

1 **Late Holocene Indian summer monsoon variability revealed from a 3,300-year-long**  
2 **lake sediment record from Nir'pa Co, southeastern Tibet**

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19 Anomaly, Little Ice Age, Hydroclimate, Pluvial, Drought

20

21 **Abstract**

22 Sedimentological and geochemical results from Nir'pa Co, an alpine lake on the  
23 southeastern Tibetan Plateau detail late Holocene ISM hydroclimate during the last 3,300

24 years. Constrained by modern calibration elevated silt and lithics and low sand and clay  
25 between 3.3-2.4 ka and 1.3 ka to the present indicate two pluvial phases with lake levels  
26 near their current overflow elevation. Between 2.4-1.3 ka, a sharp increase in sand and  
27 corresponding decrease in lithics and silt suggest drier conditions and lower lake levels at  
28 Nir'pa Co. Hydroclimate expressions in the sedimentological proxies during the  
29 Medieval Climate Anomaly (MCA) and Little Ice Age (LIA) are not statistically  
30 significant, suggesting that these events were minor compared to the millennial scale  
31 variability on which they were superimposed. However, decreasing sand and increasing  
32 lithics and silt during the MCA between 950-800 cal yr B.P. may suggest briefly wetter  
33 conditions while increasing sand and reduced lithics and silt from 500-200 cal yr B.P.  
34 suggest potentially drier conditions during the LIA. Similarities with regional records  
35 from lake, ice core and speleothem records from the central and eastern Tibetan Plateau,  
36 India, and the Arabian Sea, suggest generally coherent late Holocene ISM variability in  
37 these regions. Increased late Holocene ISM intensity occurred during times when Tibetan  
38 Plateau surface air temperatures were warmer, Indo-Pacific sea surface temperatures  
39 were elevated, and the tropical Pacific was in a La Niña-like mean state. Conversely,  
40 aridity between 2.4-1.3 ka occurred in concert with cooling on the Tibetan Plateau and in  
41 the Indo-Pacific with more El Niño-like conditions in the tropical Pacific. Differences  
42 with western Tibetan records may reflect a weakened ISM and stronger westerlies in this  
43 region during the late Holocene.

44

## 45 **1. Introduction**

46 Holocene paleoclimate records from the Tibetan Plateau and surrounding regions,  
47 an area referred to as the Third Pole Environment (Qiu, 2008), show that the Indian  
48 summer monsoon (ISM) reached a maximum during the early and middle Holocene  
49 between 10 and approximately 6 ka when boreal summer insolation peaked and strong  
50 tropical Pacific sea surface temperature (SST) gradients resembled a La Niña-like mean  
51 state (Koutavas et al., 2006; He et al., 2004; Bird et al., 2014; Cai et al., 2012; Yanhong  
52 et al., 2006; Marchitto et al., 2010; Morrill et al., 2003). The late Holocene, in contrast,  
53 has been characterized by increasing aridity as summer insolation waned and an  
54 increasingly El Niño-like mean state developed in the tropical Pacific. Despite an  
55 improved understanding of long-term Holocene ISM trends and the likely driving  
56 mechanisms, higher frequency ISM variations have been more difficult to characterize  
57 because there are comparatively few high-resolution records from the Third Pole (Chen et  
58 al., 2015). The late Holocene is under represented by high-resolution paleoclimate  
59 records in part because mean state aridity resulted in decreased resolution (e.g., Bird et  
60 al., 2014) or depositional hiatuses in many ISM archives (Berkelhammer et al., 2012;  
61 Morrill et al., 2006; Cai et al., 2012). The paucity of climate records detailing ISM  
62 variability during the last few millennia limits our understanding of monsoon dynamics  
63 during a time when climatic boundary conditions were generally similar to the present,  
64 but perturbations in radiative and other forcings, including ocean-atmosphere processes,  
65 produced widely recognized climatic events including: the Medieval Climate Anomaly  
66 (MCA; 950-1250 AD) and Little Ice Age (LIA; 1450-1900 AD)(Mann et al., 2009).  
67 Paleoclimate records from the Third Pole that span at least part of the late  
68 Holocene suggest that within this period of general aridity there were regional scale

69 fluctuations in ISM precipitation, lake levels and temperature (Bird et al., 2014;  
70 Herzsuh et al., 2006; Mischke and Zhang, 2010). Similar variability observed on longer  
71 timescales in Holocene-length paleoclimate records suggest that sub-orbital climate  
72 changes may be influenced by a variety of direct and indirect factors, including solar  
73 variability (Cai et al., 2012), Atlantic and Indo-Pacific ocean-atmosphere teleconnections  
74 (Chen et al., 2015; Bird et al., 2014; Hong et al., 2003; Goswami et al., 2006), and  
75 Tibetan Plateau surface air temperatures (Bird et al., 2014). At Paru Co, an alpine lake in  
76 the monsoon dominated southeastern Tibetan Plateau, Bird et al. (2014) noted that low  
77 lake stands and high hydrogen isotopic compositions of sedimentary leaf waxes ( $\delta D_{\text{wax}}$ )  
78 coincided with warmer Indo-Pacific SST (Koutavas et al., 2006; Stott et al., 2004), cooler  
79 eastern equatorial Pacific SST (Marchitto et al., 2010) and elevated surface air  
80 temperatures on the central Tibetan Plateau (Herzsuh et al., 2006). The opposite  
81 associations were noted during low lake stands. Late Holocene aridity on the Third Pole  
82 occurred within the context of more El Niño-like Pacific SST structure and cooling over  
83 the Tibetan Plateau (Bird et al., 2014). Centennial lake level variability followed this  
84 general warm-wet, cool-dry pattern. Persistently low lake levels at Paru Co during the  
85 late Holocene, however, limited the lake's sensitivity to hydroclimate changes, making it  
86 difficult to discern ISM variability at centennial or shorter time scales. Because many  
87 other records also lose resolution during this time, the role that regional SAT and/or Indo-  
88 Pacific SST played in ISM variability during the last ~4000 years is not well defined.  
89 Filling this gap in knowledge is important because changes in radiative forcing and  
90 ocean-atmosphere systems, particularly during the MCA and LIA, have been suggested  
91 to have produced significant precipitation anomalies in the East Asian and South

92 American monsoon systems (Zhang et al., 2008; Bird et al., 2011; Chen et al., 2015). In  
93 the ISM region, however, hydroclimatic signatures of the MCA and LIA in paleoclimate  
94 records are more equivocal. Some records, such as accumulation at Dasuopu (Thompson  
95 et al., 2000) and  $\delta^{18}\text{O}$  from Dandak Cave (India; Sinha et al., 2006) and Sahiya Cave  
96 (Sinha et al., 2015), suggest possibly wetter conditions during the MCA and drier  
97 conditions during the LIA, while other studies suggest the opposite (Liu et al., 2009).  
98 Excluding the Third Pole, the MCA and LIA appear to have produced an east-west  
99 hydroclimate dipole response north of 30°N whereby northwest and central Asia were  
100 drier and northern to northeastern China wetter during the MCA (Chen et al., 2015). The  
101 opposite conditions are suggested for the LIA. In order to help resolve questions  
102 regarding late Holocene ISM dynamics and their relationships with hydroclimate  
103 variability across Asia, additional records from core ISM regions are needed.

104 To this end, we present a 3,300-year-long multi-proxy hydroclimate record from  
105 Nir'pa Co that details the hydroclimatic variability of a region highly sensitive to changes  
106 in ISM precipitation. Through comparison with existing paleoclimate records from the  
107 Third Pole and other climatically significant regions, we investigate the regional  
108 expression of late Holocene ISM precipitation (including the MCA and LIA), its  
109 relationship with known climatic and ocean-atmosphere influences and whether Indo-  
110 Pacific and ENSO-like forcing mechanisms of ISM variability suggested for the early and  
111 middle Holocene can similarly account for late Holocene ISM variations.

112 [Insert Figure 1]

113 *1.1 Study Area: Nir'pa Co*

114 Nir'pa Co is a small (0.1 km<sup>2</sup>) alpine lake situated at 4775 m above sea level (asl)  
115 in the Nyanchentangula Mountains on the southeastern Tibetan Plateau (Fig. 1, 2a; N  
116 29.734°, E 92.386°). Nir'pa Co's watershed is defined by a 5.38 km-long and 1.3 km-  
117 wide glacially carved valley that trends northwest-southeast with a maximum elevation of  
118 5150 m asl (Fig. 2a). This valley is part of a larger plateau that contains numerous glacial  
119 features, such as moraines and glacial lakes, including Paru Co (Bird et al., 2014). The  
120 local bedrock is geologically homogeneous, consisting of Pleistocene age andesite  
121 (Choubert et al., 1983). The lake is deep relative to others in the area, with a maximum  
122 depth of 18.6 m. Nir'pa Co's bathymetry is generally steep sided with a flat bottom that  
123 is deepest near the center of the lake (Fig. 2b). An ephemeral stream enters on the lake's  
124 northwestern shore, at which point it forms a delta on a shallow littoral shelf (Fig. 2b). A  
125 channel at the lake's southeast shore provides an outlet for overflow that has incised into  
126 the terminal glacial moraine that dams the lake. This outlet was overflowing, albeit with  
127 restricted flow, when cores were collected from Nir'pa Co in May 2011 and June 2015,  
128 suggesting that the lake is currently at or near its maximum surface elevation but  
129 becomes seasonally restricted.

130 [Insert Figure 2]

131 Indian summer monsoon moisture largely derived from the Bay of Bengal is the  
132 primary source of precipitation for regions of the Tibetan Plateau between the Himalayan  
133 and Tanggula Mountains and for interior monsoonal regions in India (Tian et al., 2001;  
134 Tian et al., 2007; Sinha et al., 2015). Consistent with this, 92 % of mean annual  
135 precipitation at Nir'pa Co occurs during the monsoon season with peaks between May  
136 and September when temperatures rise above freezing and the lake is ice free and

137 sediment can be transported from the watershed and deposited in Nir'pa Co (Fig. 2c;  
138 Peterson and Vose, 1997; Lawrimore et al., 2011; Bird et al., 2014). Precipitation during  
139 the remainder of the year accounts for only 8 % of the annual hydrologic budget and is  
140 received primarily as snow when average monthly temperatures are near or below 0°C  
141 from November through April. During these months, Nir'pa Co is covered with ice,  
142 limiting sediment input to the lake. Tropical Rainfall Measuring Mission (TRMM –  
143 3B43) precipitation data estimates suggest regional annual rainfall totals 1118 mm yr<sup>-1</sup>  
144 while instrumental weather station data indicate annual rainfall amounts to 479 mm yr<sup>-1</sup>.  
145 The difference between these amounts may be due in part to over estimations of rainfall  
146 at elevation by TRMM data and underestimations of rainfall at elevation based on rainfall  
147 gauge measurements made in valleys. Regardless, the majority of precipitation at Nir'pa  
148 Co in either data set occurs during the summer monsoon season when the lake is ice-free  
149 and sediment can be mobilized from the lake's catchment and transported to Nir'pa Co  
150 via fluvial processes.

151         A number of ocean-atmosphere and terrestrial climate feedbacks influence ISM  
152 precipitation (Peings and Douville, 2010; Ashok et al., 2004; Shaman and Tziperman,  
153 2005). Interannual Indo-Pacific ocean-atmosphere variability associated with the El  
154 Niño-Southern Oscillation (ENSO) is a primary modulator of ISM rainfall (Ashok et al.,  
155 2004; Gadgil, 2003; Kumar et al., 2006; Krishnamurthy and Kirtman, 2009).  
156 Observational records show that cooling in the Indo-Pacific during warm El Niño events  
157 coincides with increased atmospheric subsidence and reduced precipitation over the  
158 Asian monsoon region. During cool La Niña events, warm Indo-Pacific SST enhance  
159 convergence and vertical air motion over the Asian monsoon region, thereby contributing

160 to enhanced monsoon rainfall. ENSO-like mechanisms have been invoked on longer  
161 timescales to explain multi-decadal to millennial trends in ISM and Asian monsoon  
162 rainfall, whereby persistent El Niño or La Niña-like conditions in the tropical Pacific  
163 created background conditions of either persistently stronger or weaker monsoon  
164 circulation over Asia (Bird et al., 2014; Hong et al., 2005; Krishnan and Sugi, 2003).

