 shallow lacustrine systems. As a biomarker of planktonic APB at the chemocline of lakes, the  
782 presence/absence and variability of Bphe *a* in Lake Son Kol sediments provides crucial  
783 information about the timing and forcing mechanisms that control the advent and vanishing of  
784 lake stratification over the last thousands of years. Increased Bphe *a*, Br, S abundances along  
785 with higher terrestrial inputs from the catchment are recorded at 8500, 8400, 8200–7800,  
786 7700–7500, 7300–7000, 6500–6100, 6000–5700 and 5500–5250 cal. yr BP, and involve  
787 recurring episodes of hypolimnetic anoxia and euxinic bottom water conditions coeval to the  
788 deposition of dark organic sediments on the lake floor. The presence of a chemocline and the  
789 associated deposition of coarse minerogenic detritus transported by concentrated flows  
790 appears to be dependent on recurrent seasonal snowmelt episodes, and thus on the quantity of  
791 snow and spring/summer temperatures that regulate the potential for meltwater events. Our  
792 data thus suggest that low lake levels and warmer climate conditions regularly prevailed  
793 during 8500–5250 cal. yr BP, and promoted massive APB blooms in a well-stratified lake  
794 system. After 5250 cal. yr BP, Son Kol water level **recovered** and the lake became persistently  
795 well mixed, leading to the disappearance of the chemocline. This abrupt ecosystem shift was  
796 associated with cooler conditions and strengthened wind forcing, when the influence of the  
797 Westerlies increased from 5250 cal. yr BP onwards. Intensified wind forcing also most

798 probably coincided with a significant increase of wave activity at the lake shoreline acting as  
799 an erosive agent. The imprint of wind forcing on coastal sedimentary dynamics (e.g., littoral  
800 sand drift) and geomorphological processes can be observed based on the diagnostic  
801 sedimentary features along the northeastern and southeastern shores of the lake, where sand  
802 spits, wave-dominated beaches, barrier spits and associated open lagoons have developed,  
803 hence suggesting that Lake Son Kol is presently a wind-driven water body (as defined by  
804 Nutz et al., 2018). Further studies are required in order to properly date such  
805 geomorphological features and assess the relationships between wind forcing, past lake levels  
806 and climatic trends during the Mid- to Late Holocene.

807

#### 808 **Acknowledgments**

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810 on core SK06. The hyperspectral imaging acquisition was conducted at the University of  
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812 TPE). All data from Son Kol core SK06 are available as supplementary data via the QSR  
813 network.

814

#### 815 **Figure captions:**

816

817 **Figure 1:** Location of Lake Son Kol: (a) global map with focus on western central Asia; (b)  
818 central Tian Shan including south Kazakhstan and Lake Balkash (image courtesy of NASA  
819 GSFC); and (c) bathymetric map of the lake, with isobaths given in meters below lake level  
820 (modified after Academy of Science of the Kyrgyz SSR, 1987). Orange points refer to  
821 sediment core SK06, while the two yellow and the white points indicate cores SK07 and  
822 SK04 (Mathis et al., 2014; Huang et al., 2014; Pacton et al., 2015) and SONK\_11\_D1/2

823 (Lauterbach et al., 2014), respectively.

824

825 **Figure 2:** (a) Detailed stratigraphic correlation between cores SK06 (this study) and cores  
826 SK04 and SK07 (Huang et al., 2014; Mathis et al., 2014) from Lake Son Kol using distinct  
827 lithological marker layers. Studied cores are located on the map provided in Fig. 1. (b) Age  
828 model for core SK06 based on the *P\_Sequence* depositional model (model parameter  $k=1$ )  
829 established using Oxcal 4.4 (Ramsey, 1995, 1998, 2001). The blue envelope corresponds to  
830 the  $2\sigma$  probability range. See text for detail.

831

832 **Figure 3:** (a) Visible signatures and (b) continuum removal spectra with associated main  
833 absorptions for Chl *a* (and derivatives), phycocyanin and Bphe *a* (Bacteriopheophytin *a*) from  
834 three selected intervals in which these pigments are abundant in core SK06.

835

836 **Figure 4:** (a) PCA biplot of major elements obtained by XRF core scanner and hyperspectral  
837 analyses (PC1 vs PC2). PC1 vs PC3 and PC2 vs PC3 are shown in Fig. S4. (b) Comparison  
838 between Ca, Fe and Ti content from XRF core-scanner and discrete XRF analyses in core  
839 SK07.

840

841 **Figure 5:** Hyperspectral image and spectral indices (relative absorption band area or RABA)  
842 for core SK06. From left to right: 500- $\mu\text{m}$  resolution Br and S (XRF core-scanner),  
843 distribution maps and plots of Chl *a* plus derivatives (RABA630–700), phycocyanin  
844 (RABA600–630), Bacteriopheophytin (Bphe *a*) (RABA750–900) and aromatic organic  
845 matter (ARO; RABA1660–1690), respectively. Grey shadings outline intervals with higher  
846 Br, S, phycocyanin and Bphe *a* contents, which are tied to the occurrence of dark organic

847 sediments in core SK06. Yellow shadings outline intervals with elevated content in Chl *a* and  
848 phycocyanin in core SK06.

849

850 **Figure 6:** Average changepoint model's slope calculated for PC1. The green curve in the  
851 individual plots outlines the most pronounced regime shift identified around 5050 cal. yr BP.  
852 Purple-filled curves correspond to abrupt shifts identified in both PC1 and PC2.

