Originally published as:

Dietze, E., Wünnemann, B., Hartmann, K., Diekmann, B., Jin, H., Stauch, G., Yang, S., Lehmkuhl, F.
(2013): Early to mid-Holocene lake high-stand sediments at Lake Donggi Cona, northeastern Tibetan
Plateau, China. - Quaternary Research, 79, 3, 325-336

DOI: 10.1016/j.yqres.2012.12.008

1   **Early to mid-Holocene lake high-stand sediments at Lake Donggi Cona,**
2   **north-eastern Tibetan Plateau, China**

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Abstract

Lake high-stand sediments are found in three onshore terraces at Lake Donggi Cona, north-eastern Tibetan Plateau, and reveal characteristics of hydrological changes on lake shorelines triggered by climate change, geomorphological processes, and neo-tectonic movements. The terraces consist of fluvial, alluvial to littoral-lacustrine facies. End-member modeling of grain size distributions allowed quantification of sediment transport processes and relative lake levels during times of deposition. Radiocarbon dating revealed higher than modern lake levels during the early and mid-Holocene. Lake levels follow the trend of Asian monsoon dynamics, and are modified by local non-climatic drivers. Site-specific impacts explain fluctuations during the initial lake level rise ~11 cal ka BP. Maximum lake extension reached ~9.2 cal ka BP, at ~16.5 m above present lake level (a.p.l.l.). Littoral and lacustrine sediment deposition paused during a phase of fluvial activity and post-depositional cryoturbations at ~8.5 cal ka BP, when the lake level fell to ~8 m a.p.l.l. After a second maximum at ~7.5 cal ka BP, lake level declined slightly at ~6.8 cal ka BP, probably due to a non-climatic pulse that caused lake opening. The level remained high until a transition towards drier conditions of ~4.7 cal ka BP. Though discontinuous, high-stand sediments provide a unique, high-resolution archive.

Keywords

Lake level fluctuations; Lake high-stand sediment; End-member modeling (EMMA); Grain size distributions; Tibetan Plateau; Asian monsoon

Introduction

The Tibetan Plateau (TP) provides water to billions of people in Asia and heavily influences the global climate circulation (Qiu, 2008). Moisture availability on the TP is regulated by different circulation systems – wet summer monsoons from the Indian and Pacific Oceans interact with the westerlies and dry winter air masses from the Siberian anticyclone. However, the past interplay of these systems is still under discussion (Chen et al., 2008, Wang et al., 2010).

One way to better understand moisture variability is to study lake level changes. Ancient lake stands can record shifts in the precipitation-evaporation balance of a catchment (Cohen, 2003). Lake stands respond to water supply from glaciers and thawing permafrost, but can also reflect modifications of the basin and catchment configurations. Indications of lake high-stands have been observed on the TP for a long time (e.g., De Terra and Hutchinson, 1934). More recently, shorelines and beach ridges were dated with ^{14}C and OSL (Lee et al. 2009; Li et al., 2009). There are only a few studies from the western TP (Gasse et al., 1991), Ladakh (Wünnemann et al., 2010), and the northern foreland of the TP (Madsen et al. 2008, Long et al., 2010) that consider lake high-stand sediment sequences in detail, i.e., intercalations of lacustrine, littoral and terrestrial sediments deposited above present lake level (a.p.l.l.).

Using stratigraphy and granulometry, this study reconstructs sedimentological processes from intercalations of terrestrial and lacustrine sediment deposited during high-stands of Lake Donggi Cona. Because the Donggi Cona catchment is part of an active tectonic fault system, a major aim is to distinguish climatic from non-climatic triggers of lake level changes using environmental reconstructions of the lake and its vicinities.

Study area

Lake Donggi Cona fills a pull-apart basin at the Kunlun Fault (35°18'N, 98°32'E, lake area: ~230 km², maximum depth: ~92 m, catchment area: ~3200 km²; Figure 1), which has a mean slip rate of ~10.3 mm/a along the Donggi Cona segment (Van der Woerd et al., 2002). Seismic studies (Dietze et al., 2010) and fluvial terraces (Van der Woerd et al., 2002) suggest that vertical motion has been of minor importance in the lake and its catchment during the Holocene.

Climatic conditions in the vicinity of Donggi Cona are characterized by a total pan evaporation of ~1375 mm/year, mean air temperatures in January (July) of -16.8 (7.5) °C and mean annual precipitation of ~300 mm (Madoi station, 4272 m a.s.l., ~50 km southwest, Chinese Central Meteorological Office, 2008). Precipitation falls mainly in summer (~280 mm between May and October) as intense as the torrential rain from local convection of the Asian summer monsoon air masses, which reach their northern limit here (Domrös and Peng, 1988). Overland flows, reactivation of ephemeral channels, and rivers transport the main suspension load to the lake. The lake is normally frozen from late November until early April, when westerlies and the winter monsoon prevail, and bring dry winds that mobilize aeolian sediments from dune fields, loess, and loess-like sediments (Lehmkuhl, 1997; Ujmer et al., 2012a). In addition, dust can be captured discontinuously in the lake ice (Dietze et al., 2012). Several phases of aeolian activity and reworking of aeolian

sediment were reconstructed for the Holocene (Stauch et al., 2012). Yak and sheep grazing on alpine steppes and meadows also cause sediment mobilization (Schlütz and Lehmkuhl, 2009). Wet hummocky land surfaces and small thermokarst depressions indicate that discontinuous permafrost affects saturated sediments close to the lake and in local depressions.

Figure 1

The heterogeneous geological and geomorphological catchment configuration includes dissected series of carbonates, limestones and sandstones faulted against quartzites and shales, all partly overlain by conglomerates. In the upper catchment, lateral moraines and U-shaped valleys indicate past glaciations, while fluvial incision formed V-shaped valleys in the middle reaches. Major perennial inflow enters the lake from the east via a large alluvial plain. The only other perennial stream enters the lake from a flat basin north of Lake Donggi Cona via a transverse valley (Figure 1).

The current lake level is at 4090 m a.s.l. Palaeo-shorelines and ancient onshore terraces that contain lake sediments encircle the lake and indicate four higher lake stands at 3.5 ± 0.4 m, 6.1 ± 1.0 m, 10.1 ± 0.9 m and 16.7 ± 1.2 m a.p.l.l. (T1 to T4, Lockot, 2010). Three lower Pleistocene lake stands at 24, 39 and 57 m below the present lake level were reconstructed using bathymetry and seismic stratigraphy (Dietze et al., 2010). A deglaciation-related lake level rise with strong fluctuations between 17.5 and 14 cal ka BP, was followed by a drop at 14 cal ka BP that produced saline conditions (Mischke et al., 2010, Opitz et al., 2012), while higher levels prevailed between ~4.3 and 11.5 cal ka BP (Opitz et al., 2012). There is debate regarding when the lake switched to an open system: at ~6.8 cal ka BP when a change in ostracod assemblages from euryhaline (brackish) to freshwater species occurred (Mischke et al., 2010); or at ~4.3 cal ka BP as suggested by changes in the geochemical and mineralogical properties of lake sediments (Opitz et al., 2012).