165 The Indian Ocean Dipole (IOD) is another recognized driver of ISM variability,  
166 which may also contribute to long-term trends in monsoon rainfall (Ashok et al., 2004).  
167 The IOD, however, is not wholly independent from ENSO and is considered by some to  
168 be an extension of ENSO variability in the Indian Ocean, albeit with some independent  
169 behavior (Ashok et al., 2001; Ashok et al., 2004). Warm IOD events are typically  
170 associated with La Niñas and enhanced ISM rainfall. Cool IOD events are conversely  
171 associated with El Niño events and reduced ISM rainfall. The IOD's influence on ISM  
172 rainfall during the Holocene is not well known, however, because there are a limited  
173 number of SST reconstructions from key centers of IOD action in the Indian Ocean and  
174 none that continuously span the late Holocene.

175

## 176 **2. Methods**

### 177 *2.1 Sediment core and sample collection*

178 A 90 cm surface core (A-11) was collected from Nir'pa Co in May 2011 in 17.5  
179 meters of water using an Aquatic Research hammer core (Fig. 2b). Two additional  
180 surface cores (A-15 & B-15) were acquired in June 2015 using a piston coring system in  
181 water depths of 16.5 and 1.0 m, respectively (Fig. 2b). The top 10 cm of A-11 were  
182 extruded at 0.5 cm intervals in the field immediately following collection. Cores A-15



183 and B-15 were not extruded in the field, but instead secured with absorbent foam, which  
184 stabilized the sediments for transport. Visible stratigraphy in the upper portion of B-15  
185 and the lack of void space in A-15 indicated that the sediments didn't mix during  
186 transport. In order to characterize modern sediment distributions, grab samples of the  
187 upper 2 cm were collected using an Ekman grab sampler along a southeast-northwest  
188 transect that extended from the lake's profundal zone (18 m) to the inlet (0.6 m; Fig. 2b).  
189 All cores and grab samples were stored at 4°C until they were analyzed at the Indiana  
190 University-Purdue University, Indianapolis (IUPUI) Paleoclimatology and  
191 Sedimentology Laboratory.

192

### 193 *2.2 Initial core description*

194 Surface cores were split, described, imaged and volumetrically subsampled (1 cm<sup>3</sup>)  
195 at 2 cm intervals for % water, dry bulk density and loss on ignition (LOI). Magnetic  
196 susceptibility (MS) was measured on the archive half of the cores using a Bartington  
197 MS2 magnetic susceptibility meter with units SI x10<sup>-5</sup>. Wet 1 cm<sup>3</sup> samples were weighed,  
198 dried for 24 hours at 60°C and reweighed to determine dry bulk density ( $\rho_{\text{dry}}$ ; g cm<sup>-3</sup>) and  
199 % water. Percent total organic matter (% TOM) and carbonate (% TC) were determined  
200 by weight loss after combustion at 550°C (4 hours) and 1000°C (2 hours), respectively  
201 (modified from Heiri et al., 2001).

202

### 203 *2.3 Grain size*

204 Approximately 1.0 g of wet sediment was collected from homogenized grab  
205 samples and at 1 cm intervals from surface cores A-11, dried for 24 hours at 60°C and

206 weighed. Organic matter was removed prior to grain size analysis by soaking samples in  
207 a 50 ml aliquot of 35 % H<sub>2</sub>O<sub>2</sub> at room temperature for 24 hours before being treated  
208 sequentially with five 20 ml aliquots of 35 % H<sub>2</sub>O<sub>2</sub> at 65°C (Gray et al., 2010). Biogenic  
209 silica was removed with a 20 ml 1N NaOH digestion (6 hr at 60°C) and verified by smear  
210 slide for 10 % of the samples. Acid washing was not performed because LOI indicated  
211 that carbonate was not present. After freeze drying, the samples were soaked for 24 hours  
212 in a 2.5 % sodium metaphosphate solution. Particle size distributions of the lithic fraction  
213 were measured using a Malvern Mastersizer 2000 with reported values calculated using  
214 the average of three replicate measurements.

215

#### 216 *2.4 Elemental and isotopic or organic carbon and sedimentary nitrogen*

217 Homogenized subsamples at ~2 cm intervals (n = 41) from core A-11 were  
218 analyzed for total organic carbon (TOC), total nitrogen (TN) concentrations, and δ<sup>13</sup>C  
219 and δ<sup>15</sup>N using a Costech ECS 4010 elemental analyzer interfaced to a Thermo Delta V  
220 Advantage continuous flow isotope ratio mass spectrometer. Isotope ratios are reported in  
221 per mil units (‰) according to the relationship  $\delta X = [(R_{\text{sample}} / R_{\text{standard}}) - 1] \cdot 1000$ , where  
222 X is the element of interest and R a measured isotopic ratio of either a sample or a  
223 standard. Carbon isotope measurements are relative to the Vienna Peedee Belemnite  
224 (VPDB) standard and all nitrogen measurements are relative to atmospheric nitrogen.  
225 Replicate measurements of internal standards run during sample analysis yielded  
226 coefficients of variation of 4.4 %, 6.9 %, for TOC and TN, respectively, and precision  
227 better than 0.2 ‰ for the isotopic data. Isotope measurements were made at Idaho State  
228 University.

229

### 230 *2.5 Biogenic silica and % lithics*

231 Biogenic silica was quantitatively determined on samples at ~ 2 cm resolution (n  
232 = 41) from A-11 using a wet-alkali dissolution method (0.1M Na<sub>2</sub>CO<sub>3</sub> at 85°C for 5 hr)  
233 modified from Mortlock and Froelich (1989) and measured by molybdate-blue  
234 spectrophotometry. Values are reported as 10 % hydrated silica (SiO<sub>2</sub> · 0.4 H<sub>2</sub>O) using a  
235 multiplier of 2.4 times the weight percent of biogenic silica content (Mortlock and  
236 Froelich, 1989). An estimated error of <4.6 % (calculated as the coefficient of variation)  
237 is based on replicate measurements of two internal sediment standards.

238 The lithic abundance was calculated by subtracting percent abundances of  
239 biogenic silica and % TOM from 100 % assuming that carbonate was not a significant  
240 sediment constituent based on LOI data.

241

### 242 *2.6 Scanning X-ray Fluorescence Geochemistry*

243 Sedimentary titanium (Ti) concentrations were determined by scanning X-ray  
244 fluorescence using an ITRAX XRF scanner at the University of Minnesota, Duluth (See  
245 Supplemental Information for concentrations of other elements measured). Measurements  
246 were made at 0.5 cm increments with a molybdenum X-ray tube and a sample exposure  
247 time of 15 seconds per spot. Values are reported as counts per second (cps) where greater  
248 elemental concentrations correspond to higher cps.

249

### 250 *2.7 Geochemistry*

251 Two samples composed of bedrock and glacial till from the adjacent Paru Co  
252 watershed and one of bedrock from the Nir'pa Co watershed were analyzed by ALS  
253 Global for a suite of elements, of which only Ti is discussed here (See supplemental  
254 materials for other elements measured and concentrations). Following ultra-trace level  
255 method ME-MS61, approximately 0.25 g of sample were digested using perchloric, nitric  
256 and hydrofluoric acids with the residue leached with dilute hydrochloric acid and diluted  
257 to volume. Measurements were made by inductively coupled plasma-atomic emission  
258 spectrometry and inductively coupled plasma-mass spectrometry with results corrected  
259 for spectral inter-element interferences (Table S1).

260

#### 261 *2.8 Powder X-ray Diffraction*

262 The mineralogical composition of four samples, three from core A-11 at 7, 45 and  
263 93 cm and 1 of bedrock from Nir'pa Co's watershed, were analyzed using powder X-ray  
264 diffraction. The XRD analyses were conducted using a Bruker D8 Discover instrument,  
265 equipped with an IÁS Cu microsource of 1.54 Angstrom wavelength which provided an  
266 X-ray beam through a 2mm point collimator. The X-ray diffraction pattern was detected  
267 by a Lynxeye 1D detector. The X-ray source's operating conditions were voltage of 50  
268 kV, and current of 1000uA. [Qualitative mineral phase identification was performed using](#)  
269 [Bruker's Diffrac.EVA software and the Crystallographic Open Database \(COD; Table](#)  
270 [S2\).](#)

271

#### 272 *2.9 Sediment age*

273 Surface sediments from core A-11 were dated using  $^{210}\text{Pb}$ ,  $^{214}\text{Pb}$  and  $^{137}\text{Cs}$   
274 measured by direct gamma counting on 10 samples from the upper 9.5 cm (Table 1;  
275 Appleby and Oldfield, 1983; Schelske et al., 1994). Radiocarbon ages from A-11 ( $n = 2$ )  
276 and B -15 ( $n = 2$ ) were determined by accelerator mass spectrometry (AMS  $^{14}\text{C}$ ) at the  
277 University of California, Irvine (UCI), Keck AMS Facility on samples consisting of  
278 charcoaled grass fragments, which were isolated by wet sieving sediment at 63  $\mu\text{m}$   
279 (Table 2). The samples were manually cleaned and chemically pretreated with an acid-  
280 alkali-acid wash of 1N HCl and 1N NaOH following UCI protocols. Radiocarbon ages  
281 were calibrated with the online program CALIB 6.0 (Stuiver and Reimer, 1993) using the  
282 IntCal 09.14c data set. We report the median probability age and  $2\sigma$  error as thousands of  
283 calendar years (ka) before present (B.P.; present = AD 1950).

284 [Insert Table 1]

285 [Insert Table 2]

### 286 **3. Results**

#### 287 *3.1 Lithology*

288 Surface core A-11 is characterized by reddish brown fine-grained sediment that  
289 becomes progressively darker reddish brown with depth. Banding was noted in the upper  
290 30 cm and below 60 cm. Scattered organic fragments, including grass and charcoal are  
291 visible throughout the core. Sediments in surface core A-15 were dark to medium gray  
292 with no apparent banding. In surface core B-15, sediments were gray and coarse grained  
293 with some banding distinguished by zones of increased organic matter from 94 to 74 cm.  
294 Macroscopic charcoal and organic matter were particularly visible from 94 to 81 cm.  
295 From 74 to 65 cm, B-15 sediments transition to a tan color with a darker brown band of

296 mottled sediment between 69 and 65 cm. From 65 to 44 cm, sediments are tan and fine  
297 grained with some banding and still visible, but fewer organic particles. At 44 cm,  
298 macroscopic organic matter particles increased until 5 cm. Above 5 cm, the sediments are  
299 more olive green and the abundance of macroscopic plant fragments decreases.

300

### 301 *3.2 Chronology*

302 Excess  $^{210}\text{Pb}$  ( $^{210}\text{Pb}_{\text{ex}}$ ) activities were highest in core A-11 between 0 and 3.5 cm.  
303 Below this depth,  $^{210}\text{Pb}_{\text{ex}}$  decays until reaching background activities at 7.5 cm. Using the  
304 constant rate of supply model (CRS; Appleby and Oldfield, 1983), an age model was  
305 constructed for the upper most sediment depths between 0 and 7.5 cm (Fig. 3). One  
306 sample at  $4.25 \pm 0.25$  cm contained measurable  $^{137}\text{Cs}$  activities. Based on the  $^{210}\text{Pb}$  CRS  
307 age model, this sample dates to  $1950-60 \pm 4$  CE. Samples above and below this, which  
308 are dated at 1969-77 and 1929-40 CE, respectively, contain no measurable  $^{137}\text{Cs}$ .  
309 Although slightly predating the 1963 CE  $^{137}\text{Cs}$  maximum, the 4.25 cm sample's age is  
310 within error of 1963 CE and consistent with the period during which atmospheric tests of  
311 nuclear weapons were being conducted, supporting the  $^{210}\text{Pb}$  CRS age model. Below 7.5  
312 cm, two AMS  $^{14}\text{C}$  ages of charcoalized grass fragments anchor the core chronology. A  
313 second order polynomial was fit to the combined  $^{210}\text{Pb}$  and AMS  $^{14}\text{C}$  ages to  
314 quantitatively establish the age-depth relationship, which spans the last 3.4 ka.

315 The polynomial age model uncertainty was evaluated by comparison with an age  
316 model derived using the Bchron age modeling software (Blaauw and Christen, 2011: Fig.  
317 S1). The polynomial age model is well within the 2-sigma confidence interval of the  
318 Bchron age model while a regression between the two age models yielded a slope of 1.01

319 and an  $r^2$  of 0.998. The Bchron age model, however, produces abrupt changes in  
320 sedimentation rates at the  $^{14}\text{C}$  age control points that are unlikely to reflect actual  
321 sedimentation rate changes. For this reason, we utilize the polynomial age model, which  
322 estimates more gradual changes in sedimentation rates.