853

854 **Figure 7:** Average changepoint model's slope calculated for PC2. The green curve in the  
855 individual plots outlines the most pronounced regime shift identified around 5250 cal. yr BP.  
856 Purple-filled curves correspond to abrupt shifts identified in both PC1 and PC2.

857

858 **Figure 8:** Close-up on XRF and hyperspectral data for the interval 131.5–151.5 cm in core  
859 SK06 (ca. 6500–5600 cal. yr BP), plotted against grain size (clay, sand), stable isotope ( $\delta^{18}\text{O}$ ,  
860  $\delta^{13}\text{C}_{\text{org}}$ ), total organic matter (TOC), total carbon (TC), total nitrogen (TN) and C/N data from  
861 core SK07 in Lake Son Kol (Huang et al. 2014; unpublished data). Yellow shadings refer to  
862 intervals characterized by dark organic layers with coarser minerogenic detritus, high Br, S,  
863 Bphe *a*, ARO in core SK06, as well as high TOC (and TC, TN and C/N) and depleted  $\delta^{18}\text{O}$   
864 and  $\delta^{13}\text{C}_{\text{org}}$  values in core SK07.

865

866 **Figure 9:** Time series of hyperspectral, geochemical and PCA data for Lake Son Kol (core  
867 SK06), plotted with proxy data from core SONK\_11\_D1/2 from Lake Son Kol (Lauterbach et  
868 al., 2014; Schwarz et al., 2017). XRF (9-point running means) and hyperspectral data (15-  
869 point running means) for core SK06 are represented as standardized values (mean = 0,  
870 standard deviation = 1). Yellow shadings outline periods for which hypolimnetic euxinia  
871 prevailed in Lake Son Kol. See text for detail.

872

873 **Figure 10:** Comparison of geochemical and palynological proxies from Lake Son Kol to  
874 other regional records in Central Asia. Shaded areas indicate periods characterized by  
875 relatively warm summers and milder winters/springs with intensified precipitation (rain or  
876 snow) at Lake Son Kol, favoring enhanced winter/spring snow thaw episodes; (a) *Bphe a*  
877 record from Lake Qinghai, China (Ji et al., 2009); (b) alkenone-inferred temperature record at  
878 Lake Qinghai (Hou et al., 2016); (c) speleothem  $\delta^{18}\text{O}$  record from Tonnel'naya (Ton) Cave,  
879 Uzbekistan (Chen et al., 2016); (d) speleothem  $\delta^{18}\text{O}$  record from Baluk Cave, China (Liu et  
880 al., 2019); (e) influx of coprophilous fungi *sensu stricto* (core SK06, this study); (f) arboreal  
881 pollen (%) record from Lake Son Kol, Kyrgyzstan (Mathis et al., 2014); (g) total content of  
882 Betulaceae (%) from Lake Son Kol, Kyrgyzstan (Mathis et al., 2014) and (h) *Bphe a* record  
883 (15-point running means) from Lake Son Kol (core SK06, this study).

884

885 **Figure S1:** Core-to-core correlation between sediment cores SK06, SK04 and SK06 using  
886 magnetic susceptibility.

887

888 **Figure S2:** Correlation of core SK06 (this study) with core SONK\_11\_D1/2 from Lauterbach  
889 et al., (2014) using distinct lithological marker layers.

890

891 **Figure S3:** 500- $\mu\text{m}$  resolution (grey curves) XRF core-scanner data exhibiting Br, S, Ti, Rb,  
892 Zr, Fe, Si, Ca, Sr and Mn. Red curves represent 31-point running means. Yellow shadings  
893 outline intervals with high Br, S, Ti, Rb and low Zr, Ca, Sr and Mn, which are tied to the  
894 occurrence of dark organic sediments in core SK06.

895

896 **Figure S4:** PCA biplots of major elements obtained by XRF core scanner and hyperspectral  
897 analyses (PC1 vs PC3 and PC2 vs PC3).

898

899 **Figure S5:** Average changepoint models of PC1 and PC2 during the last 8500 years in core  
900 SK06.

901

902 **Figure S6:** Thin-section images (cross-polarized light) of the detrital layer microfacies types  
903 observed in dark organic-rich sediments; (a) minerogenic detrital layer characterized by rare  
904 normal graded beds in a non-graded matrix-supported (i.e., fine silt) layer. Note the absence  
905 of a sharp erosive basal layer; (b) a typical detrital layer exhibiting mixed organic (e.g.,  
906 phytoclasts), bioclasts (e.g., ostracod valves) and minerogenic detritus. Note the absence of  
907 grading into the detrital layer; (c, d) organic detrital layer mainly consisting of phytoclasts,  
908 interpreted as reworked terrestrial material (sheet-wash) from the catchment.

909

910 **Table 1:** AMS  $^{14}\text{C}$  dates obtained on terrestrial plant remains and bulk sediment OM (with  
911 unknown proportion of aquatic OM) in cores SK07 and SK04 (Huang et al., 2014) and core  
912 SONK\_11\_D1/2 (Lauterbach et al., 2014). Published AMS  $^{14}\text{C}$  dates were transposed to core  
913 SK06 using lithological and magnetic susceptibility-inferred correlations (Figs. 2a, S1 and  
914 S2). Conventional  $^{14}\text{C}$  ages were calibrated using Oxcal 4.4 with the IntCal20 calibration  
915 dataset (Reimer et al., 2020). A reservoir correction of 150 years was applied before  
916 calibration to conventional  $^{14}\text{C}$  ages. Only the samples with asterisks (\*) were considered for  
917 age modeling (see text for detail).