Today, the lake drains through an artificial channel at the western margin, south of the active course of the Kunlun fault. It merges with a perennial stream that originates from the north and then drains towards the northwest (Figure 2A). A gauge station was set up during the 1970s to control the water supply towards the endorheic Qaidam Basin.

Methods

Field methods

During May and September 2009, nine high-stand sediment sections at Lake Donggi Cona were described at modern fluvial channel exposures, permafrost decay fronts, anthropogenic gravel pits or were dug along terrace front slopes (Figs. 1, and 2). The highest elevation of the lake sediments was determined using a differential GPS. Sedimentological structures and layering were described, including major macroscopic characteristics such as sediment color, charcoal, and snail fragments. Burrows and distorted sediment layers indicate post-depositional alteration of the sections. Additionally, pedological description focused on features of redoximorphosis, clay and carbonate relocation, and root penetration.

Laboratory methods

The complete sections of P06 and P14, and the lake sediment parts of sections P15, P02 (undisturbed parts) and P21 (including matrix material within gravel layers) were analysed. Prior to analysis, all samples (taken in 2cm slices at different intervals) were sieved to <1mm. Larger grain fractions were not considered, as only 26 samples had material >1mm. To account for the detrital components of the grain-size distributions, 262 samples were pre-treated with 10% acetic acid to remove the carbonate fraction, and 35% H₂O₂ for at least 72 h to remove organic matter, before being placed on an overhead shaker for at least 12 h, together with 10 mg of sodium pyrophosphate. Grain size distributions were measured with a laser diffraction particle size analyser (Beckmann Coulter LS 200, calculated with a Fraunhofer model). Volume percentages of 85 grain-size classes from 1000 to 0.38 µm were compiled in a data matrix.

Statistical methods

Single grain size parameters such as sand, silt, and clay contents, or the associated methods of moments (e.g., mean, skewness, kurtosis), are biased when applied to multi-modal distributions (Dietze et al., 2012), which is why such results were omitted here. Instead, the original grain size distributions were unmixed using end-member modeling analysis (EMMA) – an eigenspace decomposition with different scaling procedures that extract genetically meaningful end-member grain size distributions (i.e., loadings) and their percentages in each sample (end-member composition, i.e., scores; Dietze et al., 2012). End-members (EMs) can be interpreted in terms of sediment transport processes, and thus, characterize typical depositional environments (following e.g., Folk and Ward, 1957).

The 10th quantile ($l = 0.1$) was applied in the weight transformation after Dietze et al. (2012), which yielded the best unmixing and modeling results compared to other model configurations. The robustness of the resulting end-members was tested by different weight transformation and single section EMMA runs (results not shown). A normalized difference between the finest and coarsest end-members (i.e., EM_{diff}) was calculated as a proxy for relative lake level change – assuming that the finest particles deposit in calm, deeper water, while fractions coarser than silt settle close to the shore, shortly after entering the lake.

Radiocarbon dating

AMS dating of eight sediment sections was done at Poznan radiocarbon laboratory. Because pollen dating failed due to poor preservation, 14 bulk lake sediment samples, two charcoal remains, and three non-recrystallized (aragonite) *Radix* shells were dated. The potential hard-water error for TOC was determined by dating the topmost samples of three lake cores, including their 2σ-errors (Opitz et al., 2012). The up-to-date absolute minimum and maximum hard-water errors for bulk sediment (i.e., 1920 and 2360 years, respectively) were subtracted from all bulk and snail ¹⁴C ages. Then, all samples were calibrated with IntCal09 (Reimer et al., 2009) in Calib 6.1.0 (Stuiver and Reimer, 1993). The absolute minimum and maximum in the 2σ-ranges of the corrected and calibrated ages give a larger uncertainty than usual after calibration, but represent a more realistic consideration of the potential hard-water and reservoir effects. All mean ages mentioned in the text and figures refer to corrected, cal ka BP.

Results and interpretation

Sediment sections and radiocarbon dating

High-stand sediment sections were grouped according to relative elevation and location. They are located at the outlet (P15, P04, P06), at the northern T3 shore (P02, P21, P14, P13) and at the eastern alluvial plain (P16, P17; Figure 1 and 2; Table 1).

Figure 2

Table 1

Table 2 contains all dating results with 2σ -errors. All high-stand sediments were deposited between ~ 11.4 and ~ 3.5 cal ka BP (Figure 3). There are two main age clusters in the early and mid-Holocene. High-stand lake sediments accumulated at >8 m a.p.l.l. until ~ 7 cal ka BP in T4 (P15) and T3-terraces (P04, P02, P21, P14, and P13), while mid-Holocene lake sediments dominate lower elevations and are exposed in T2 (P06) and T1 terraces (P17). The highest section, P15, has an age of ~ 6.9 corr.cal ka BP, which is ~ 1.5 ka younger than an OSL-age of the loess at its top (section P063 at the base: 8.5 ± 0.8 ka, Stauch et al., 2012). The OSL-age coincides with a hiatus in lake sedimentation between 8.2 and 8.7 cal ka BP (Figure 3). There are also stratigraphically inverted ages in P06 and P16.

Table 2

Figure 3

Classification of depositional environments

All sections show several sedimentation unconformities (“U” in Figure 4), and have sub-catchment-specific sedimentation rates and grain sizes. Figure 4 shows generalized section sketches with depositional zones assigned to three sediment facies that represent different depositional environments along the littoral zone (Figure 5). Sites are too heterogeneous to integrate them in one “composite” section, but they can be preliminarily correlated using the elevation, ages, and characteristics of deposited sediments and their post-depositional overprint.

Sediments of Facies I bear a platy structure, high clay and carbonate contents, grayish-green or whitish-blue color with post-depositional oxidized spots. *Radix* and *Gyraulus* snails are common. Facies I can be found in all sections except for P13 and P15 (Figure 4). A large part of P06 contains mm-thick, clayey and silty laminae. In P14, high amounts of carbonate tubes were found above and below a unique thin, dark layer at ~ 67 cm depth. At sites P16 and P17 and at ~ 170 and ~ 210 cm depth of P04, Facies I sediments show a more granular structure (Figures 4, 5). Facies I is interpreted as lake sediment deposited in a deep water environment, below the wave-dominated zone (Figure 5). There, organic matter, carbonate precipitates, silt and clay can settle causing the typical platy structure, or even laminations. The black, organic-rich layer in P14 could represent short-term sapropel deposition.