323 The basal AMS  $^{14}\text{C}$  sample for core B-15 at 90 cm returned an age of 11.25 ka  
324 with the 66 cm sample returning an age of 4.5 ka (Table 1).

325 [Insert Figure 3]

326 *3.3 Bulk density, loss on ignition and the elemental and isotopic composition of organic*  
327 *matter carbon and nitrogen*

328 Dry bulk density for core A-11 averaged  $0.28 \text{ g cm}^{-3}$  during the last 3.3 ka with  
329 near average densities from 3.3 to 1.2 ka, above average densities between 1.2 ka and 60  
330 cal yr B.P. and then a marked decrease to the present (Fig. 4a). Mass accumulation rates  
331 averaged  $8.7 \times 10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$  with a maximum of  $15.9 \times 10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$  and a minimum  
332 of  $4.6 \times 10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$  and shows a similar trend with slowly increasing rates from 3.3 to  
333 1.2 ka, a steeper increase between 1.2 ka and 60 cal yr B.P. and a sharp drop off to the  
334 present (Fig. 4b).

335 [Insert Figure 4]

336 Organic matter abundances averaged 13.3 % and are characterized by a  
337 decreasing trend from 3.3 ka to 360 cal yr B.P. that is marked by variations of up to  $\pm 4$   
338 %. After 360 cal yr B.P., % OM increased sharply to the present (Fig. 4c). When  
339 considered as flux, organic matter deposition averaged  $1.1 \times 10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$  with a  
340 maximum of  $1.9 \times 10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$  and a minimum of  $0.6 \times 10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$  and an  
341 increasing trend that is similar to that of mass accumulation (Fig. 4d). Average carbonate

342 content was less than 3 %; variations in % TC most likely represent the loss of interstitial  
343 water from clays.

344 The atomic ratio of carbon to nitrogen (C:N) averaged 11.1 during the last 3.3 ka  
345 with a steady decreasing trend to the present, although with variability ranging between  
346 14.5 and 9.4 (Fig. 4e). Biogenic silica concentrations averaged 10.8 % with values  
347 ranging between 15.0 and 6.7 % (Fig. 4f). A slight decreasing trend is apparent over the  
348 length of the record, but with considerable variability. When plotted as flux, a generally  
349 increasing trend similar to organic matter flux is observed (Fig. 4). Organic matter  $\delta^{13}\text{C}$   
350 ( $\delta^{13}\text{C}_{\text{org}}$ ) was relatively constant during the last 3.3 ka, varying by less than 1 ‰ about a  
351 mean of -24.6 ‰ until 80 cal yr B.P., when  $\delta^{13}\text{C}_{\text{org}}$  decreased to -25.9 ‰ (Fig. 4g). The  
352 isotopic composition of sedimentary nitrogen averaged 2.6 ‰ and ranges from 3.1 to 2.1  
353 ‰ (Fig. 4h). The data suggest a generally increasing trend towards present, with highest  
354 values after 500 cal yr B.P.

355

### 356 *3.4 Lithics and Grain Size*

357 Down core lithic abundances in the Nir'pa Co record varied between 70.5 and  
358 81.0 % with an average of 75.0 % (Fig. 5a). Correlations were positive with Ti ( $r = 0.70$ ;  
359  $p < 0.001$ ) and dry bulk density ( $r = 0.38$ ;  $p = 0.096$ ), null with % BSi and negative with  
360 % TOM ( $r = -0.71$ ;  $p < 0.001$ ) as expected since % lithics was calculated from %TOM.  
361 Still the lithics-TOM correlation is not 1:1, suggesting there is some signal captured. The  
362 relationship between lithics and dry bulk density improves greatly ( $r = 0.64$ ;  $p < 0.001$ ) if  
363 the most recent 110 years are excluded from the regression, which may reflect the  
364 approximate time for flocculent sediment compaction to occur. Between 3.3 and 2.4 ka,



365 % lithics was near average at 74.1 % with low values between 3.3 and 2.9 ka and a peak  
366 between 2.9 and 2.4 ka. At 2.4 ka % lithics decreased to an average of 73.8 % until 1.3  
367 ka, but with variability. After 1.3 ka, % lithics increased to 77.3 %, but again with  
368 variability. Peaks in % lithics occurred at 0.9, 0.4 and 0.1 ka with intervening drops. The  
369 flux of lithic materials shows variations similar to % lithics, but with less variability prior  
370 to 1.3 ka. From 3.3 to 1.3 ka, values generally increased, ranging between 2.2 and 10.1 x  
371  $10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$  about a mean of  $5.7 \times 10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$ . After 1.3 ka, lithic flux increased  
372 more steeply plateauing at an average of  $8.4 \times 10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$  by 950 cal yr B.P. (Fig.  
373 5a). At 110 cal yr B.P., lithic flux began to decrease steeply to the present value of  $6.1 \times$   
374  $10^{-3} \text{ g cm}^{-2} \text{ yr}^{-1}$ .

375 [Insert Figure 5]

376 The grain size distribution of the lithic fraction is dominated by silt, which varies  
377 between 63.3 and 81.1 % and averages 71.5 % (Fig. 5b). Between 3.3 and 2.4 ka, % silt  
378 was 2.4 % below the mean, averaging 69 %, before decreasing by a further 1 % to 68 %  
379 between 2.4 and 1.4 ka (Fig. 5b, d). At 1.4 ka, % silt began a sustain increase to the  
380 present, increasing by 6 % to 73.8 % . Clay is the next most abundant grain size fraction,  
381 accounting for 27.6 % of the lithic component on average with a range between 18.6 and  
382 36.5 % (Fig. 5e). Clay and silt are strongly anti-correlated at a near 1:1 relationship ( $m =$   
383  $-1.01$ ;  $r^2 = 0.88$ ) over the duration of the record. Sand is the least abundant lithic  
384 constituent, averaging just 1.5 %, but ranging of between 0.0 and 5.6 % (Fig. 5c).  
385 Variations in % sand are distinct from those of silt and clay, but with transitions between  
386 higher and lower concentrations coinciding with transitions in silt and clay. Between 3.3  
387 and 2.4 ka and again between 1.4 ka and the present, % sand was generally low,

388 averaging 1.1 and 1.0 %, respectively. From 2.4 to 1.4 ka, % sand increased sharply,  
389 averaging 3 % and reaching a maximum of 5.6 %.

390 Surface grab samples show consistent relationships between grain size, depth and  
391 distance along the sampling transect (Fig. 2). As expected, % sand is highest in littoral  
392 region closest to the ephemeral inlet with peak values of 58 %. Sand abundance  
393 decreased away from the inlet toward the middle of the lake with sand in profundal sites  
394 decreasing from 7 % to less than 1% at the A-11 core site. Silt and clay abundances are  
395 lowest in the littoral zone and increase toward the center of the lake. Within the profundal  
396 zone, % silt and % clay are largely consistent across the transect, but with a subtle  
397 antiphased relationship.

398

### 399 *3.5 Analytical and XRF Ti abundances*

400 Geochemical analysis of the Nir'pa Co and Paru Co bedrock samples returned  
401 high Ti values of 2360 and 2460 ppm, respectively. Ti concentrations in the glacial till  
402 were similarly high, measuring 2290 ppm. Consistent with the sediment source  
403 geochemistry (i.e., bedrock and till samples), XRF derived Ti in the Nir'pa Co sediment  
404 core was high, averaging 7190 cps between a range of 10,510 and 5460 cps (Fig. 5g).  
405 Values of Ti were below average, but with variability, between 3.3 and 1.2 ka, after  
406 which time they increased to the top of the core. Both the long term and higher frequency  
407 Ti variability closely resemble % lithics variations ( $r = 0.70$ ;  $p < 0.001$ ).

408

### 409 *3.6 X-ray diffraction mineralogy*

410 X-ray diffraction patterns from the bedrock sample indicate the presence of  
411 quartz, feldspars, muscovite and titanite, consistent with its mapped andesitic  
412 composition. Accordingly, the down core sediment samples from Nir'pa Co A-11 reflect  
413 this bedrock mineralogy, showing similar assemblages (mainly quartz, feldspar,  
414 muscovite and titanite) with the addition of clay weathering products (Table S1).

415

### 416 *3.7 Magnetic susceptibility*

417 Magnetic susceptibilities for core A-11 averaged 10.1 during the record with a  
418 maximum of 17.5 and a minimum of 2.5 (Fig. 5f). Values were generally low between  
419 3.3 and 1.5 ka with the exception of a broad peak between 2.8 and 2.4 ka. After 1.5 ka,  
420 MS increased, but with superimposed centennial-scale variability at resembles that of the  
421 Si/Ti ratio noted above. For A-15, MS averaged 11.0 with a maximum of 16.1 and a  
422 minimum of 6.0. Trends similar to A-11 are noted in the A-15 MS with generally  
423 increasing values with decreasing depth and peaks from 70 to 75 cm, 45 to 51 cm, 31 and  
424 38 cm, and 10 to 21 cm (supplemental material). Although the magnitude of the peaks in  
425 MS vary somewhat between the two cores, the correspondence is generally good  
426 considering that A-11 was collected from near the middle of the lake and A-15 was  
427 collected closer to the shore where the profundal zone begins to slope up toward the  
428 shore.

429

## 430 **4. Discussion**

431 We focus on results from A-11 because its profundal location is less likely to be  
432 affected by internal lake processes, such as near shore currents, which could affect

433 sediment characteristics (i.e., the location of A-15). Similarities between A-11 and A-15,  
434 however, support generally consistent sedimentation across the lake basin, which give us  
435 confidence that the A-11 results are representative of the lake's hydroclimate variations  
436 during the late Holocene (Fig. S3). Core B-15 is not included in the discussion because  
437 the initial  $^{14}\text{C}$  results indicate that it likely contains hiatuses and is not a complete record  
438 of Holocene climate. Nonetheless, the basal age of 11.25 ka is consistent with the basal  
439 age from nearby Paru Co, suggesting that lakes in the region deglaciated and filled  
440 rapidly at this time.

441

#### 442 *4.1 Lithic and grain size proxy interpretation*

443 Grain size measurements of lacustrine sediments offer insight into watershed and  
444 in lake processes that are sensitive to climatic variations (e.g., Shuman et al., 2009;  
445 Dearing, 1997; Conroy et al., 2008). Today at Nir'pa Co, below freezing temperatures  
446 and minimal precipitation between November and April limit watershed erosion and the  
447 transport of clastic material to the lake during the cold season (Fig. 2c). At the same time,  
448 cold season ice cover precludes sediment influx to Nir'pa Co while also limiting  
449 autochthonous biological productivity. As such, sediment deposition at the lake is largely  
450 restricted to the warm summer months when clastic material can be mobilized by  
451 monsoonal precipitation after above freezing temperatures removes snow cover from the  
452 watershed and ice cover from the lake. Watershed and in-lake productivity also increase  
453 during this time, although the degree of productivity is controlled by multiple factors,  
454 including average temperature, precipitation, pH, nutrient availability and turbidity  
455 (Carpenter et al., 1985; Catalan et al., 2009).

456           Given the highly seasonal climate at Nir'pa Co and the regional dominance of the  
457 ISM on the summer hydrologic regime, we interpret Nir'pa Co's clastic sediment record  
458 as reflecting watershed and depositional processes that are linked to summer climatic  
459 conditions associated with the ISM. The lithic fraction, which averages more than half of  
460 the sediment total, is by far the dominant component of Nir'pa Co's sediment archive  
461 record. Of this component, silt is the primary grain size fraction, averaging 72.9 %  
462 followed by % clay and % sand. Modern grain size distributions show that silt and clay  
463 are relatively evenly distributed across Nir'pa Co's profundal zone, although with an  
464 antiphased relationship (Fig. 2). Increases in one constituent, therefore, necessarily lead  
465 to the reduction of the other. We suggest that the down core antiphased relationship  
466 between % silt and % clay reflects variations in runoff intensity, such that silt (clay)  
467 content would increase (decrease) when runoff intensity was enhanced (diminished), and  
468 vice versa. This is supported by the similarities between % silt and % lithics ( $r = 0.51$ ,  $p <$   
469  $0.001$ ), which are interpreted as the relative contribution of clastic material delivered to  
470 the lake (Sly, 1978). We suggest % lithics is primarily controlled by summer  
471 precipitation whereby enhanced rainfall results in greater lithic mobilization in the  
472 watershed. At the same time, % silt abundances reflect increased runoff energy during  
473 times of greater lithic influx.