918

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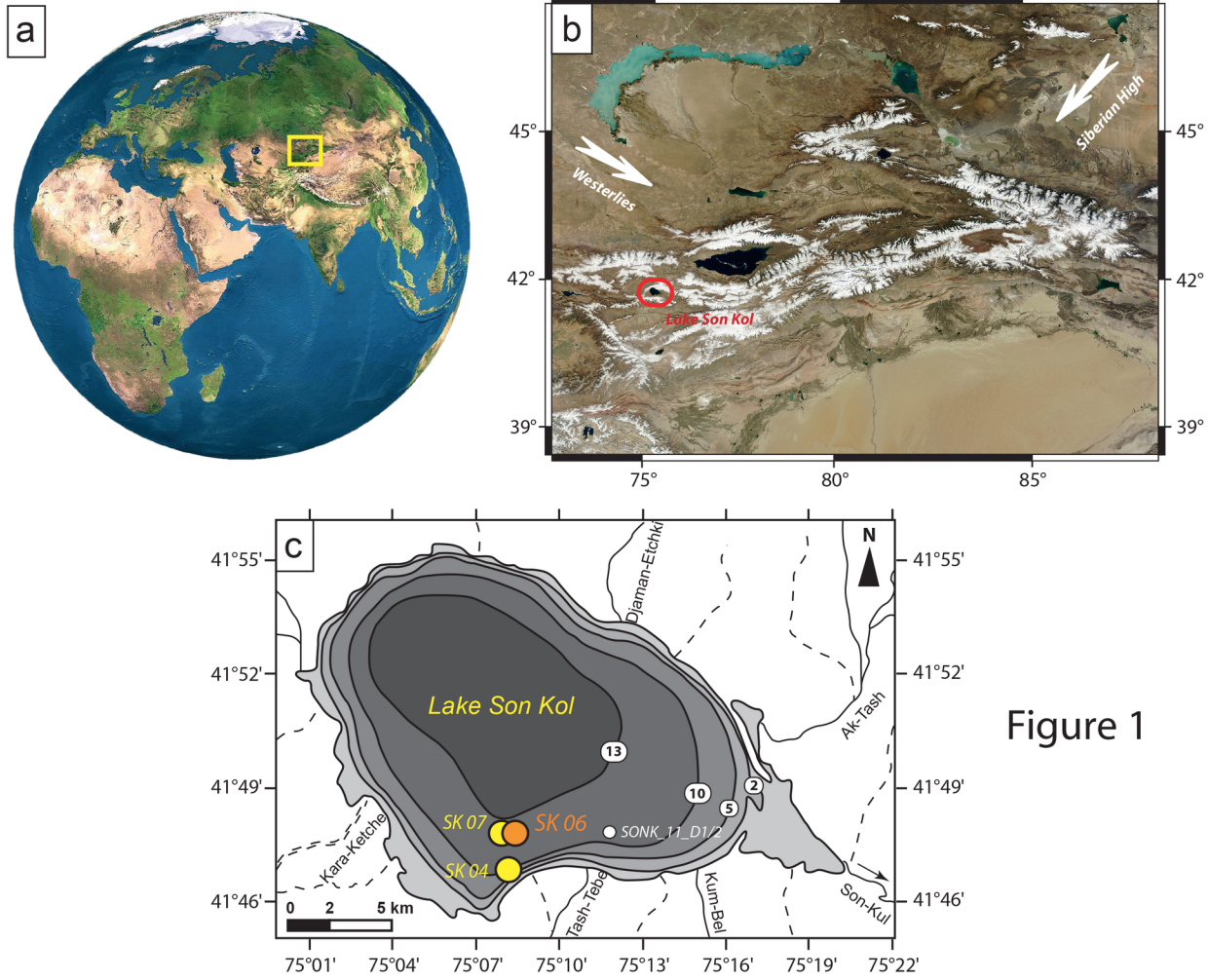