Facies II sediments are generally loose, well-sorted, mainly sandy, coarser deposits that are found in most sections. In section P21, for example, the fine sands at the base and the shingle bedding of cobbles at ~165 cm depth are assigned to this Facies (Figure 4). Facies II is interpreted as littoral sediment (Figure 5); silts and clays are washed out and distributed by currents within the lake.

Figure 4

Depending on the site, sediments of Facies III show a more heterogeneous structure with either: a) cemented or loose, horizontally-bedded gravels (e.g., at the base of most sections, as well as at ~60, 140 and 190 cm depth in P21); b) unconsolidated sandy material with homogenous bedding or a coarsening upwards sequence at ~265 cm depth of P02; or c) silty-sandy cross-laminations (e.g., 190-230 cm depth in P14, Figure 4). Facies III is interpreted as fluvial sediment, which is generally coarser than lake sediment, even in the littoral zone. The heterogeneity of this Facies results from diverse processes comprising confined (in-channel) and unconfined (out-of-channel, alluvial) sediment transport (North and Davidson, 2012). Loose layers of soft, silty aeolian sediment cover most of the sections. They were deposited on the slopes and reworked by unconfined overland flows during intense summer precipitation. As there was no in-situ loess in or on top of the sections, the dominant sedimentation process is assumed to be alluvial.

Post-depositional overprint by pedogenesis, bio-, and cryoturbation varies slightly between sites. The topmost sequences always have a brownish or dark coloration, and sometimes translocated carbonate and clay. These are interpreted as the initial stages of Kastanozem- or Cambisol-like soils at dry, steppe sites (Miehe et al., 2008). Intense root penetration and animal burrows (e.g., *Ochotona spec.*) indicate sediment mixing, throughout section P14 and at section P15. Ongoing active layer mixing created a typical cryogenic granular structure in Facies I, especially at sites that are connected to groundwater (e.g., sections P04, P16, and P17 (grain sizes not analysed)).

Figure 5

In section P02, intense post-depositional deformation and involution affect sediments of all Facies between 48 and 180 cm depth, including the gravel layers (Figure 2B). Although it is possible that these extensive involutions are seismites and their structures are related to liquefaction in formerly near-shore, water-saturated sediments they are instead attributed to past periglacial mixing (cryoturbations).

An ambiguous zone in P14 between 100 and 190 cm depth contains involutions of fine sediment, burrows, and roots, with a distinct change from Facies III to II at 130 cm depth. Neither past cryoturbations, nor fluvial reworking (Figure 2F) can be excluded as deformation mechanisms for this zone.

Detrital grain-size end-members

Unmixing of detrital grain size distributions yielded an optimal model with five end-members (EMs) explaining 89.9% of the data variance (mean r^2 between the original and modelled data generally >0.7 ($p<0.01$); Figure 6B/C), of which EMs 1 to 5 explain 6%, 24%, 26.3%, 16.5% and 27.2%,

respectively. Except for a few strata, grain size end-members independently support the field descriptions (see below, Figure 7).

Figure 6

EM 1 has a broad mode in the coarse and medium sands with a maximum at 715 μm in its loadings (Figure 6A). It contributes to the sample compositions (i.e., scores) mainly at the base of the sections and in the coarse sediments of Facies III (e.g., 10% in the matrix material of P21 gravel layers, 80% of the P02 coarsening-upwards layer, Figure 7). EM 1 is interpreted to contain grain sizes that are deposited by high-energy fluvial transport, though it is also found in littoral sediments that are affected by high-energy wave activity.

EM 2 has a narrow peak in the very fine sand (mode at 76 μm). It has high scores in the fine sediments of Facies III (e.g., ~210 cm in P14, bases of P21, P06, and 20% of the coarsening-upwards layer in P02). Hence, EM 2 represents sediment from low-energy unconfined alluvial flow, i.e., fine sand is deposited when flow velocity reduces, while finer particles are washed farther down into the lake. An aeolian origin for EM 2 is excluded, as it is too fine to represent dune sand, which would be similarly narrow-peaked but in a coarser range (Sun et al., 2002). Furthermore, dunes are unlikely to occur at the position of the included sections (Ijmker et al., 2012a).

The coarse silt EM 3 has a narrow peak at 44 μm , while EM 4 shows a broad mode in the fine silt (maximum at 13 μm). Both are present in most of the samples with fractions of $30 \pm 24\%$ and $31 \pm 19\%$, but it is found mainly at the top of the sections (Figure 7). EMs 3 and 4 are interpreted as aeolian deposits. EM 3 has the same mode as loess from the TP (Sun et al., 2007), which is transported in short-term suspension by near-surface, mainly winter monsoonal circulation (Sun et al., 2002; Ijmker et al., 2012b). EM 4 represents background sedimentation of remote dust. EM 3 and EM 4 dust can be reworked along the slopes and reach the lake in fluvial suspension, but can also be trapped in ice off-shore (Dietze et al., 2012).

The finest grain size end-member, EM 5, has a broad mode in the clay fraction at 3.5 μm (Figure 6A). EM 5 is generally high in the strata of lacustrine sediments and low in fluvial/alluvial sediments. Hence, EM 5 is interpreted as the finest suspension that reaches deeper lake areas and accumulates very slowly under calm water conditions (off the wave-dominated zone; Figure 5). EMs 4 and 5 are the robust end-members in modern lake surface sediments (Dietze et al., 2012).

EM_{diff}, the normalized difference of the low-energy fluvial EM 2 and the suspension EM 5, shows a similar but more pronounced pattern than EM 5, and can be readily interpreted as a proxy for lake depth.

Figure 7

Lake level reconstruction from sediment facies

When lacustrine, littoral/near-shore, or fluvial/alluvial sediments are identified using field observations and EMMA, phases of higher, similar, or lower lake levels are assigned relative to the present elevation of sediment in sections (Figure 4 and 7). Furthermore, post-depositional

pedogenesis, bioturbation, and cryoturbation point to a lower lake level after deposition. For example, in section P02, cryoturbations distorted all sediments between 48 and 180 cm depth. Permafrost deforms sediments, when sufficient water is available, especially at sites with imperfect drainage (Van Vliet-Lanoë, 1998). Below 180 cm depth, sediments consist of a high percentage of sand (EM1 and 2), which would have allowed good drainage if the water level was much lower. Hence, it is assumed that a lake level elevation close to 8 m a.p.l.l. triggered cryoturbations between 48 and 180 cm.