474           Down core variations in % sand are distinct from those in lithics, silt and clay, but  
475 with transitions that were contemporaneous with changes in silt, clay and lithics (Fig. 5).  
476 This suggests that different, but still climatically related processes resulted in changes in  
477 the proportion of sand deposited at the core site. Today, sand abundance is clearly linked  
478 to lake levels with minimum sand in profundal sites and increasing abundances toward

479 the littoral and inlet regions. Because much of Nir'pa Co's bathymetry is steep sided with  
480 a flat bottom, changes in lake level will not greatly increase the proximity of the littoral  
481 zone for most of the lake's basin (Håkanson, 1982). At the lake's shallow north end,  
482 however, small changes in lake level will more significantly reduce the distance between  
483 the core site and the point at which seasonal fluvial discharge enters the lake. For  
484 example, a lake level decrease of ~1.5 m would increase the proximity of fluvial inputs  
485 and the core site by approximately 130 m (Fig. 2b). Although encroachment of the littoral  
486 zone at other points around the lake basin might also be important, the net result of  
487 increased sand with lower lake levels would remain consistent. We therefore interpret %  
488 sand as a lake level indicator such that increased sandiness reflects lower lake levels with  
489 more proximal fluvial discharge and reduced sandiness reflects higher lake levels and  
490 more distal fluvial discharge relative to the core site. That sand increased during times  
491 when % silt and lithics decreased, is consistent with our interpretation because if  
492 increased sandiness were related to depositional energy, we would expect it to increase in  
493 phase with % silt and lithics, not decrease as is observed.

494

#### 495 *4.2 Magnetic susceptibility and Si/Ti*

496 The magnetic susceptibility and Ti abundance in lacustrine sediments have been  
497 widely used as indicators of the contribution of terrestrial clastic material (Seltzer et al.,  
498 2002; Metcalfe et al., 2010). At Nir'pa Co, MS and Ti closely track changes in % lithics  
499 and silt, although with important differences, particularly after 1.3 ka, generally  
500 supporting the hypothesis of terrestrial erosion control. Subtle differences between Ti and

501 MS, however, indicate that processes other than terrestrial input may be reflected in the  
502 MS profile.

503 [Insert Figure 6]

#### 504 *4.3 Late Holocene ISM rainfall at Nir'pa Co*

505 Based on the sedimentological characteristics of Nir'pa Co's clastic sediment  
506 fraction, we identified three distinct multi-century-scale phases of hydroclimate  
507 variability during the last 3.3 ka within a larger trend of generally increasing precipitation  
508 (Fig. 6). Two periods of elevated lithic and silt abundance with low clay and sand content  
509 occurred between 3.3 and 2.4 ka and 1.3 ka and the present, suggesting wetter conditions  
510 and a strengthened ISM. Although MS does not show a strong increase during the earliest  
511 inferred wet period between 3.3 and 2.4 ka, it does increase significantly after 1.3 ka in  
512 step with lithics, % silt and Ti, indicating that the increase in ISM precipitation was large  
513 enough to register in all proxies. The intervening period from 2.4 to 1.3 ka, with its lower  
514 silt and lithic content, but elevated sand and clay content, suggests drier conditions with  
515 lower lake levels.

516 Superimposed on these millennial-scale ISM variations were higher frequency  
517 century-scale ISM variations; however, these are comparatively minor when considered  
518 within the context of the lower frequency variations (Fig 6). In particular, it is notable  
519 that relatively minor sedimentological responses are apparent at Nir'pa Co during the  
520 MCA (950 to 800 cal yr B.P.). Although subtle and not statistically significant (two-  
521 sample t-test null hypothesis not rejected), decreasing sand and elevated silt and lithics  
522 during the MCA may possibly reflect an increase in ISM rainfall that resulted in higher  
523 lake levels and greater watershed erosion during this event (Fig. 6a-c). Increasing % sand

524 and slightly reduced lithics and % silt during the LIA from 500 to 200 cal yr B.P. may  
525 suggest a drier LIA at Nir'pa Co. It is important to stress, however, that sedimentological  
526 responses during the MCA and LIA were minor compared to the millennial scale increase  
527 in ISM precipitation on which they are superimposed and that additional records with  
528 greater temporal resolution and climatic sensitivity are needed to more robustly  
529 investigate MCA and LIA hydroclimate expressions in this region.

530

#### 531 *4.4 Late Holocene productivity at Nir'pa Co*

532 Paleo productivity at Nir'pa Co is interpreted from several indicators, including %  
533 OM, OM flux, the elemental abundance and isotopic composition of organic carbon and  
534 sedimentary nitrogen, and biogenic silica.

535 The decreasing trend in % OM at Nir'pa Co from 3.3 ka to 360 cal yr B.P., is  
536 attributed to dilution from increasing clastic sedimentation as indicated by greater lithic  
537 composition and mass accumulation rates, particularly after 1.3 ka (Fig. 4 & 5). When  
538 considered as flux, OM shows an increasing trend during the late Holocene that is anti-  
539 phased with decreasing C:N, implicating enhanced aquatic productivity as the cause of  
540 this trend (Fig. 5). Biogenic silica also shows an increasing trend when considered as  
541 flux, additionally supporting the idea that aquatic productivity increased through the late  
542 Holocene. At the same time,  $\delta^{13}\text{C}_{\text{org}}$  varies by less than 1 ‰, with the exception of one  
543 surface sample that was markedly more negative than the late Holocene average,  
544 suggesting that aquatic productivity was the source of sedimentary organic matter and  
545 that only its production intensity changed through time.

546



547 *4.5 Late Holocene ISM variability on the Third Pole*

548 Despite generally more arid late Holocene conditions (Gasse et al., 1996; Bird et  
549 al., 2014; Cai et al., 2012; Zhu et al., 2008; Mügler et al., 2010), comparison of Nir'pa Co  
550 results with other Third Pole paleoclimate records indicates periods of coherent  
551 hydroclimatic variability on the central and eastern Tibetan Plateau (Fig. 6). At nearby  
552 Paru Co, for example, reconstructed lake levels were higher between 3.3 and 2.5 ka and  
553 1.3 ka and the present, with an intervening period of declining lake level between 2.4 and  
554 1.3 ka (Fig 6d). Erosion at Paru Co was also significantly enhanced after 1.3 ka,  
555 consistent with trends at Nir'pa Co (Fig. 6e). Although low resolution, the hydrogen  
556 isotopic composition of leaf wax n-alkanes ( $\delta D_{\text{wax}}$ ) at Paru Co broadly suggests similar  
557 late Holocene hydrologic variability with lower  $\delta D_{\text{wax}}$  corresponding to periods of  
558 increased precipitation, erosion and high lake levels and higher  $\delta D_{\text{wax}}$  corresponding to  
559 periods of reduced ISM precipitation, reduced erosion and lower lake levels at Paru Co  
560 and Nir'pa Co (Fig. 6f).

561 At Seling Co, the largest lake in Tibet and located approximately 370 km  
562 northwest of Nir'pa Co, late Holocene  $\delta^{18}\text{O}_{\text{cal}}$  values were relatively low from 3.3 to 2.4  
563 ka and after 1.4 ka, suggesting wetter conditions during these times (Fig. 6f; Gu et al.,  
564 1993). Between 2.4 and 1.4 ka, higher  $\delta^{18}\text{O}_{\text{cal}}$  at Seling Co suggests enhanced aridity.  
565 Similar trends are noted in grain size, mineralogical and geochemical proxies at Nam Co,  
566 the second largest lake on the Tibetan Plateau, which is approximately 180 km northwest  
567 of Nir'pa Co (Mügler et al., 2010; Zhu et al., 2008). A generally stable environment is  
568 suggested for Nam Co between 4 and 1.6 ka with a marked period of aridity between 1.6  
569 and 1.4 ka. After 1.4 ka, wetter conditions persisted until 800 cal yr B.P., after which

570 point aridity was again inferred. Considered as a whole, the similar trends at Seling Co  
571 and Nam, Nir'pa Co and Paru Co, particularly since 1.4 ka, suggest regionally similar  
572 multi-centennial to millennial-scale hydroclimate on the southeastern and central Tibetan  
573 Plateau.

574 The hydroclimatic variations inferred from Tibetan lake sediment records are  
575 generally similar with inferred changes in surface air temperatures as reconstructed with a  
576 pollen biome index technique applied to results from Lake Zigetang (Fig. 6i; Herzschuh  
577 et al., 2006). Elevated air temperatures are suggested to have persisted between 3.3 and  
578 2.3 ka and after 1.4 ka with cooler air temperatures between 2.3 and 1.4 ka. This suggests  
579 that increased ISM precipitation between 3.3 and 2.4 ka and 1.3 ka and the present was  
580 broadly associated with warmth on the Tibetan Plateau whereas reduced ISM rainfall  
581 between 2.4 and 1.3 ka coincided with cooler air temperatures. This pattern was also  
582 noted for other lake high stands and precipitation increases during the middle and early  
583 Holocene at Paru Co (Bird et al., 2014).

584 In contrast with the largely similar expression of ISM variability on central and  
585 eastern Tibetan Plateau, the Bangong Lake  $\delta^{18}\text{O}$  record from western Tibet shows  
586 different trends during the last 3 ka (Fig. 6h; Gasse et al., 1996). Increasing  $\delta^{18}\text{O}$  after 3.3  
587 ka is interrupted at 2.4 ka by a period of steady to lower  $\delta^{18}\text{O}$  that lasted until 1.3 ka.  
588 After 1.3 ka,  $\delta^{18}\text{O}$  at Bangong Lake increased abruptly to maximum late Holocene values  
589 that then decrease slowly. It is possible that the Bangong Lake  $\delta^{18}\text{O}$  data reflects regional  
590 changes in effective moisture, with the last 1.3 ka being a period of considerably higher  
591 evaporation and reduced ISM precipitation that did not penetrate into western Tibet. This  
592 idea is consistent with recent work that showed lakes levels in western Tibet were more

593 sensitive to long-term reductions in ISM precipitation than those in the east, which are  
594 closer to the source of ISM moisture (Hudson and Quade, 2013). It is also possible that  
595 moisture from the mid-latitude westerlies, which has elevated  $\delta^{18}\text{O}$  relative to monsoonal  
596 precipitation (Tian et al., 2007), increased during this time as the ISM weakened.

597

#### 598 *4.6 Regional late Holocene ISM expressions*

599 The ISM trends discussed above are also expressed in proxy records more distal  
600 to the Tibetan Plateau, but still within the sphere of ISM influence. A foraminifera-based  
601 upwelling record from the Arabian Sea, for example, shows enhanced upwelling,  
602 suggesting a strengthened ISM during the two pluvial phases identified at Nir'pa Co  
603 between 3.3 and 2.4 ka and 1.3 ka to the present with the latter stronger than the former  
604 (Fig. 7e; Gupta et al., 2005; Overpeck et al., 1996). Between 2.4 and 1.5 ka, upwelling is  
605 reduced relative to the preceding and following periods. The Oman speleothem  $\delta^{18}\text{O}$   
606 record, which has been interpreted as reflecting the northerly (wet) or southerly (dry)  
607 position of the Intertropical Convergence Zone (ITCZ), shows relatively little  $\delta^{18}\text{O}$   
608 variability, but contains a hiatus between 2.7 and 1.4 ka (Fig. 7f). This may suggest that  
609 reduced precipitation during this time, possibly reflecting a southerly mean position of  
610 the ITCZ, precluded drip water percolation and calcite formation (Fleitmann et al., 2007).