Figure 1



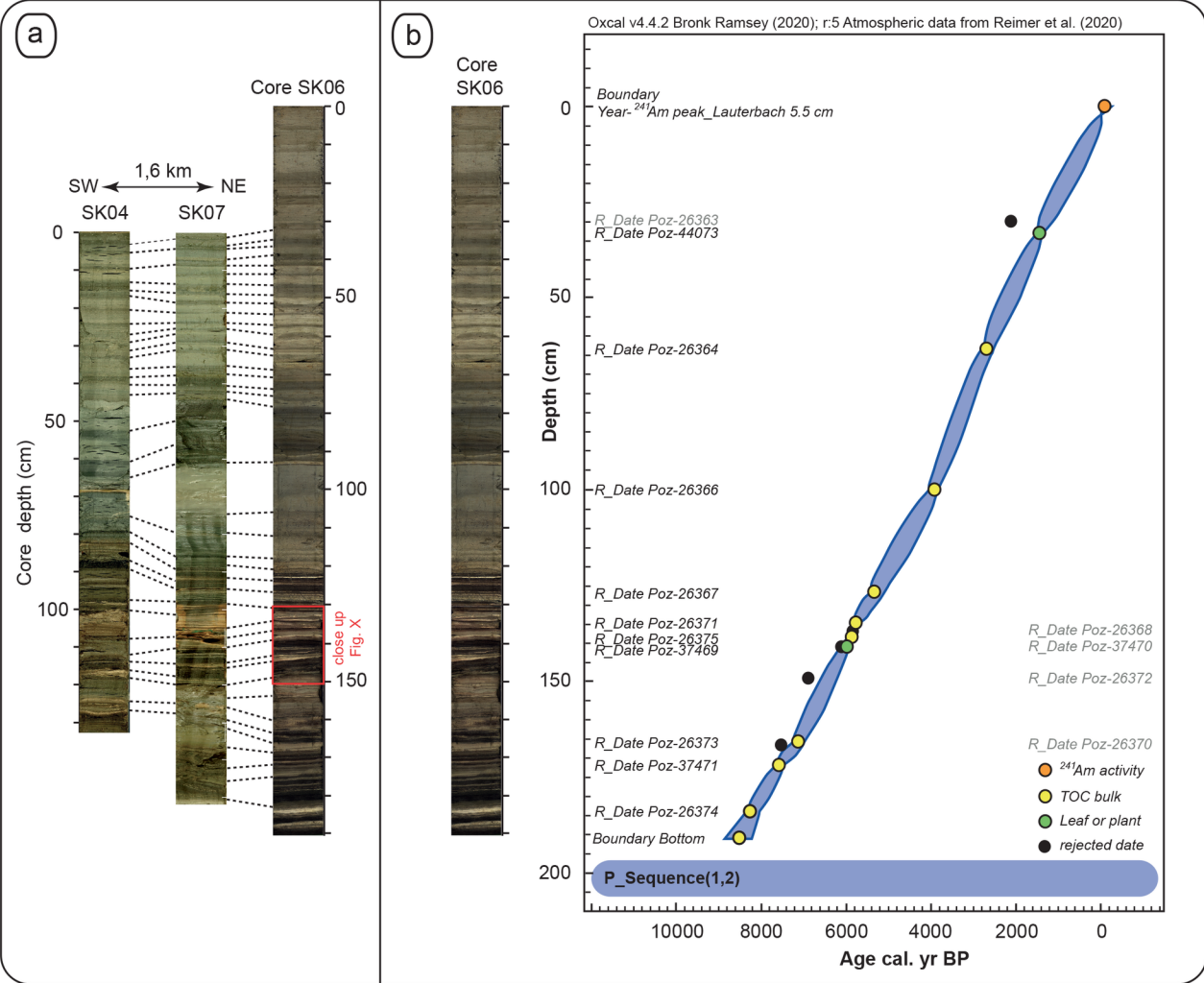


Figure 2

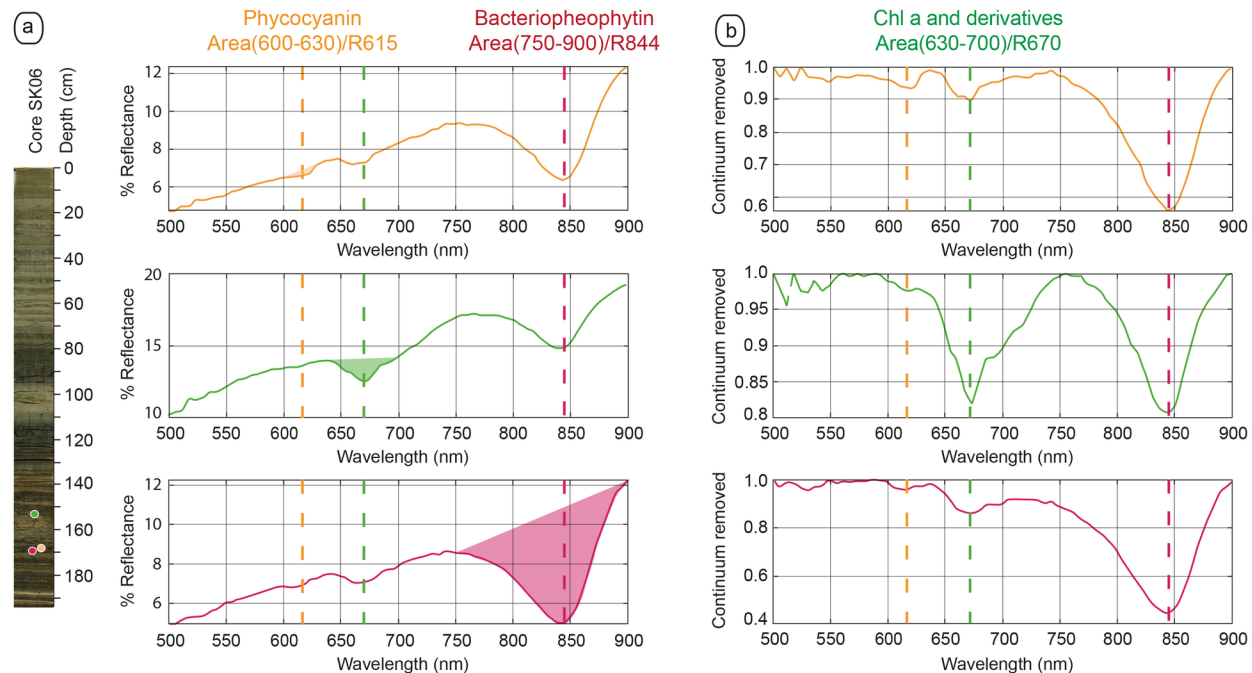


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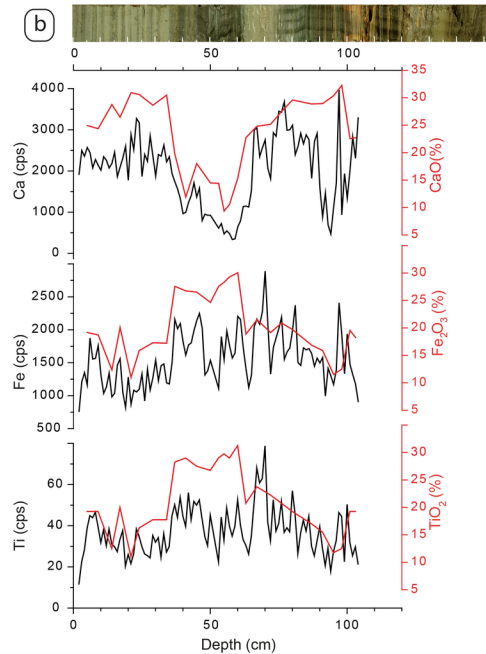