Discussion

Interpretation of Facies and depositional environments, as well as respective elevation is especially difficult for the ambiguous, fine-grained strata marked with “?” in Figure 4. Some strata contain sediments with platy structure, but high silt content and low amounts of suspension EM 5 (e.g., ~95 cm in P04, P14; Figure 7), as well as redoximorphic features (e.g., at ~80 and 110 cm depth of P02), and gastropods (e.g., P14 at 140 cm). An explanation is that littoral and near-shore environments can host fine-grained sediments of all Facies depending on the position along shore. Additionally, the presence of gastropods generally indicates near-shore conditions, but *Radix spec.* can also develop in small riverine and beach lagoons (Taft et al., 2012), independent of a large permanent lake. Alternatively, some strata contained high amounts of suspension EM 5 but lacked other Facies I features (e.g., P14 at 295 cm). However, wave action, fluvial deltas, and backshore configurations vary strongly between sites in this heterogeneous setting (Dietze et al., 2010), and a littoral deposit may lack typical, representative grain size end-members. Finally, other strata have been overprinted post-depositionally, convoluting their original depositional characteristics (e.g., P02, P14 and the top of P21). Hence, all of the ambiguous strata are tentatively assigned to the littoral zone or a deeper lake environment and, therefore, lake level reconstruction is also tentative.

Inverted ages may result from reworking of older material during lake level change or from varying hard-water contribution across time and space, which can only initially be considered here. High percentages of limestone in the catchment, and probably old CO₂ from groundwater, bring dead carbon into the system. The inverse ages in P06 – a section close to a limestone ridge at the outlet – may be related to hard-water effects, as there are no significant stratigraphic or granulometric differences in this depth range. Another likely explanation considers random mixing by tectonic shaking of saturated and unconsolidated sediments. However, no obvious signs for tectonic influence (e.g., vertical faults, or (micro)kinks in the sediments, Van Vliet-Lanoë, 1998) were found in any of the sediments.

An underestimation of bulk sediment ages may result from modern fine roots, bio- and/or cryoturbation that contribute younger atmospheric carbon to the sediments (symbols in Figure 3). Section P17 may suffer from all these effects. P16 has a similar intense cryogenic granular fabric and is much younger than sections at the same elevation. Neither is considered in the discussion. The ages of the root-rich, near-shore sediments in P15 and P02 might be ~2 ka older due to the age of the overlying loess (Stauch et al., 2012), and correlation with lake sediments preserved at similar

elevation and stratigraphic position (top of P14, no. 10; Figure 3), respectively. Only unbiased ages and their large uncertainties were included in the following chronology (Figure 8).

Chronology of lake level changes

The high-stand sediments are in agreement with the lake-core stratigraphy of Opitz et al. (2012). Their litho-unit 3 represents the time interval presented here, which can now be resolved in more detail.

Early Holocene (~11.4 to 6.8 cal ka BP)

After the Younger Dryas, the lake level was fluctuating, but rising overall: at ~11 cal ka BP, littoral sediments intercalated with lacustrine and fluvial/alluvial sediments, forming the base of P14 and P02 at a minimum elevation of 7.1 m a.p.l.l. (Figure 3: age no. 7, 12). A discontinuous rise in lake level between ~10.5 and 9.8 cal ka BP is inferred from lacustrine and littoral sediments accumulated at the base of all northern shore sections and P04 at the outlet. Strong fluctuations in lake level and reworking of older sediment may be responsible for the inverted ages in P02. A prominent lake level drop around 10 cal ka BP is indicated by a clear erosional surface followed by a zone of reddish alluvial sediment at 180 cm depth of P02 (at 8.2 m a.p.l.l.). In P04 and P14 a thin gravel layer at 140 and 240 cm depth (8.1 and 8.7 m a.p.l.l.), respectively, and the first gravel/shingle phase in P21 (130-200 cm depth, ~8.5 m a.p.l.l.) might coincide with this lower lake level.

Subsequently, lake level rose rapidly, with apparent fluctuations allowing deposition of near-shore sediments intercalating with fluvial Facies at 9 to 10 m a.p.l.l. (e.g., P02-sediments at 50-180 cm depth, up to 20 cm in P21). Starting at ~9.8 cal ka BP, this rise probably lasted a few centuries until ~8.7 cal ka BP, as the small base-top age differences in P02 and P21 suggest (Figure 3: age no. 4-7, 8-9). The highest lake levels (16.5 m a.p.l.l.) are tentatively assumed around 9.2 cal ka BP from associated lake sediments found in the cryoturbated strata of P02 (Figure 3: age no. 4) and from the OSL-age of the loess on top of the highest section P15. An upper erosional boundary is prominent in all northern shore sections (e.g., the thin pebble layer in P02 at 47 cm depth and a shift to cover sediments in P04 and P21). Hence, a significant lake level decline took place after this first lake maximum.

In P14, a distinct boundary between lacustrine and low-energy floodplain sediments (high scores of EM 2) occurs at 230 cm depth. The 1-m-thick floodplain sediments were deposited during a time of lower lake levels. They may contain a phase of lacustrine sedimentation, as end-members suggest (high EM_{diff} between 110 and 150 cm, Figure 7). However, when the lake shifted to a lower level it is possible that dynamic fluvial activity, coupled with a much larger catchment than at P02 and P21 (Figure 1), intensively eroded sediments of the previous high stand – leaving behind only the thin lacustrine layer at ~235 cm below the floodplain sediments (i.e., at ~8.75 m a.p.l.l.).

A prominent lake level drop likely occurred at ~8.7 cal ka BP. So far no lacustrine sediments have been dated that were deposited between ~8.7 and 8.2 cal ka BP (Figure 3). Instead, P02-sediments are intensively cryoturbated, supported by saturated sediments below, which provided the water for frost action. Therefore, the lake level was probably at an elevation close to ~8 m a.p.l.l. Slightly cooler conditions are suggested by a short-term reduction in offshore TOC and biological activity (Aichner et

al., 2012). At the same time, the floodplain sediments in P14 may have accumulated (above 8.8 m a.p.l.l.), supporting the idea that the local river bed was slightly above 8 m a.p.l.l. However, as discussed above, these sediments might be much older, and the distortions seen in the ambiguous zone could be dominantly of cryogenic origin, similar to P02. Regardless, all explanations suggest a lower lake level after the lake maximum.