611

[Insert Figure 7]

612 Although temporally limited, trends in a 900-yr-long  $\delta^{18}\text{O}$  record from Dandak  
613 Cave, India, show some similarities with the Third Pole records (Fig. 6j; Sinha et al.,  
614 2006). The initiation of calcite precipitation at Dandak Cave at approximately 1.3 ka  
615 coincides with the end of the arid interval at Nir'pa Co between 2.4 and 1.3 ka. This may

616 indicate that a shift to increased precipitation at Dandak Cave at time permitted calcite to  
617 form. Generally low  $\delta^{18}\text{O}$  between 1020 and 690 cal yr B.P., suggests an interval of  
618 enhanced ISM precipitation that coincides with the MCA. This is consistent with another  
619 speleothem from Sahiya Cave (30° 36'N, 77° 52'E, 1190 m asl) that shows low  $\delta^{18}\text{O}$   
620 during the MCA between 1050 and 650 cal yr B.P., indicating enhanced ISM  
621 precipitation and moisture delivery from the Bay of Bengal (Sinha et al., 2015). This  
622 supports the tentative evidence from Nir'pa Co that the MCA may have been a period of  
623 slightly greater ISM precipitation on the central/eastern Tibetan Plateau. Increasing  $\delta^{18}\text{O}$   
624 after 650 and 690 cal yr B.P, in the Dandak and Sahiya Cave records, respectively,  
625 although with variability, and the cessation of calcite precipitation at 390 cal yr B.P. at  
626 Dandack cave, suggest a shift to increasing aridity during the LIA that coincides with  
627 potentially drier conditions at Nir'pa Co. It is important to stress, however, that the MCA  
628 and LIA signals in the Nir'pa Co record are not statistically significant, which suggests  
629 that these events were not strongly expressed in at least the southeastern portion of the  
630 Tibetan Plateau. Additional records are needed from this and other regions across Tibet in  
631 order to better evaluate the expression of these events and whether or not they were  
632 indeed climatically significant intervals.

633

#### 634 *4.7 Mechanisms of late Holocene ISM precipitation trends*

635 The largely coherent multi-centennial ISM trends noted in lake sediment records  
636 from the central and eastern Tibetan Plateau suggest common, large-scale forcing  
637 mechanisms. Although some work suggests that the modern ENSO-ISM relationship was  
638 only established during the last 2000 years, previous work investigating Holocene ISM

639 on the Third Pole showed that it closely followed Indo-Pacific sea surface temperatures  
640 (SST) and ENSO-like ocean-atmosphere variability on centennial and millennial scales  
641 through the Holocene (Bird et al., 2014). Comparison of the Nir'pa Co results and other  
642 Third Pole records shows that periods of inferred wet conditions between 3.3 and 2.4 ka  
643 and from 1.3 ka and the present similarly coincided with Indo-Pacific warming and  
644 eastern equatorial Pacific cooling suggestive of La Niña-like conditions (Fig. 7h & i;  
645 Koutavas et al., 2006; Marchitto et al., 2010). Although SST reconstructions spanning the  
646 last 2 ka are not available from the EEP, an El Niño record from El Junco, a crater lake in  
647 the Galapagos Islands, provides supporting evidence for late Holocene ENSO variability  
648 that is consistent with SST records from the Indo-Pacific (Fig. 7g; Conroy et al., 2008).  
649 Generally low % sand in the El Junco record from 3.3 to 2.4 ka and 1.3 ka to the present  
650 is interpreted to reflect reduced precipitation over the lake's watershed, which occurs  
651 today during La Niña conditions. Persistent La Niña-like conditions from 3.3 to 2.4 ka  
652 and 1.3 ka to the present is consistent with the Nir'pa Co results and other Third Pole  
653 records, which suggest wetter and warmer conditions during these times. Peak % sand at  
654 El Junco between 2.4 and 1.3 ka, suggesting persistent El Niño-like conditions, coincides  
655 with lake level decreases at Nir'pa Co and Paru Co, aridity at Seling Co and Nam Co and  
656 cooling on the Tibetan Plateau. The suggested occurrence of El Niño-like conditions  
657 between 2.4 and 1.3 ka is consistent with Indo-Pacific cooling and the modern  
658 relationship between El Niño, reduced ISM precipitation and cooling as a result of  
659 enhanced Tibetan Plateau snow pack (Shaman and Tziperman, 2005). Similar ENSO  
660 patterns are observed at Laguna Pallcacocha, Ecuador, with the exception of the period  
661 between 2.4 and 1.7 ka, when fewer than 5 ENSO events per century were observed and

662 the El Junco record suggests enhanced El Niño-like conditions (Moy et al., 2002). The El  
663 Junco record, however, may better reflect EEP ENSO-like variability because its location  
664 in the Galapagos Islands is more directly linked to Pacific SST and convective activity  
665 (Conroy et al., 2008).

666 The relatively minor expressions of the MCA and LIA in the higher resolution  
667 record from Nir'pa Co suggests that lower-frequency climate processes, particularly  
668 those acting in the Indo-Pacific, may have been more fundamental in determining the  
669 mean state of the late Holocene ISM. Still, that the ISM appears to have strengthened  
670 during the MCA and weakened during the LIA is consistent with evidence for La Niña-  
671 like and El Niño-like conditions during these events, respectively (Mann et al., 2009).

672

## 673 **5. Summary and Conclusions**

674 The Nir'pa Co sediment record provides a detailed view of late Holocene ISM  
675 variability from the southeastern Tibetan Plateau spanning the last 3.3 ka. Within the  
676 context a generally weakened late Holocene ISM, two pluvial phases are identified. The  
677 strongest pluvial phase, which is recognized in other records from the central and eastern  
678 Tibetan Plateau occurred between 1.3 ka and the present. A second, less intense, pluvial  
679 phase occurred between 3.3 and 2.4 ka.

680 Between these pluvial phases, a pronounced arid interval occurred between 2.4  
681 and 1.3 ka. Other paleoclimate records from the Third Pole also indicate aridity during  
682 this time, suggesting that this was a widespread event. The two pluvial phases occurred in  
683 conjunction with warmer surface air temperatures on the Tibetan Plateau, elevated SST in  
684 the Indo-Pacific and lower SST in the EEP. These conditions are consistent with a La

685 Niña-like mean state in the tropical Pacific, suggesting that Indo-Pacific ocean-  
686 atmosphere variability contributed to these events. This idea is supported by a lake  
687 sediment record from El Junco in the ENSO-sensitive Galapagos Islands, which suggests  
688 La Niña-like conditions occurred during inferred pluvial phases at Nir'pa Co.  
689 Conversely, aridity at Nir'pa Co between 2.4 and 1.3 ka occurred in conjunction with  
690 reduced Indo-Pacific SST and cooling on the Tibetan Plateau, which are suggestive of  
691 enhanced El Niño-like conditions. The El Junco record again supports this idea, showing  
692 the greatest increase in % sand during the late Holocene, which has been interpreted as  
693 reflecting enhanced El Niño conditions. These associations between variations in ISM  
694 precipitation, Tibetan Plateau surface air temperatures and Pacific ocean atmosphere  
695 variability are consistent with modern relationships observed in the instrumental record  
696 whereby El Niño/El Niño-like conditions produce reductions in ISM precipitation and La  
697 Niña/La Niña-like conditions enhance the ISM.

698         Although the above ISM phases are noted in multiple records from the Third Pole,  
699 the Bangong Co record from western Tibet shows opposite trends during the late  
700 Holocene. This suggests that ISM variability during the last 3 ka differed across the  
701 Tibetan Plateau. This may reflect the inability of the ISM to exert a strong influence  
702 across the Tibetan Plateau as a result of its generally weakened mean state during the late  
703 Holocene. As a result, evaporation in western Tibet may have increased with little  
704 moderation by the two pluvial events noted in the central and eastern Tibetan Plateau.  
705 This is consistent with an east-west asymmetry in Holocene lake level trends that show  
706 greater evaporation sensitivity in western Tibet. The last 1000 years may also have

707 experienced increased moisture from westerly sources as a result of strengthened  
708 westerlies that delivered isotopically lighter precipitation.

709         Superimposed on the millennial-scale ISM variations noted above were minor  
710 higher frequency, but lower amplitude changes in ISM precipitation. Although subtle and  
711 not statistically significant, possible increases in ISM rainfall and lake levels occurred  
712 during the MCA while reductions in ISM rainfall and lower lake levels are suggested for  
713 the LIA. These hydroclimate expressions of the MCA and LIA are consistent with  
714 regional evidence from speleothems and recent work combining proxy records and  
715 climate model simulations suggesting a La Niña-like mean state in Pacific ocean-  
716 atmosphere system during the MCA and an El Niño-like mean state during the LIA.  
717 Additional records with improved spatial and temporal resolution are needed, however, to  
718 better assess the apparently subtle expressions of the MCA and LIA in the ISM region.

719         The patterns and mechanisms of late Holocene ISM variability identified in the  
720 Nir'pa Co and other Third Pole records are consistent with those identified for the early  
721 and middle Holocene (e.g. Bird et al., 2014). This suggests that the tropical Pacific's  
722 response to warming (Clement et al., 1996; Marchitto et al., 2010; England et al., 2014)  
723 to warming may continue to enhance ISM precipitation on the Tibetan Plateau, as has  
724 been apparent from the expansion of lakes across the plateau, but particularly in the  
725 central and eastern sections over the last few decades (Lei et al., 2013). Additional and  
726 higher-resolution records are needed from the Third Pole region, however, in order to  
727 confirm our findings, evaluate their spatiotemporal coherence, and investigate additional  
728 controls on multi-decadal ISM rainfall through the Holocene (e.g., solar and Atlantic  
729 Ocean-atmosphere variability).



730

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737

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742

743 **Data**

744 All original data presented in this paper are publically available via the NOAA  
745 Paleoclimate Database (a url to the data will be included in the final manuscript pending  
746 acceptance).

747

748 **References**

- 749 Appleby PG and Oldfield F. (1983) The assessment of  $^{210}\text{Pb}$  data from sites with varying  
750 sedimentation accumulation rates. *Hydrobiologia* 103: 29-35.  
751 Ashok K, Guan Z, Saji N, et al. (2004) Individual and combined influences of ENSO and  
752 the Indian Ocean dipole on the Indian summer monsoon. *Journal of Climate* 17:  
753 3141-3155.  
754 Ashok K, Guan Z and Yamagata T. (2001) Impact of the Indian Ocean Dipole on the  
755 relationship between the Indian monsoon rainfall and ENSO. *Geophysical*  
756 *Research Letters* 28: 4499-4502.

757 Berkelhammer M, Sinha A, Stott L, et al. (2012) An abrupt shift in the Indian monsoon  
758 4000 years ago. *Climates, Landscapes, and Civilizations*. Washington, DC: AGU,  
759 75-87.

760 Bird BW, Abbott MB, Vuille M, et al. (2011) A 2300-year-long annually resolved record  
761 of the South American summer monsoon from the Peruvian Andes. *Proceedings*  
762 *of the National Academy of Sciences* 108: 8583-8588.

763 Bird BW, Polisar PJ, Lei Y, et al. (2014) A Tibetan lake sediment record of Holocene  
764 Indian summer monsoon variability. *Earth and Planetary Science Letters* 399: 92-  
765 102.

766 Blaauw M and Christen JA. (2011) Flexible paleoclimate age-depth models using an  
767 autoregressive gamma process. *Bayesian Analysis* 6: 457-474.

768 Cai Y, Zhang H, Cheng H, et al. (2012) The Holocene Indian monsoon variability over  
769 the southern Tibetan Plateau and its teleconnections. *Earth and Planetary Science*  
770 *Letters* 335-336: 135-144.

771 Carpenter SR, Kitchell JF and Hodgson JR. (1985) Cascading trophic interactions and  
772 lake productivity. *BioScience*: 634-639.

773 Catalan J, Pla S, García J, et al. (2009) Climate and CO<sub>2</sub> saturation in an alpine lake  
774 throughout the Holocene. *Limnology and Oceanography* 54: 2542.

775 Chen J, Chen F, Feng S, et al. (2015) Hydroclimatic changes in China and surroundings  
776 during the Medieval Climate Anomaly and Little Ice Age: spatial patterns and  
777 possible mechanisms. *Quaternary Science Reviews* 107: 98-111.

778 Choubert G, Faure-Muret A and Chanteux P. (1983) *Atlas géologique du monde*: Unesco.

779 Clement AC, Seager R, Cane MA, et al. (1996) An ocean dynamical thermostat. *Journal*  
780 *of Climate* 9: 2190-2196.

781 Conroy JL, Overpeck JT, Cole JE, et al. (2008) Holocene changes in eastern tropical  
782 Pacific climate inferred from a Galápagos lake sediment record. *Quaternary*  
783 *Science Reviews* 27: 1166-1180.

784 Dearing J. (1997) Sedimentary indicators of lake-level changes in the humid temperate  
785 zone: a critical review. *Journal of Paleolimnology* 18: 1-14.

786 England MH, McGregor S, Spence P, et al. (2014) Recent intensification of wind-driven  
787 circulation in the Pacific and the ongoing warming hiatus. *Nature Climate Change*  
788 4: 222-227.