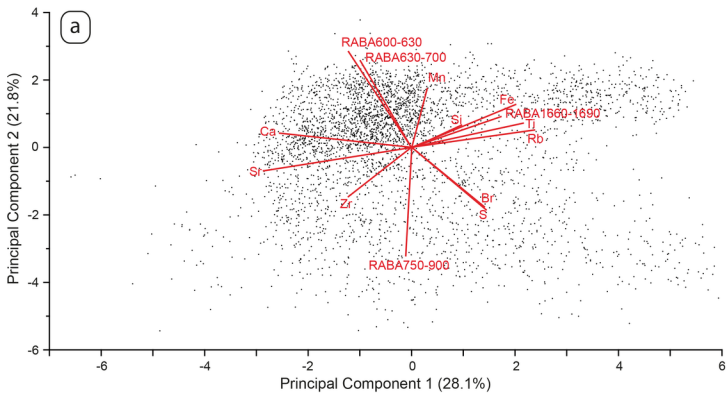


Figure 4

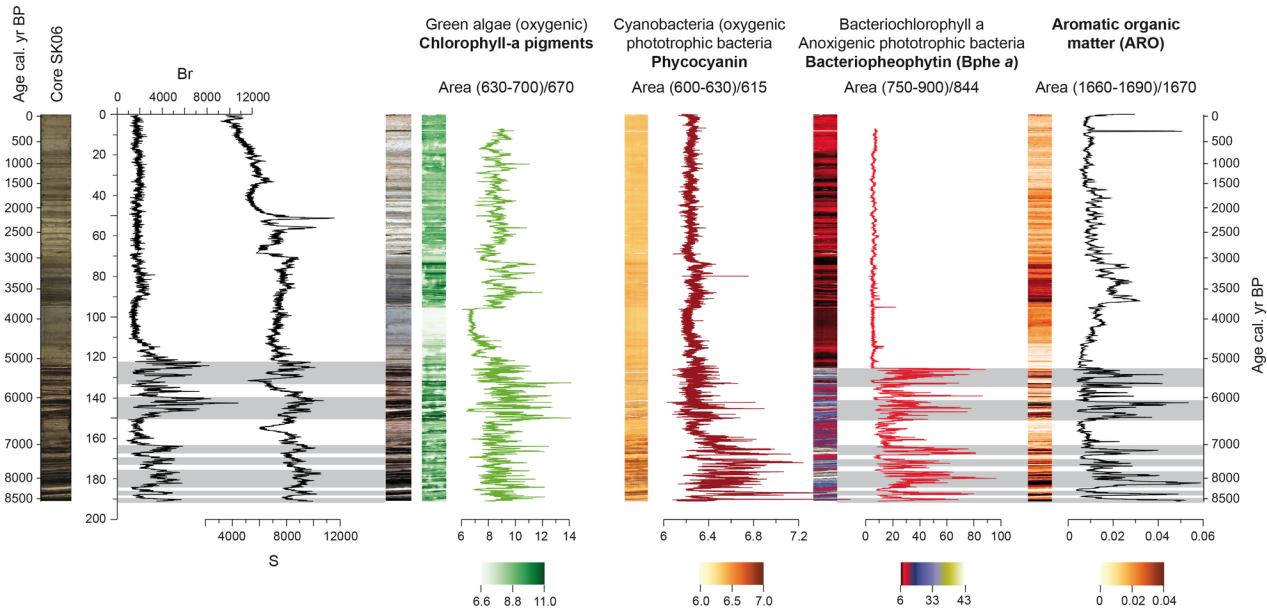


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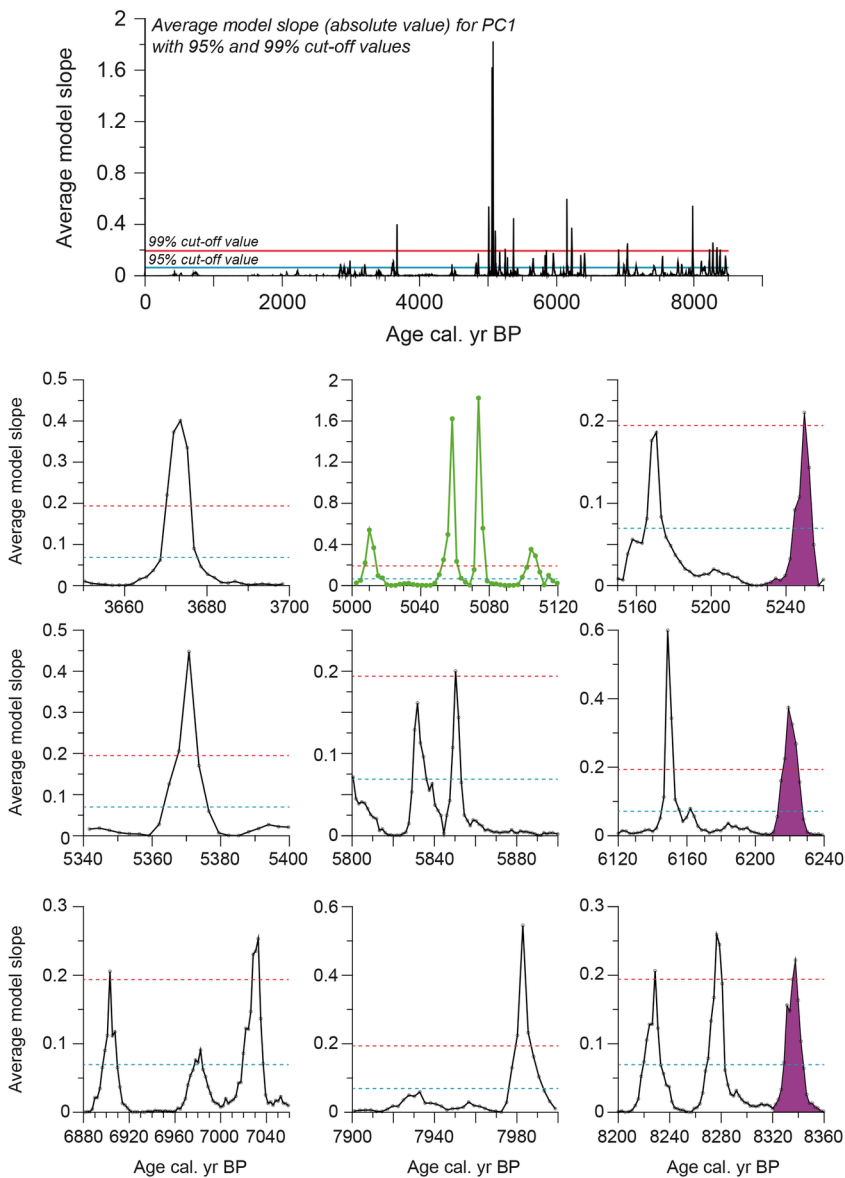