After ~8.2 cal ka BP, another significant lake extension led to the accumulation of lake sediments that overlie the distorted strata in P02 and comprise the topmost lake sediment in P14. This lake stand at ~7.5 cal ka BP (Figure 3: age no. 10) reached far into the transverse valley, to a minimum of 10.6 m a.p.l.l. in section P14. However, because there are no higher lake sediments at other sites with unbiased ages or similarly clayey sediments, as found in P14, a secondary lake maximum likely occurred at this time (Figure 8). The sapropel layer in P14 has probably formed during a slight lake reduction (see EM_{diff} , Figure 7) during times of high biological productivity at ~8 cal ka BP (offshore TOC maximum, Aichner et al., 2012). Lacustrine sedimentation also commenced at the outlet at ~7.8 cal ka BP (P06 base, Figure 4).

Mid-Holocene (6.8 – 4.3 cal ka BP)

A further lake level decline to below 10 m a.p.l.l. caused the topmost transition from lake to alluvial sediment in P02, P13 and P14. Compared to the lake decline at ~8.5 cal ka BP, this sequence has not been overprinted post-depositionally by cryoturbation, suggesting either different climatic conditions, or a decoupling of the topmost sediments from ascending water. To date, there are no further recognized lake sediments in T3 or T4 terraces with unbiased ages younger than 7 cal ka BP. Hence, this lake decline might coincide with the abrupt lake opening at ~6.8 cal ka BP, when ostracods in offshore lake sediments changed from euryhaline to freshwater assemblages (Mischke et al., 2010).

However, massive, finely laminated lake sediment accumulated at P06 in the T2 terrace (Figure 7), even after this transition. The only sedimentological indication for a lake opening at this site might be the reduction of EM_{diff} and an increase in aeolian activity (dust EMs) between 143 and 170 cm depths.

Afterwards, EM_{diff} suggests a lake rise to a level slightly lower than before. The inverted ages in P06 after this transition may suggest either a change in the hard-water nature of the lake, or a reworking of some older organic material. After ~5 cal ka BP the lake gradually reduced with some fluctuations (unconformity overlain by a sandy layer at 100 cm depth, and silty alluvial sediment at 45 cm depth; Figure 7). P06 top sediments still contain some detrital clay associated either with soil formation, or near-shore conditions, but consist mainly of (reworked) dust marking the end of high-stand sedimentation at a minimum elevation of 4.3 m a.p.l.l. This transition might coincide with the suggested change in lake chemistry and stratification recorded at ~4.3 cal ka BP in offshore lake sediments (Opitz et al., 2012).

Synthesis

Although a truly quantitative lake level reconstruction is restricted by the high spatial variability in the accumulation, erosion, and post-depositional overprint of the different sites, and the importance

of water contributions from local sources (i.e., melt and ground water) is difficult to assess, the mechanisms involved in Lake Donggi Cona hydrological variations can now be discussed in a larger context. The basin-wide correlation of facies and lake level assumptions (Figure S1) can be compared to known patterns in Asian monsoon precipitation. Both the Indian and East Asian monsoons may have dominated the site. However, in a comparison with several monsoon reconstructions (not shown) the pattern of an Indian monsoon proxy record, i.e., the $\delta^{18}\text{O}$ of Oman speleothems from Fleitmann et al. (2003) (Figure 8) showed trends similar to this study. Additionally, Wang et al. (2010) found that the Indian monsoon was the dominant influence at the TP during the early Holocene.

The trend of rapidly rising lake level correlates well with the trend of increasingly warmer and wetter conditions during the early Holocene. These resulted from stronger Asian monsoons and are related to the Northern Hemisphere insolation maximum (e.g., Fleitmann et al., 2003; Dykoski et al., 2005). As a result, lakes extended all over the TP during the early Holocene, fed also by melting glaciers, permafrost (Mischke and Zhang, 2010), and increased fluvial activity (Schlütz and Lehmkuhl, 2009). The ~16.5 m a.p.l.l. maximum of Lake Donggi Cona at ~9.2 cal ka BP correlates with this pattern. However, the onset and timing of the ‘Holocene Optimum’ and maximum lake levels varied between sites across Tibet (Mischke et al., 2009; Wang et al., 2010). Lake high-stands during the early Holocene are reported, e.g., from Lake Qinghai (~200 km north-east of Donggi Cona, +8-12 m prior to 8.4 cal ka BP, Madsen et al., 2008), Selin Co (+48 m at ~9.2 ka, Li et al., 2009), and others (Lehmkuhl and Haselein, 2000).

Lake Donggi Cona may have responded to the prominent centennial cooling and drying at ~8.5 cal ka BP described at various sites around the world (e.g., Rohling and Pälike, 2005; Wanner et al., 2011). Between ~8.2 and 8.7 cal ka BP, intense cryoturbations at site P02, probable floodplain deposition at the second largest inflow (P14), and incision at many sites took place (e.g., probably formation of T4-lake terrace). However, the cooling’s impact and timing on the TP are debated (Morrill et al., 2003; Jin et al., 2007) and it is only recognized at some sites in Tibet between 8.7 and 8.2 cal ka BP (e.g., W-Tibet: Gasse et al., 1991; central Tibet: Herzsuh et al., 2006; Lake Qinghai: Colman et al., 2007; E-Tibet: Mischke et al., 2008), suggesting site-specific response times.

Donggi Cona again reached a level of ~11.5 m a.p.l.l. at ~7.5 cal ka BP, correlating with a maximum in monsoon intensity (e.g., Fleitmann et al., 2003; Figure 8). Lake Kuhai, around 40 km to the east of Lake Donggi Cona, reached its highest levels between 12.8 and 7.1 cal ka BP (Mischke et al., 2009). Several other lakes reached their maximum extent during the Mid-Holocene (e.g., Lake Koucha ~230 km SW of Lake Donggi Cona, Mischke et al., 2008) in accordance with the dominance of the East Asian monsoon on the TP (Wang et al., 2010), though this maximum is often attributed to local non-climatic dynamics (Wünnemann et al., 2010).

Figure 8

After 7 cal ka BP Asian monsoon intensities gradually decreased (Fleitmann et al., 2003). This led to a transition towards a cooler and drier late Holocene at around 4.5 cal ka BP that can be observed globally (so-called “Neoglacial”, e.g., Wanner et al., 2011) and across the TP (e.g., Colman et al., 2007; Mischke and Zhang, 2010, Wünnemann et al., 2010). Similarly, this transition might be

reflected by a basin-wide change that is prominent in most offshore lake sediment proxies between 3.9 and 4.7 cal ka BP (Opitz et al., 2012) and is associated with the gradual lake decline seen in section P06. Donggi Cona lake level fell permanently below 4.3 m a.p.l.l. after ~4.7 cal ka BP. This climatic transition towards drier conditions could have allowed a deeper incision of the outflow, probably associated with a reduction of alluvial aggradation there and the initiation of T2-terrace formation. Incision probably stopped at the level of T1, which was formed prior to the artificial extension of the outlet in the 1970s (Lockot, 2010).