789 Fleitmann D, Burns SJ, Mangini A, et al. (2007) Holocene ITCZ and Indian monsoon  
790 dynamics recorded in stalagmites from Oman and Yemen (Socotra). *Quaternary*  
791 *Science Reviews* 26: 170-188.

792 Gadgil S. (2003) The Indian monsoon and its variability. *Annual Review of Earth and*  
793 *Planetary Sciences* 31: 429-467.

794 Gasse F, Fontes JC, Van Campo E, et al. (1996) Holocene environmental changes in  
795 Bangong Co basin (Western Tibet). Part 4: discussion and conclusions.  
796 *Palaeogeography, Palaeoclimatology, Palaeoecology* 120: 79-92.

797 Goswami BN, Madhusoodanan MS, Neema CP, et al. (2006) A physical mechanism for  
798 North Atlantic SST influence on the Indian summer monsoon. *Geophysical*  
799 *Research Letters* 33: L02706.

800 Gray AB, Pasternack GB and Watson EB. (2010) Hydrogen peroxide treatment effects on  
801 the particle size distribution of alluvial and marsh sediments. *The Holocene* 20:  
802 293-301.

803 Gu Z, Liu J, Yuan B, et al. (1993) The changes in monsoon influence in the Qinghai-  
804 Tibetan Plateau during the past 12,000 years. Geochemical evidence from the L.  
805 Selin sediments. *Chinese Science Bulletin* 38: 61-64.

806 Gupta AK, Das M and Anderson DM. (2005) Solar influence on the Indian summer  
807 monsoon during the Holocene. *Geophysical Research Letters* 32: L17703.

808 Håkanson L. (1982) Lake bottom dynamics and morphometry: The dynamic ratio. *Water*  
809 *Resources Research* 18: 1444-1450.

810 He Y, Theakstone WH, Zhonglin Z, et al. (2004) Asynchronous Holocene climatic  
811 change across China. *Quaternary Research* 61: 52-63.

812 Heiri O, Lotter AF and Lemcke G. (2001) Loss on ignition as a method for estimating  
813 organic and carbonate content in sediments: reproducibility and comparability of  
814 results. *Journal of Paleolimnology* 25: 101-110.

815 Herzschuh U, Winter K, Wünnemann B, et al. (2006) A general cooling trend on the  
816 central Tibetan Plateau throughout the Holocene recorded by the Lake Zigetang  
817 pollen spectra. *Quaternary International* 154-155: 113-121.

818 Hong YT, Hong B, Lin QH, et al. (2005) Inverse phase oscillations between the East  
819 Asian and Indian Ocean summer monsoons during the last 12000 years and paleo-  
820 El Niño. *Earth and Planetary Science Letters* 231: 337-346.

821 Hong YT, Hong B, Lin QH, et al. (2003) Correlation between Indian Ocean summer  
822 monsoon and North Atlantic climate during the Holocene. *Earth and Planetary*  
823 *Science Letters* 211: 371-380.

824 Hudson AM and Quade J. (2013) Long-term east-west asymmetry in monsoon rainfall on  
825 the Tibetan Plateau. *Geology* 41: 351-354.

826 Koutavas A, Olive GC and Lynch-Stieglitz J. (2006) Mid-Holocene El Niño–Southern  
827 Oscillation (ENSO) attenuation revealed by individual foraminifera in eastern  
828 tropical Pacific sediments. *Geology* 34: 993-996.

829 Krishnamurthy V and Kirtman BP. (2009) Relation between Indian monsoon variability  
830 and SST. *Journal of Climate* 22: 4437-4458.

831 Krishnan R and Sugi M. (2003) Pacific decadal oscillation and variability of the Indian  
832 summer monsoon rainfall. *Climate Dynamics* 21: 233-242.

833 Kumar KK, Rajagopalan B, Hoerling M, et al. (2006) Unraveling the mystery of Indian  
834 monsoon failure during El Niño. *Science* 314: 115-119.

835 Lawrimore JH, Menne MJ, Gleason BE, et al. (2011) An overview of the Global  
836 Historical Climatology Network monthly mean temperature data set, version 3.  
837 *Journal of Geophysical Research* 116: D19121.

838 Lei Y, Yao T, Bird BW, et al. (2013) Coherent lake growth on the central Tibetan Plateau  
839 since the 1970s: Characterization and attribution. *Journal of Hydrology* 483: 61-  
840 67.

841 Liu X, Dong H, Yang X, et al. (2009) Late Holocene forcing of the Asian winter and  
842 summer monsoon as evidenced by proxy records from the northern Qinghai–  
843 Tibetan Plateau. *Earth and Planetary Science Letters* 280: 276-284.

844 Mann ME, Zhang Z, Rutherford S, et al. (2009) Global signatures and dynamical origins  
845 of the Little Ice Age and Medieval Climate Anomaly. *Science* 326: 1256-1260.

846 Marchitto TM, Muscheler R, Ortiz JD, et al. (2010) Dynamical response of the tropical  
847 Pacific Ocean to solar forcing during the early Holocene. *Science* 330: 1378-  
848 1381.

849 Metcalfe SE, Jones MD, Davies SJ, et al. (2010) Climate variability over the last two  
850 millennia in the North American Monsoon region, recorded in laminated lake  
851 sediments from Laguna de Juanacatlán, Mexico. *The Holocene*.

852 Mischke S and Zhang C. (2010) Holocene cold events on the Tibetan Plateau. *Global and*  
853 *Planetary Change* 72: 155-163.

854 Morrill C, Overpeck JT and Cole JE. (2003) A synthesis of abrupt changes in the Asian  
855 summer monsoon since the last deglaciation. *The Holocene* 13: 465-476.

856 Morrill C, Overpeck JT, Cole JE, et al. (2006) Holocene variations in the Asian monsoon  
857 inferred from the geochemistry of lake sediments in central Tibet. *Quaternary*  
858 *Research* 65: 232-243.

859 Mortlock RA and Froelich PN. (1989) A simple method for the rapid determination of  
860 biogenic opal in pelagic marine sediments. *Deep Sea Research Part A.*  
861 *Oceanographic Research Papers* 36: 1415-1426.

862 Moy CM, Seltzer GO, Rodbell DT, et al. (2002) Variability of El Niño/Southern  
863 Oscillation at millennial timescales during the Holocene epoch. *Nature* 420: 162-  
864 165.

865 Mügler I, Gleixner G, Günther F, et al. (2010) A multi-proxy approach to reconstruct  
866 hydrological changes and Holocene climate development of Nam Co, Central  
867 Tibet. *Journal of Paleolimnology* 43: 625-648.

868 Overpeck J, Anderson D, Trumbore S, et al. (1996) The southwest Indian Monsoon over  
869 the last 18000 years. *Climate Dynamics* 12: 213-225.

870 Peings Y and Douville H. (2010) Influence of the Eurasian snow cover on the Indian  
871 summer monsoon variability in observed climatologies and CMIP3 simulations.  
872 *Climate Dynamics* 34: 643-660.

873 Peterson TC and Vose RS. (1997) An overview of the Global Historical Climatology  
874 Network temperature database. *Bulletin of the American Meteorological Society*  
875 78: 2837-2849.

876 Qiu J. (2008) The Third Pole. *Nature* 454: 393-396.

877 Schelske CL, Peplow A, Brenner M, et al. (1994) Low-background gamma counting:  
878 applications for <sup>210</sup>Pb dating of sediments *Journal of Paleolimnology* 10: 115-128.

879 Seltzer GO, Rodbell DT, Baker PA, et al. (2002) Early Warming of Tropical South  
880 America at the Last Glacial-Interglacial Transition. *Science* 296: 1685-1686.

881 Shaman J and Tziperman E. (2005) The effect of ENSO on Tibetan Plateau snow depth:  
882 A stationary wave teleconnection mechanism and implications for the South  
883 Asian monsoons. *Journal of Climate* 18: 2067-2079.

884 Shuman B, Henderson AK, Colman SM, et al. (2009) Holocene lake-level trends in the  
885 Rocky Mountains, USA. *Quaternary Science Reviews* 28: 1861-1879.

886 Sinha A, Cannariato KG, Stott LD, et al. (2006) A 900-year (600 to 1500 A.D.) record of  
887 the Indian summer monsoon precipitation from the core of the monsoon zone of  
888 India. *Geophysical Research Letters* 34: L16707.

889 Sinha A, Kathayat G, Cheng H, et al. (2015) Trends and oscillations in the Indian  
890 summer monsoon rainfall over the last two millennia. *Nature communications* 6.

891 Sly P. (1978) Sedimentary processes in lakes. *Lakes*. Springer, 65-89.

892 Stott L, Cannariato K, Thunnell R, et al. (2004) Decline of surface temperature and  
893 salinity in the western tropical Pacific Ocean in the Holocene epoch. *Nature* 431:  
894 56-59.

895 Stuiver M and Reimer PJ. (1993) Extended  $^{14}\text{C}$  database and revised CALIB radiocarbon  
896 calibration program. *Radiocarbon* 35: 215-230.  
897 Thompson LG, Yao T, Mosley-Thompson E, et al. (2000) A high-resolution millennial  
898 record of the South Asian monsoon from Himalayan ice cores. *Science* 289: 1916-  
899 1919.  
900 Tian L, Masson-Delmotte V, Stievenard M, et al. (2001) Tibetan Plateau summer  
901 monsoon northward extent revealed by measurements of water stable isotopes.  
902 *Journal of Geophysical Research* 106: 28081-28088.  
903 Tian L, Yao T, MacClune K, et al. (2007) Stable isotopic variations in west China: A  
904 consideration of moisture sources. *Journal of Geophysical Research* 112:  
905 D10112.  
906 Yanhong W, Lücke A, Zhangdong J, et al. (2006) Holocene climate development on the  
907 central Tibetan Plateau: A sedimentary record from Cuoe Lake.  
908 *Palaeogeography, Palaeoclimatology, Palaeoecology* 234: 328-340.  
909 Zhang P, Cheng H, Edwards RL, et al. (2008) A test of climate, sun and culture  
910 relationships from an 1810-year Chinese cave record. *Science* 322: 940-942.  
911 Zhu L, Wu Y, Wang J, et al. (2008) Environmental changes since 8.4 ka reflected in the  
912 lacustrine core sediments from Nam Co, central Tibetan Plateau. *The Holocene*  
913 18: 831-839.

914

915 **Figure Captions**

916 **Figure 1:** Map of the broader study region showing the location of paleoclimate sites  
917 discussed in the text. Nir'pa Co is marked with a red circle. Other lakes records are  
918 indicated with black circles. Cave records are identified with an open square. Ice core  
919 records are marked with a triangle. Marine records are identified with a diamond.  
920 Abbreviations for paleoclimate records are as follows: PC = Paru Co; AC = Ahung Co;  
921 LZ = Lake Zigetang; SC = Seling Co; TC = Tianmen Cave; NC = Nam Co; MC =  
922 Mawmulh Cave; DC = Dandak Cave; DAS = Dasoupu, BC = Bangong Co, SHC =  
923 Sahiya Cave.

924

925 **Figure 2:** (a) Detailed digital elevation map of the Nir'pa Co watershed (outlined with a  
926 dashed line and shaded blue) showing the position of moraines (while solid lines) and  
927 seasonal (dashed blue line) and permanent streams (solid blue line). (b) Expanded view

928 of Nir'pa Co showing the location of inflow (white dashed line) and outflows (solid white  
929 line), the delta/shelf at the lakes north end and the location from which cores A-11, A-15  
930 and B-15 were collected (white circles). The location of surface samples from the transect  
931 are show with green circles. (c) Monthly average temperature (red) and precipitation  
932 (blue) at Nir'pa Co (Bird et al., 2014). (d) Grain size results from surface sample transect  
933 sites (green circles in 2b).

934

935 **Figure 3:** (a) Excess (red) and supported (blue)  $^{210}\text{Pb}$  activities are shown versus depth  
936 with  $^{137}\text{Cs}$  activities for the top 10 cm of core A-11. Where error bars are not visible, they  
937 are smaller than the symbol size. (b) Sediment depth ages based on the CRS age model  
938 are shown with the depth at which maximum  $^{137}\text{Cs}$  activity occurred (4.5 cm). (c) The  
939 composite age model for Nir'pa Co core A-11 with a second order polynomial fit to the  
940  $^{210}\text{Pb}$  (red circles) and AMS  $^{14}\text{C}$  (blue squares) ages with the  $^{137}\text{Cs}$  peak depth shown  
941 with a black diamond.