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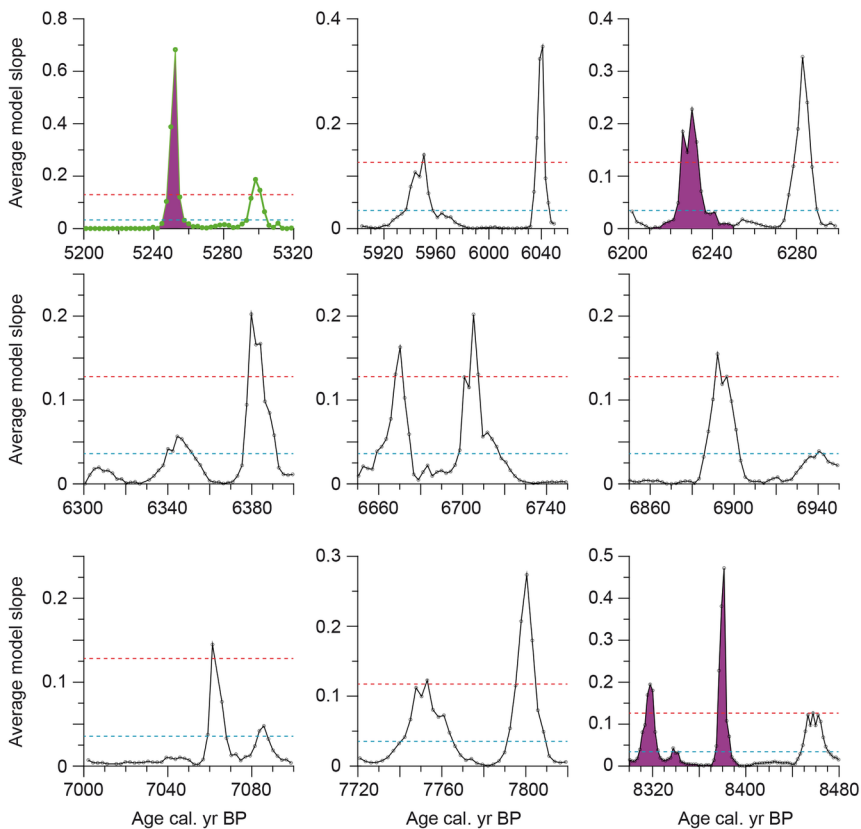
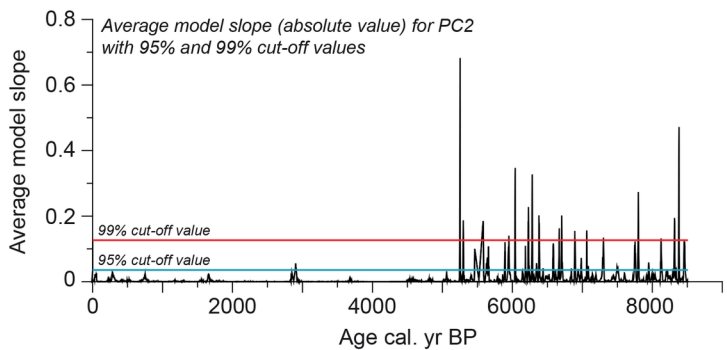