However, as in other Tibetan lake catchments, the Donggi Cona area is affected by active tectonic and geomorphological dynamics that can randomly intensify or weaken the sedimentological response to climatic change significantly, possibly invalidating hydrological reconstructions. Two main stratigraphic features may dominantly reflect these local processes. One is the fluctuation during the early Holocene lake level rise. It can hardly be correlated between different high-stand sections, and may instead be associated with varying melt-water contributions and distinct adaptations of the sub-catchments to the changing climatic conditions (e.g., sediment mobilisation, interaction with changing vegetation, and permafrost thawing). The site most sensitive to such variations is the outlet spillway that could have randomly been eroded or blocked by the large alluvial fan from the north (Figure 2A).

Another feature is the abrupt opening of Lake Donggi Cona as reflected by ostracod assemblages at ~6.8 cal ka BP (Mischke et al., 2010). A tectonic and/or geomorphological event is more likely responsible for the change than hydrological and climatic drivers. On- and offshore lake sediment cores show short-term excursions in sedimentological, mineralogical, and geochemical parameters at this time that may be linked to a short-term decline of lake level, probably incising T3 terrace. However, no consistent shift in stratification and geochemical processes took place (Opitz et al., 2012). The lake remained at a high level afterwards (5 to 7 m a.p.l.l., i.e., T2-level after Lockot, 2010). Furthermore, to date no significant lake level decline or moisture reduction has been reconstructed for the north-eastern TP (Wang et al., 2010). A significant global cooling phase at ~6.3 ka rather occurred after the opening (Wanner et al., 2011, Figure 8).

Conclusions

Complex geomorphological and sedimentological processes interact in the littoral zone that can be recorded in onshore high-stand sediments. At Lake Donggi Cona, these sediments, while more discontinuous, provide a unique, multi-process archive of past lake level variations and associated sedimentological dynamics that are more diverse and detailed than profundal lake sediments. Stratigraphy, end-member modeling of grain size distributions, and radiocarbon dating considering hard-water uncertainty ranges, allowed quantification of sediment transport processes and relative lake levels during times of deposition.

Lake levels generally follow the trend of Asian monsoon dynamics, and are modified by local non-climatic drivers. During a warmer, wetter climate, Donggi Cona rose from its glacial low-stands to above its present level (starting ~11 cal ka BP), reaching separate maximums during the early and

mid-Holocene. A major cold and dry phase at ~8.5 cal ka BP caused a reduction in lake size and significantly overprinted the high-stand sediments by cryoturbation. The end of high-stand sedimentation correlates with the decline of monsoon dynamics at the transition to the Neoglacial.

However, Lake Donggi Cona is one of many lakes in an active tectonic setting. Its spillway can be affected by small-scale tectonic and geomorphologic processes that could cause the lake to outflow, incise, and erase information that is otherwise recorded. Such non-climatic dynamics may explain the lake opening at ~6.8 cal ka BP, while sub-catchment specific processes may explain the spatial variations between sites during lake level fluctuations at the beginning of the high-stand period.

Acknowledgements

We thank all Chinese and German colleagues for the support and discussions during the field campaigns. M. Runge and M. Paprotzki helped with lab analysis. DFG-priority program 1372 provided financial support. Hucai Zhang and an anonymous reviewer made some useful suggestions.

References

- Aichner, B., Herzschuh, U., Wilkes, H., Schulz, H.-M., Wang, Y., Plessen, B., Mischke, S., Diekmann, B., Zhang, C., 2012. Ecological development of Lake Donggi Cona, north-eastern Tibetan Plateau, since the late glacial on basis of organic geochemical proxies and non-pollen palynomorphs. *Palaeogeography, Palaeoclimatology, Palaeoecology* 313–314, 140–149.
- Chen, F., Yu, Z., Yang, M., Ito, E., Wang, S., Madsen, D. B., Huang, X., Zhao, Y., Sato, T., Birks, H.J.B., Boomer, I., Chen, J., An, C., Wünnemann, B., 2008. Holocene moisture evolution in arid central Asia and its out-of-phase relationship with Asian monsoon history. *Quaternary Science Reviews* 27, 351–364.
- Chinese Central Meteorological Office, 2008. Average climatology for 841 meteorological stations throughout China, 1951–2007 (Meteorological Data of China). Meteorology Press, Beijing.
- Cohen, A. S., 2003. *Paleolimnology: the history and evolution of lake systems*. Oxford University Press, New York.
- Colman, S. M., Yu, S.-Y., An, Z., Shen, J., Henderson, A. C. G., 2007. Late Cenozoic climate changes in China's western interior: a review of research on Lake Qinghai and comparison with other records. *Quaternary Science Reviews* 26, 2281–2300.
- De Terra, H., Hutchinson, G.E., 1934. Evidence of recent climate change shown by Tibetan highland lakes. *Geographical Journal* 84, 311–320.
- Dietze, E., Wünnemann, B., Diekmann, B., Aichner, B., Hartmann, K., Herzschuh, U., IJmker, J., Jin, H., Kopsch, C., Lehmkuhl, F., Li, S., Mischke, S., Niessen, F., Opitz, S., Stauch, G., Yang, S., 2010. Basin morphology and seismic stratigraphy of Lake Donggi Cona, north-eastern Tibetan Plateau, China. *Quaternary International* 218, 131–142.
- Dietze, E., Hartmann, K., Diekmann, B., IJmker, J., Lehmkuhl, F., Opitz, S., Stauch, G., Wünnemann, B., Borchers, A., 2012. An end-member algorithm for deciphering modern detrital processes from lake sediments of Lake Donggi Cona, NE Tibetan Plateau, China. *Sedimentary Geology* 243–244, 169–180.

502 Domrös, M., Peng, G., 1988. The climate of China. Springer Verlag, Berlin, Heidelberg, New York.

503 Dykoski, C. A., Edwards, R. L., Cheng, H., Yuan, D., Cai, Y., Zhang, M., Lin, Y., Qing, J., An, Z.,
504 Revenaugh, J., 2005. A high-resolution, absolute-dated Holocene and deglacial Asian monsoon record
505 from Dongge Cave, China. *Earth and Planetary Science Letters* 233, 71–86.

506 Fleitmann, D., Burns, S. J., Mudelsee, M., Neff, U., Kramers, J., Mangini, A., Matter, A., 2003.
507 Holocene Forcing of the Indian Monsoon Recorded in a Stalagmite from Southern Oman. *Science*
508 300, 1737–1739.