942

943 **Figure 4:** Sedimentological and geochemical results from Nir'pa Co core A-11: (a) dry  
944 bulk density; (b) mass accumulation; (c) percent organic matter; (d) organic matter flux;  
945 (e) the ratio of elemental carbon to nitrogen; (f) percent biogenic silica (orange) and flux  
946 (pink); (g)  $\delta^{13}\text{C}$  of organic matter; (h)  $\delta^{15}\text{N}$  of sedimentary nitrogen. Vertical lines  
947 represent the late Holocene mean of each record.

948

949 **Figure 5:** Physical sedimentology and XRF results from Nir'pa Co core A-11: (a)  
950 percent lithics and lithic flux; (b) histogram showing the down core abundance

951 distribution of clastic grain sizes; (c) percent sand; (d) percent silt; (e) percent clay on a  
952 reversed axis to highlight similarities with %silt; (f) magnetic susceptibility; (g) XRF  
953 derived Ti. Vertical lines as in Figure 4.

954

955 **Figure 6:** Comparison of results from Nir'pa Co (a-c) with regional paleoclimate records  
956 (d-j). Nir'pa Co (a) %lithics; (b) %sand on a reversed axis; (c) %silt; Paru Co (d) PC1  
957 lake level curve (e) PC2 erosion curve; (f) leaf wax hydrogen isotopes; (g) Seling Co  
958  $\delta^{18}\text{O}$  of calcite; (h) Bangong Co  $\delta^{18}\text{O}$  of calcite; (i) Lake Zigetang pollen biome index of  
959 relative surface air temperature; (j) Dandak Cave  $\delta^{18}\text{O}$  of calcite. Blue horizontal bars  
960 indicate pluvial phases while the gray bar indicates aridity. Vertical lines as in Figure 4  
961 and 5.

962

963 **Figure 7:** Comparison of the Paru Co (a-b) and Nir'pa Co (c-d) results with other more  
964 distal ISM records from (e) the Arabian Sea marine sediment cores and (f) Qunf Cave  
965  $\delta^{18}\text{O}$ , Oman. Also shown are the (g) El Junco %sand El Niño record and SST  
966 reconstructions from (h) the eastern tropical Pacific and (i) Indo-Pacific. Vertical lines as  
967 in Figures 4-6.

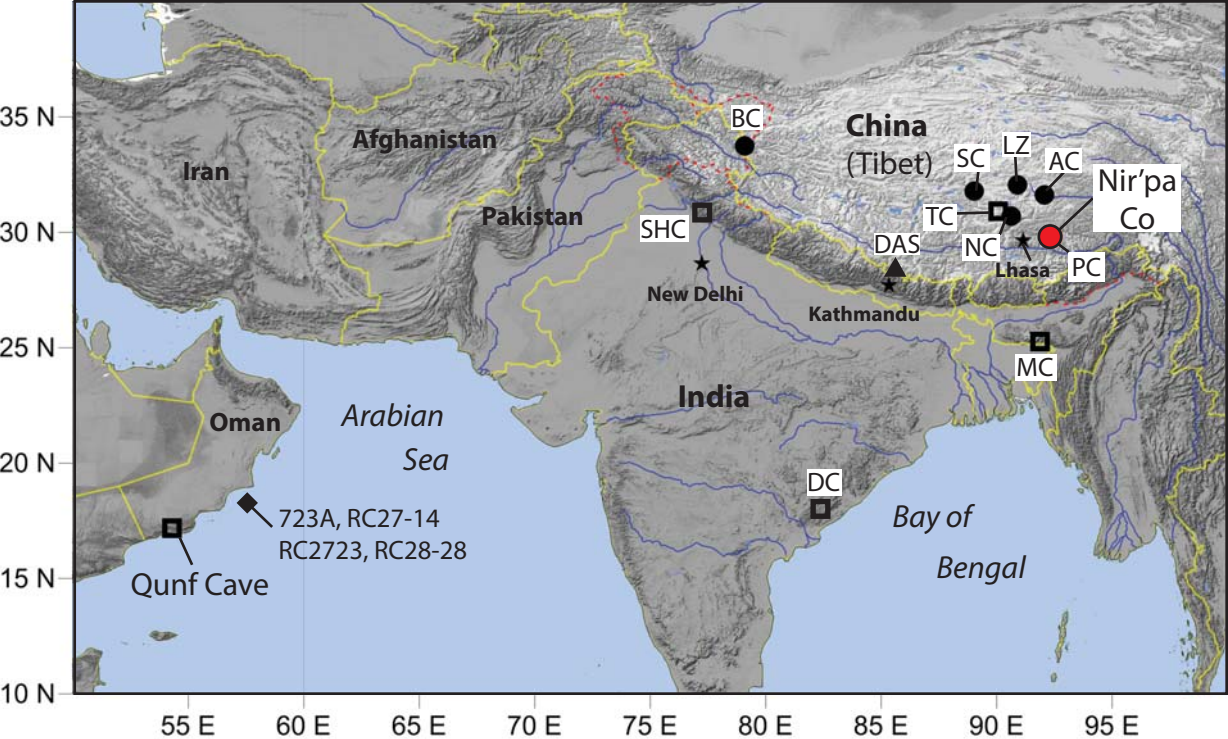


Figure 1: Bird et al. (2016)



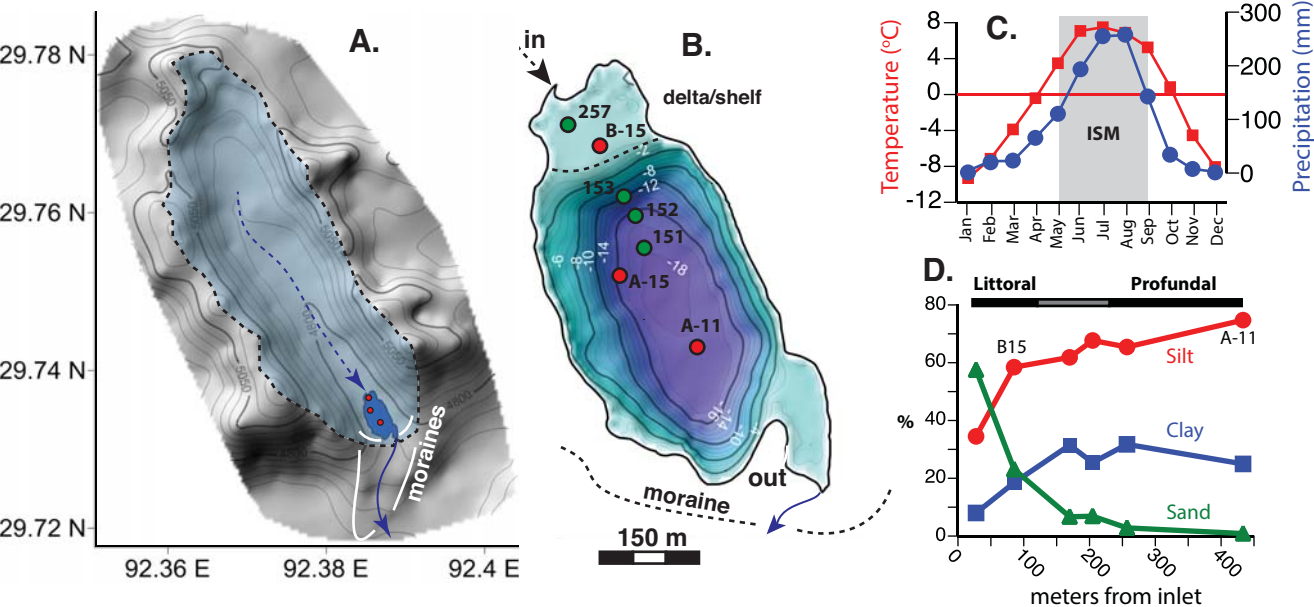


Figure 2 Bird et al. (2016)

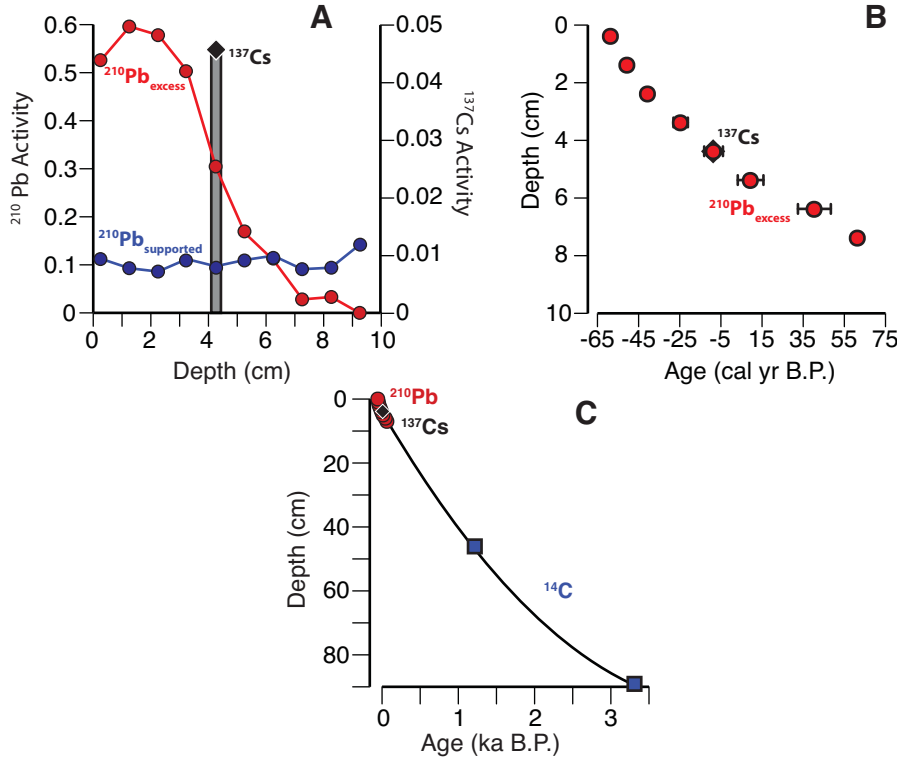


Figure 3: Bird et al. (2016)

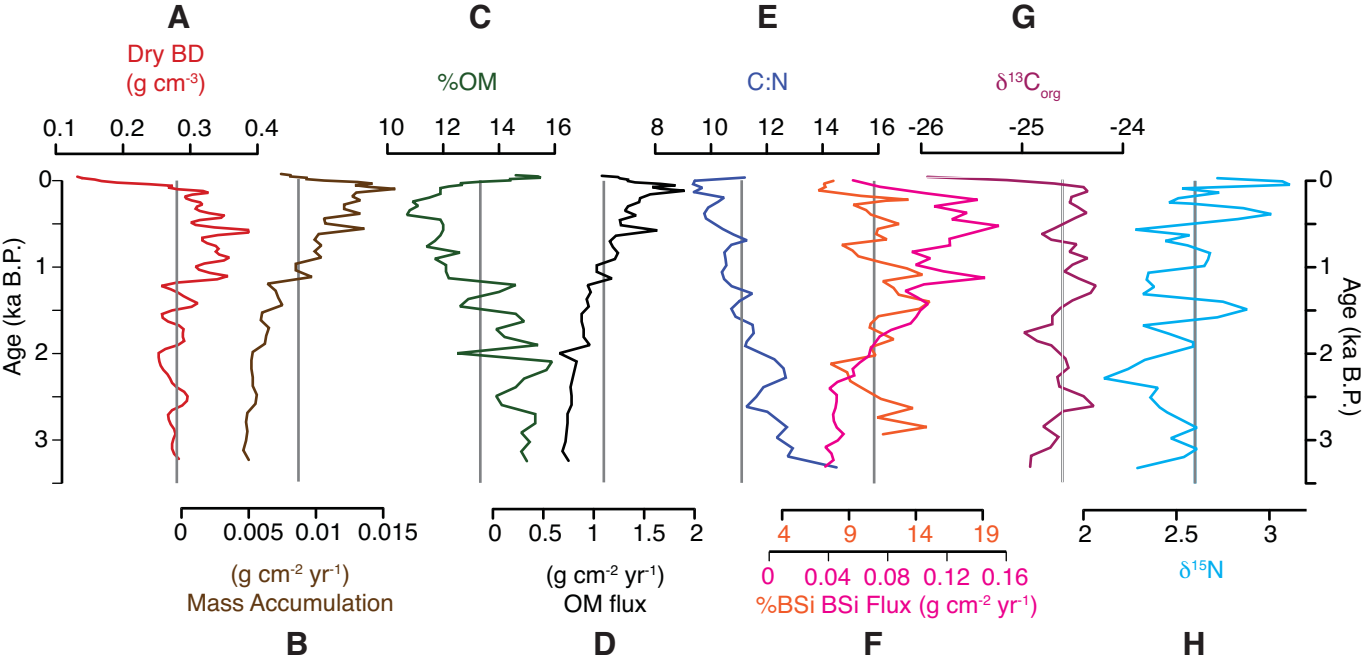


Figure 4: Bird et al. (2016)