Figure 7

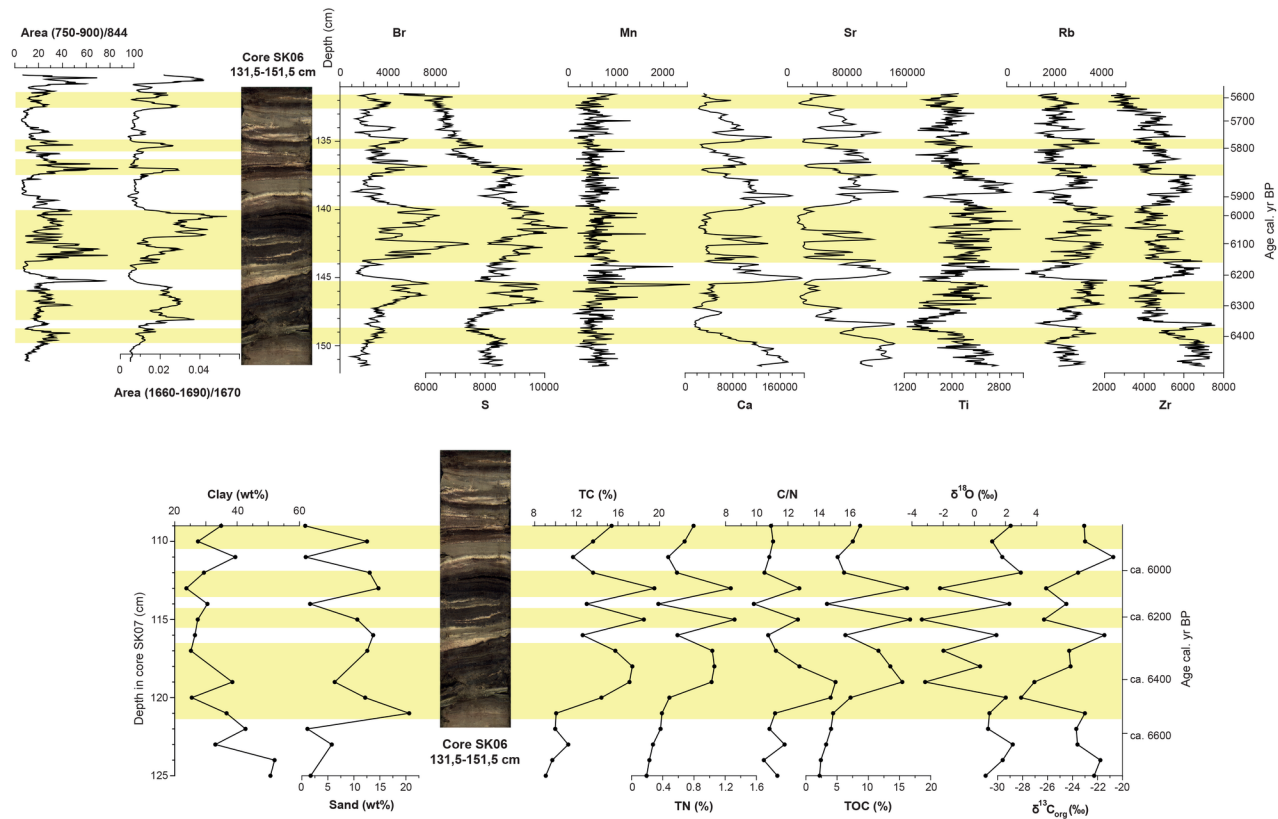


Figure 8

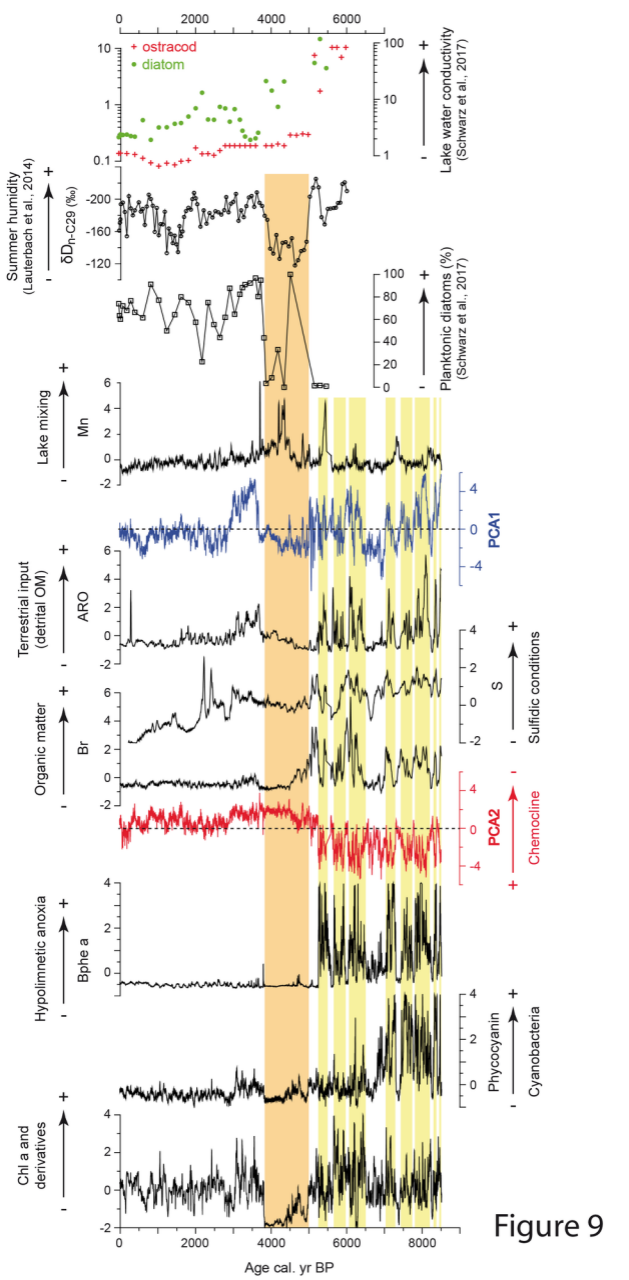


Figure 9



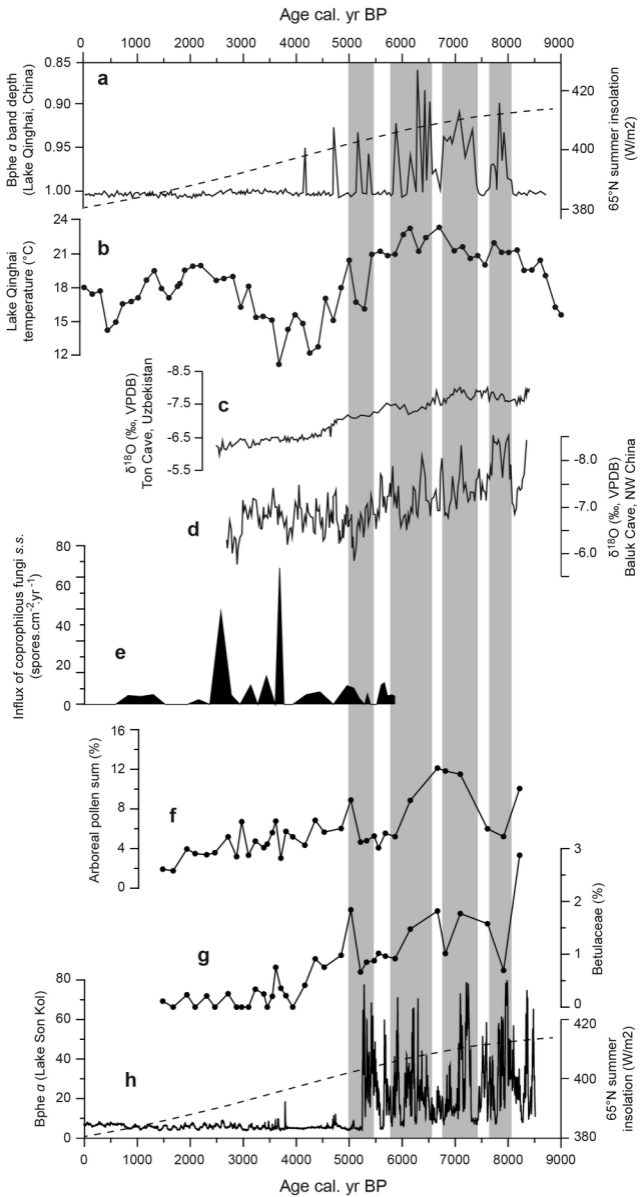


Figure 10

Lab. ID	Sampling interval	Dated material	Correlation LITHO in Core SK6 (cm)	AMS <sup>14</sup> C age (yr BP + $\sigma$ )	Corr. AMS <sup>14</sup> C age (yr BP + $\sigma$ )	Calibrated age (cal. yr BP + 2 $\sigma$ )
	SONK_11_D1/2 (Lauterbach et al., 2014)	<sup>241</sup> Am peak (5,5 cm)	2			-13
<b>Poz-26363</b>	<b>Sonkul 4A 0-75 0-1cm</b>	<b>Bulk sediment</b>	<b>30</b>	<b>2000 ± 30</b>	<b>1850 ± 30</b>	2139**
*Poz-44073	SONK_11_D1/2 (Lauterbach et al., 2014)	Terrestrial plant remains	33	1575 ± 35		
*Poz-26364	Sonkul 4A 0-75 30-31cm	Bulk sediment	63,5	2770 ± 40	2620 ± 40	
*Poz-26366	Sonkul 4A 75-150 0-1cm	Bulk sediment	100	3750 ± 35	3600 ± 35	
*Poz-26367	Sonkul 4A 75-150 20-21cm	Bulk sediment	127	4730 ± 40	4580 ± 40	
*Poz-26371	Sonkul 7A 120-180 0-1cm	Bulk sediment	135	5270 ± 40	5120 ± 40	
<b>Poz-26368</b>	<b>Sonkul 4A 75-150 40-41cm</b>	<b>Bulk sediment</b>	<b>137</b>	<b>5240 ± 40</b>	<b>5090 ± 40</b>	5817**