509 Folk, R.L., Ward, W.C., 1957. Brazos river bar: a study in the significance of grain size parameters.
510 *Journal of Sedimentary Petrology* 27, 3–26.

511 Gasse, F., Arnold, M., Fontes, J. C., Fort, M., Gibert, E., Huc, A., Bingyan, L., Yuanfang, L., Qing, L.,
512 Melieres, F., Campo, E. V., Fubao, W., Qingsong, Z., 1991. A 13,000-year climate record from western
513 Tibet. *Nature* 353, 742–745.

514 Herzsuh, U., Winter, K., Wünnemann, B., Li, S., 2006. A general cooling trend on the central Tibetan
515 Plateau throughout the Holocene recorded by the Lake Zige tang pollen spectra. *Quaternary*
516 *International* 154-155, 113–121.

517 IJmker, J., Stauch, G., Pötsch, S., Diekmann, B., Wünnemann, B., Lehmkuhl, F., 2012a. Dry periods on
518 the NE Tibetan Plateau during the late Quaternary. *Palaeogeography, Palaeoclimatology,*
519 *Palaeoecology* 346-347, 108–119.

520 IJmker, J., Stauch, G., Hartmann, K., Diekmann, B., Dietze, E., Opitz, S., Wünnemann, B., Lehmkuhl, F.,
521 2012b. Environmental conditions in the Donggi Cona lake catchment, NE Tibetan Plateau, based on
522 factor analysis of geochemical data. *Journal of Asian Earth Sciences* 44, 176–188.

523 Jin, Z., Yu, J., Chen, H., Wu, Y., Wang, S., Chen, S., 2007. The influence and chronological uncertainties
524 of the 8.2 ka cooling event on continental climate records in China. *The Holocene* 17, 1041–1050.

525 Lee, J., Li, S.-H., Aitchison, J. C., 2009. OSL dating of paleoshorelines at Lagkor Tso, western Tibet.
526 *Quaternary Geochronology* 4, 335–343.

527 Lehmkuhl, F., 1997. The spatial distribution of loess and loess-like sediments in the mountain areas of
528 Central and High Asia. *Zeitschrift f. Geomorphologie, N.F., Suppl.-Bd.* 111: 97–116.

529 Lehmkuhl, F., Haselein, F., 2000. Quaternary paleoenvironmental change on the Tibetan Plateau and
530 adjacent areas (Western china and Western Mongolia). *Quaternary International* 65-66, 121–145.

531 Li, D., Li, Y., Ma, B., Dong, G., Wang, L., Zhao, J., 2009. Lake-level fluctuations since the Last Glaciation
532 in Selin Co (lake), Central Tibet, investigated using optically stimulated luminescence dating of beach
533 ridges. *Environmental Research Letters* 4, doi:10.1088/1748-9326/4/4/045204.

534 Locket, G., 2010. Geomorphologisch-fernerkundliche Untersuchungen zur litoralen Entwicklung des
535 Donggi Cona, Tibet Plateau (China). Diplom thesis, Institute of Geographical Sciences, Free University
536 Berlin.

537 Long, H., Lai, Z., Wang, N., Li, Y., 2010. Holocene climate variations from Zhuyeze terminal lake
538 records in East Asian monsoon margin in arid northern China. *Quaternary Research* 74, 46–56.

539 Madsen, D. B., Haizhou, M., Rhode, D., Brantingham, P. J., Forman, S. L., 2008. Age constraints on the
540 late Quaternary evolution of Qinghai Lake, Tibetan Plateau. *Quaternary Research* 69, 316–325.

541 Miehe, G., Kaiser, K., Sonam Co, Zhao, X.Q., Liu, J.Q., 2008. Geo-ecological transect Studies in
542 Northeast Tibet (Qinghai, China) reveal human-made mid-Holocene environmental changes in the
543 upper Yellow River catchment changing forest to grassland. *Erdkunde* 62, 187–199.

544 Mischke, S., Zhang, C., 2010. Holocene cold events on the Tibetan Plateau. *Global and Planetary*
545 *Change* 72, 155–163.

546 Mischke, S., Kramer, M., Zhang, C., Shang, H., Herzsuh, U., Erzinger, J., 2008. Reduced early
547 Holocene moisture availability in the Bayan Har Mountains, northeastern Tibetan Plateau, inferred
548 from a multi-proxy lake record. *Palaeogeography, Palaeoclimatology, Palaeoecology* 267, 59–76.

549 Mischke, S., Zhang, C., Börner, A., Herzsuh, U., 2009. Lateglacial and Holocene variation in aeolian
550 sediment flux over the northeastern Tibetan Plateau recorded by laminated sediments of a saline
551 meromictic lake. *Journal of Quaternary Science* 25, 162–177.

552 Mischke, S., Aichner, B., Diekmann, B., Herzsuh, U., Plessen, B., Wünnemann, B., Zhang, C., 2010.
553 Ostracods and stable isotopes of a late glacial and Holocene lake record from the NE Tibetan Plateau.
554 *Chemical Geology* 276, 95–103.

555 Morrill, C., Overpeck, J. T., Cole, J. E., 2003. A synthesis of abrupt changes in the Asian summer
556 monsoon since the last deglaciation. *The Holocene* 13, 465–476.

557 North, C. P., Davidson, S. K., 2012. Unconfined alluvial flow processes: Recognition and interpretation
558 of their deposits, and the significance for palaeogeographic reconstruction. *Earth-Science Reviews*
559 111, 199–223.

560 Opitz, S., Wünnemann, B., Aichner, B., Dietze, E., Hartmann, K., Herzsuh, U., Ijmker, J., Lehmkuhl,
561 F., Li, S., Mischke, S., Plotzki, A., Stauch, G., Diekmann, B., 2012. Late Glacial and Holocene
562 development of Lake Donggi Cona, north-eastern Tibetan Plateau, inferred from sedimentological
563 analysis. *Palaeogeography, Palaeoclimatology, Palaeoecology* 337–338, 159–176.

564 Qiu, J., 2008. The third pole. *Nature* 454, 393–396.

565 Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Blackwell, P.G., Bronk Ramsey, C., Buck,
566 C.E., Burr, G.S., Edwards, R.L., Friedrich, M., Guilderson, T.P., Hajdas, I., Heaton, T.J., Hogg, A.G.,
567 Hughen, K.A., Kaiser, K.F., Kromer, B., McCormac, G., Manning, S.W., Reimer, R.W., Richards, D.A.,
568 Southon, J.R., Talamo, S., Turney, C.S.M., van der Plicht, J., Weyhenmeyer, C.E., 2009. IntCal09 and
569 Marine09 radiocarbon age calibration curves, 0–50,000 years cal BP. *Radiocarbon* 51, 1111–1150.