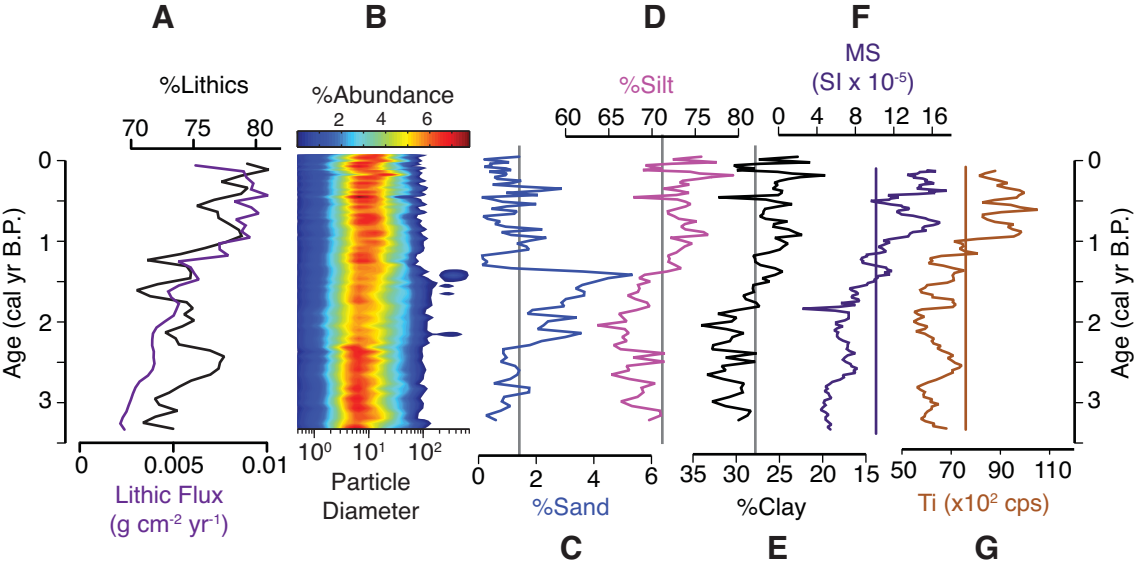


Figure 5: Bird et al. (2016)

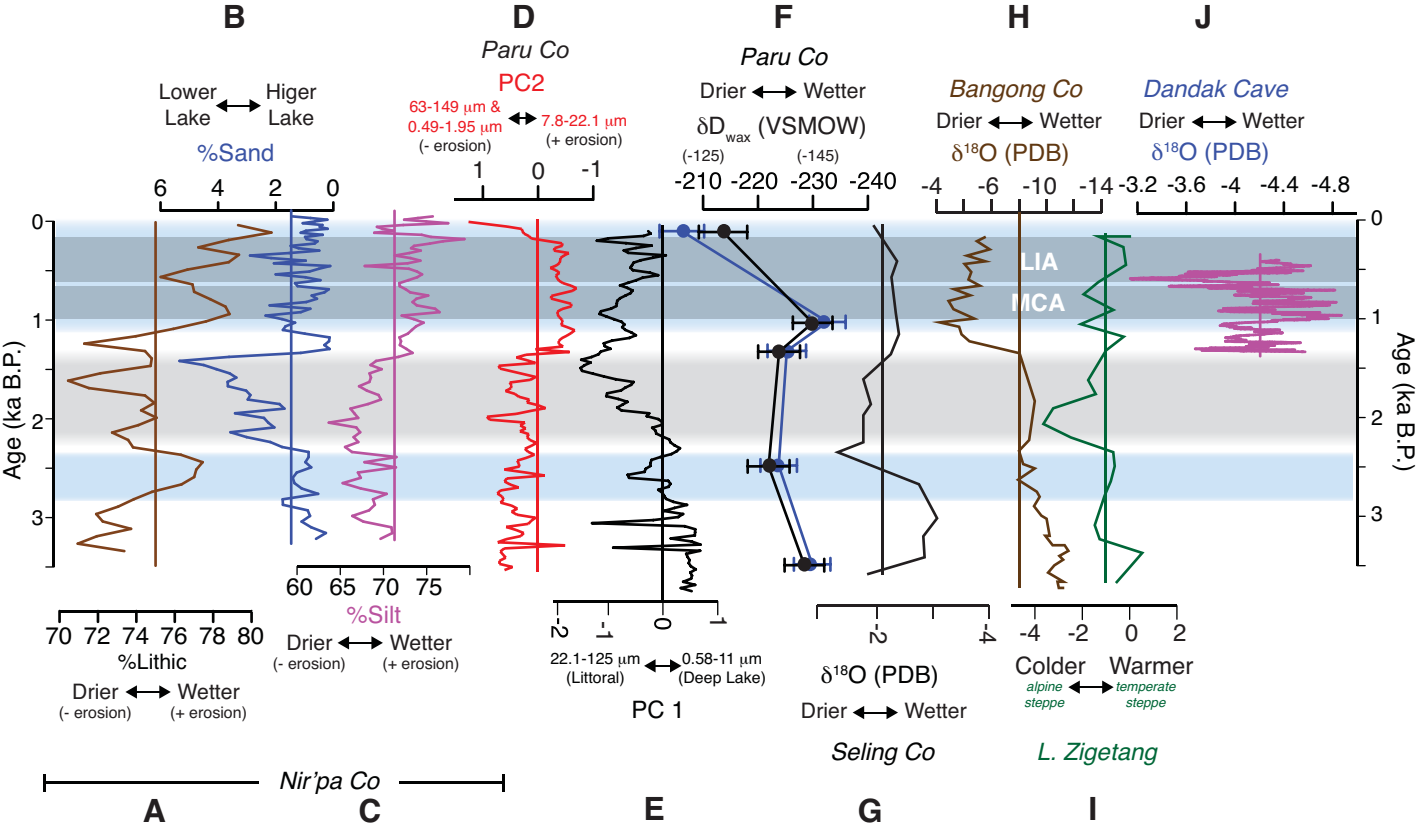


Figure 6: Bird et al. (2016)

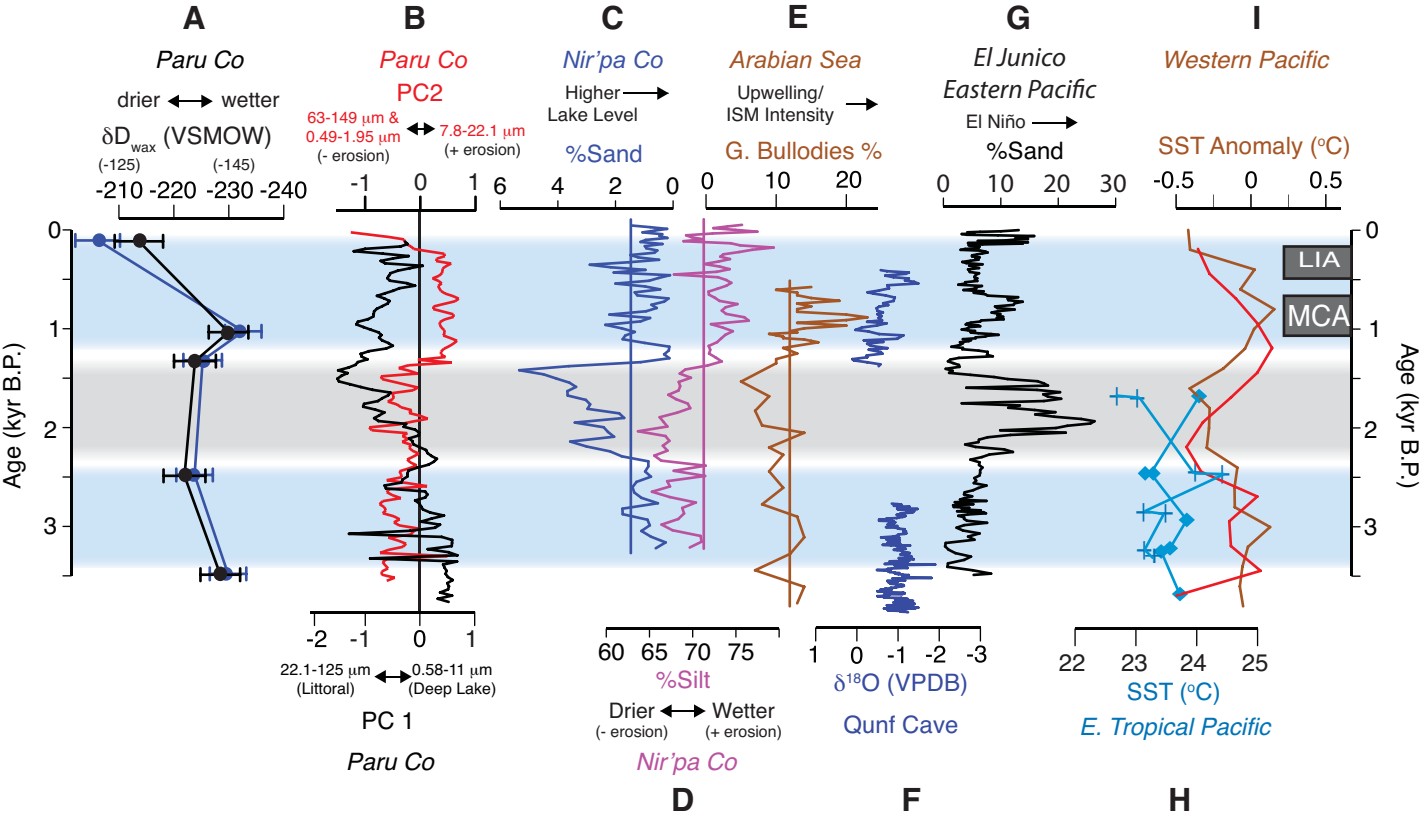


Figure 7: Bird et al. (2016)

**Table 1** <sup>210</sup>Pb and <sup>137</sup>Cs results from Nir'pa Co core A-11.

Mean Depth (cm)	<sup>210</sup> Pb		Supported <sup>210</sup> Pb or <sup>214</sup> Pb activity		Excess <sup>210</sup> Pb		<sup>137</sup> Cs		cal yr B.P.
	Activity	±	(Bq/g)	±	(Bq/g)	±	Activity (Bq/g)	±	
0.25	0.64	0.077	0.113	0.014	0.527	0.063	<DET	0	-56
1.25	0.69	0.077	0.093	0.011	0.597	0.066	<DET	0	-48
2.25	0.664	0.0775	0.086	0.012	0.579	0.065	<DET	0	-38
3.25	0.616	0.0735	0.112	0.016	0.504	0.057	<DET	0	-22
4.25	0.4	0.0545	0.097	0.013	0.303	0.042	0.0449	0.0174	-6
5.25	0.277	0.0431	0.110	0.013	0.167	0.030	<DET	0	12
6.25	0.229	0.0417	0.119	0.013	0.110	0.029	<DET	0	43
7.25	0.12	0.02335	0.091	0.009	0.029	0.015	<DET	0	64
8.25	0.128	0.028	0.094	0.010	0.034	0.018	<DET	0	n/a
9.25	0.151	0.03395	0.142	0.014	0.000	0.020	<DET	0	n/a

**Table 2** Radiocarbon results from Nir'pa Co cores A-11 & B-15.

UCI AMS #	Core	Depth (cm)	Fraction Modern	±	$\Delta^{14}\text{C}$ (‰)	±	$^{14}\text{C}$ yr B.P.	±	Cal yr B.P.	+/-
104262	A-11	46-47	0.8551	0.0029	-144.9	2.9	1255	30	1210	70/40
104263	A-11	88.5-89.5	0.6812	0.0030	-318.8	3.0	3085	40	3290	92/83
165845	B-15	66-67	0.5809	0.0026	-419.1	2.6	4365	40	4930	60/80
165846	B-15	90-91	0.2931	0.0014	-706.9	1.4	9860	40	11250	90/50