*Poz-26375	Sonkul 7A 120-180 5cm	Bulk sediment	139	5250 ± 40	5100 ± 40	
<b>Poz-37470</b>	<b>Sonkul 7A 120-180 cm 7.2 cm- 8.6cm</b>	<b>Bulk sediment</b>	<b>141</b>	<b>5410 ± 40</b>	<b>5260 ± 40</b>	6052**
*Poz-37469	Sonkul 7A 120-180 cm 7.2 cm- 8.6cm	Terrestrial plant remains	141	5460 ± 40	5310 ± 40	
<b>Poz-26372</b>	<b>Sonkul 7A 120-180 15.5-16.5cm</b>	<b>Bulk sediment</b>	<b>149</b>	<b>6200 ± 50</b>	<b>6050 ± 50</b>	6900**
<b>Poz-26370</b>	<b>Sonkul 4A 75-150 65-66cm</b>	<b>Bulk sediment</b>	<b>167?</b>	<b>6790 ± 40</b>	<b>6640 ± 40</b>	7521**
*Poz-26373	Sonkul 7A 120-180 26.5-27.5cm	Bulk sediment	166	6330 ± 50	6180 ± 50	
*Poz-37471	Sonkul 7A 120-180 cm 37.5-38.5cm	Bulk sediment	172	6880 ± 50	6730 ± 50	
*Poz-26374	Sonkul 7A 120-180 45.5-46.3cm	Bulk sediment	184	7570 ± 50	7420 ± 50	

The asterisks (\*) denotes the radiocarbon dates applied to establish the age modelling. Bold samples correspond to excluded dates from age-dep  
Double asteriks (\*\*) correspond to median probability ages as calculated online with Oxcal8.2