570 Rohling, E. J., Palikey, H., 2005. Centennial-scale climate cooling with a sudden cold event around
571 8,200 years ago. *Nature* 434, 975–979.

572 Schlütz, F., Lehmkuhl, F., 2009. Holocene climatic change and the nomadic Anthropocene in Eastern
573 Tibet: palynological and geomorphological results from the Nianbaoyeze Mountains. *Quaternary*
574 *Science Reviews* 28, 1449–1471.

575 Stauch, G., IJmker, J., Pötsch, S., Zhao, H., Hilgers, A., Diekmann, B., Dietze, E., Hartmann, K., Opitz, S.,
 576 Wünnemann, B., Lehmkuhl, F., 2012. Aeolian sediments on the north-eastern Tibetan Plateau.
 577 Quaternary Science Reviews, 57, 71–84.

578 Stuiver, M., Reimer, P.J., 1993, Radiocarbon, 35, 215–230.

579 Sun, D., Bloemendal, J., Rea, D. K., Vandenberghe, J., Jiang, F., An, Z., Su, R., 2002. Grain-size
 580 distribution function of polymodal sediments in hydraulic and aeolian environments, and numerical
 581 partitioning of the sedimentary components. Sedimentary Geology 152, 263–277.

582 Sun, J., Li, S.-H., Muhs, D. R., Li, B., 2007. Loess sedimentation in Tibet: provenance, processes, and
 583 link with Quaternary glaciations. Quaternary Science Reviews 26, 2265–2280.

584 Taft, L., Wiechert, U., Riedel, F., Weynell, M., Zhang, H., 2012. Sub-seasonal oxygen and carbon
 585 isotope variations in shells of modern *Radix* sp. (Gastropoda) from the Tibetan Plateau: potential of a
 586 new archive for palaeoclimatic studies. Quaternary Science Reviews 34, 44–56.

587 Van der Woerd, J., Tapponnier, P., Ryerson, F. J., Meriaux, A.-S., Meyer, B., Gaudemer, Y., Finkel, R.
 588 C., Caffee, M. W., Guoguang, Z., Zhiqin, X., 2002. Uniform postglacial slip-rate along the central 600
 589 km of the Kunlun Fault (Tibet), from ²⁶Al, ¹⁰Be, and ¹⁴C dating of riser offsets, and climatic origin of
 590 the regional morphology. Geophysical Journal International 148, 356–388.

591 Van Vliet-Lanoë, B., 1998. Frost and soils: implications for paleosols, paleoclimates and stratigraphy.
 592 Catena 34, 157–183.

593 Wang, Y., Liu, X., Herzschuh, U., 2010. Asynchronous evolution of the Indian and East Asian Summer
 594 Monsoon indicated by Holocene moisture patterns in monsoonal central Asia. Earth-Science Reviews
 595 103, 135–153.

596 Wanner, H., Solomina, O., Grosjean, M., Ritz, S. P., Jetel, M. T., 2011. Structure and origin of
 597 Holocene cold events. Quaternary Science Reviews 30, 3109–3123.

598 Wünnemann, B., Demske, D., Tarasov, P., Kotlia, B. S., Reinhardt, C., Bloemendal, J., Diekmann, B.,
 599 Hartmann, K., Krois, J., Riedel, F., Arya, N., 2010. Hydrological evolution during the last 15 kyr in the
 600 Tso Kar lake basin (Ladakh, India), derived from geomorphological, sedimentological and
 601 palynological records. Quaternary Science Reviews 29, 1138–1155.

602

603 **Figures**

604 Figure 1: Donggi Cona lake basin and its catchment on the north-eastern TP [inset]. Main Kunlun fault
 605 after Van der Woerd et al. (2002) and assumed in-lake continuation after Dietze et al. (2010).
 606 Triangles indicate the locations of high-stand sediment sections (cf. Google Earth kmz-file).

607 Figure 2. Sites of onshore lake high-stand sediment sections at the outlet (A, view to north), at the
 608 northern shore (B, E, F) and on the eastern alluvial plain (C, D). B shows the stratigraphy of section
 609 P02, exemplarily. For locations see Figure 1.

610 Figure 3: Corrected and calibrated ages of high-stand sediments including error bars. Sections are
 611 separated by bold lines. Dated sample numbers refer to Table 2 (left axis) and are sorted by depth in
 612 each section. Remarks are added, when distinct sediment properties may bias the reliability of the

age. The grey blocks in the background show the associated onshore terrace (right axis, Lockot, 2010). Red and blue crosses mark the median and mean ages. *OSL-age ($\pm 2\sigma$) at base of loess section P063 for comparison (Stauch et al., 2012)

Figure 4: Generalized stratigraphic descriptions of the Donggi Cona high-stand sediment sections, including absolute position above sea level, Facies, interpreted lake level implications, and mean corrected, calibrated radiocarbon ages (BP, cf. Table 2, Figure 3). They are sorted by location and elevation of the section tops. Unspecified colors represent strata colors.

Figure 5: Examples for strata of different sediment facies and their association with a terrestrial, littoral or lake depositional environment. The lower row shows strata that are overprinted by pedogenesis, cryoturbation, and ongoing active layer mixing (left to right).

Figure 6: End-member loadings of all high-stand sediment grain size distributions (A). Mean total r^2 of data modelled with five end-members versus original data in variable (B) and sample space (C), after Dietze et al. (2012).

Figure 7: End-member scores, normalized difference of EM 2 and EM 5 ($EM_{diff} = (EM\ 2 - EM\ 5)/(EM\ 2 + EM\ 5)$), and their relation to field stratigraphy. Grey bars refer to sediment Facies and relative lake level assumption (see Figure 4 for legend). P15 and P06 are sections at the outlet (A), and P02, P21 and P14 are sections at the northern shore (B).

Figure 8: Higher than present lake level phases at Lake Donggi Cona compared with the lake-core proxies, ages of loess on top of onshore terraces, and Indian monsoon-dominated $\delta^{18}O$ of Oman speleothems in the background (Fleitmann et al., 2003). * Opitz et al. (2012); ** Mischke et al. (2010); *** OSL ages of loess sections on lake terraces, Stauch et al. (2012)

Figure S1: Higher than present lake level phases at Lake Donggi Cona compared with the lake-core proxies, and ages of loess on top of onshore terraces (for separate use and future comparison). * Opitz et al. (2012); ** Mischke et al. (2010); *** OSL ages of loess sections on lake terraces, Stauch et al. (2012)